Water's Journey from Rain to Stream

by

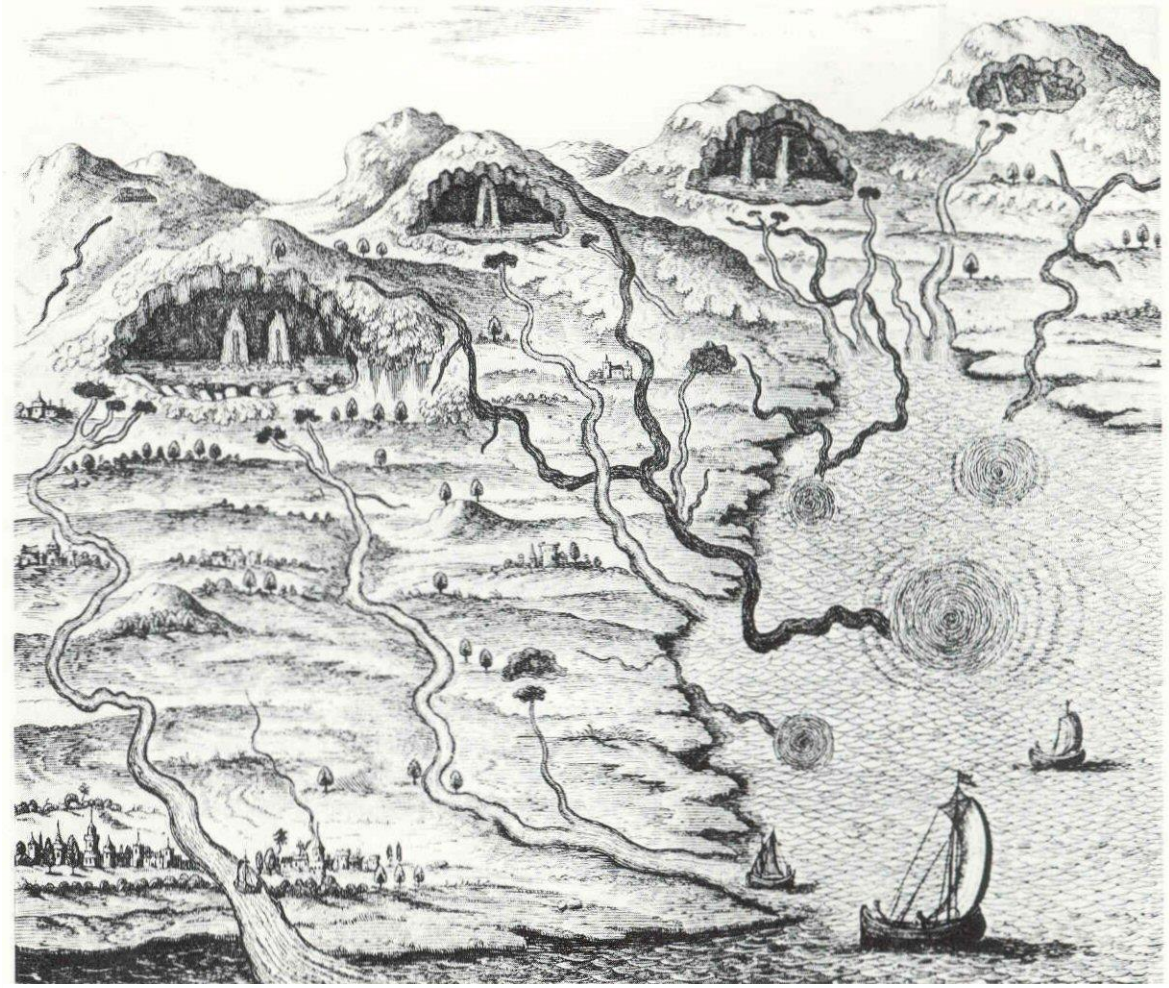
Harald Grip and Allan Rodhe

Preliminary translation into English by Harald Grip and Kevin Bishop
(edited January 2015)

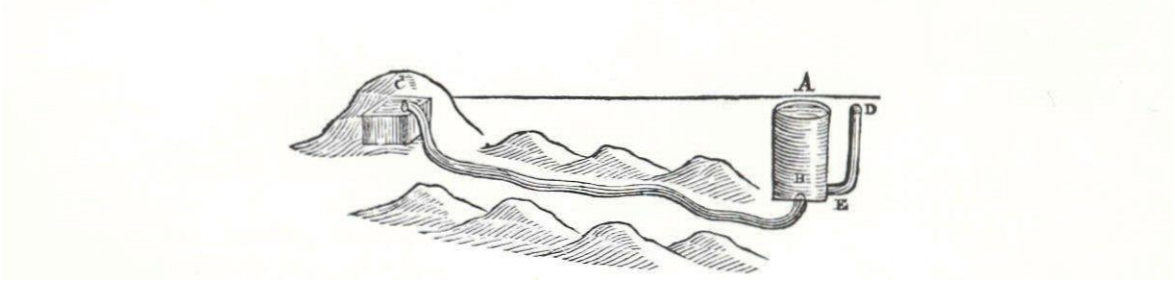
The Swedish edition:

Grip, H. and Rodhe, A., 2000, *Vattnets väg från regn till bäck*, Hallgren & Fallgren, Uppsala
(earlier edition Forskningsrådens förlagstjänst 1985)

Contents



In his great work Mundus Subterraneus (The underground world) from 1664 the German Jesuit munch and scientist Athanasius Kircher suggests that the rivers get their water from lakes in the interior of the mountains. These cave-lakes are replenished from the sea via underground rivers (dark in the figure). By laboratory experiments Kircher meant that he could show how the water could rise to the mountains when the pressure in the sea increased, e.g. by the influence of winds and tides.



1. The origin of streamwater – a controversy

In 1674 the French philosopher Pierre Perrault published the book "On the origin of springs". The book was a contribution to a debate on a question that had been discussed since the Antiquity: What is the origin of the water in springs and streams? Today the answer seems obvious. Anyone knows that it is rain or melting snow that delivers water to streams. But it is far from an obvious understanding that the rain, giving a few mm or cm of water if it falls into a bucket, is enough to feed the huge volume of water in rivers. Many suggestions had been discussed, e.g., that the water came from the ocean transmitted by subsurface rivers or that the air was transformed into water in the cold interior of the mountains. Based on measurements, Perrault showed that the rain and snowfall was quite enough to feed the Seine River with its water. According to his calculations the yearly flow of water in this river constituted only one sixth of the amount of precipitation that fell within what we today call its *catchment*, i.e., the area that according to the topography may collect water to the river. (Modern calculations give runoff as about one third of the precipitation.)

Which are the pathways for water from rain to stream?

Perrault laid the foundation for our understanding of the water balance of a catchment, i.e., that the precipitation over an area is temporarily stored, evaporated or discharged. But it is a big step from grasping the water balance to understanding how water passes through an area to a stream. Even today, 300 years after Perrault, there are many question marks concerning the processes that transform precipitation over an area into water flow in a stream, the so called *runoff generation*. Many basic questions still remain to be answered and put into context with climate, topography, geology, vegetation etc. Which pathways will water particles take through the area? How large are the fluxes and residence times on the soil surface and in different soil layers? In what way do different parts of the catchment contribute to the flux in the stream? How is the water pressure transmitted from the infiltration to the discharge into or close to the stream?

Knowledge about the process of runoff generation is necessary for predicting groundwater supply, water flow rates in streams and rivers as well as for predicting how these are affected by human activities. Such knowledge is also necessary for understanding the chemical changes that water undergoes during its flow through an area, a burning question in connection to e.g. acid precipitation or eutrophication.

Overland flow and base flow – the traditional view

The traditional understanding of runoff generation that has dominated textbooks in hydrology and practice during the 20th century is that flow events in streams are generated by water flow on the soil surface, i.e. surface runoff or *overland flow*. According to this theory overland flow is generated over the whole catchment. It is generated when more rain or meltwater is delivered to the soil surface than can be infiltrated into the soil. The water on the soil surface is first collected in depressions forming puddles. When they are filled, the water will start to run off in ephemeral rills, which merge, grow and eventually reach the perennial stream system. The streams also gain water from groundwater, a so called base flow, to which the overland flow is added. This base flow is considered to be only marginally influenced by the storm that causes the flow events. Between flow events the streamwater consists of groundwater, but during the flow events this

groundwater flow is only a small part of the total flow. According to this view, the entry of water into the soil, i.e. the *infiltration*, means that water is withdrawn from stream runoff. The *infiltration capacity*, i.e. the soil's ability to absorb water, thus determines whether a storm of a given intensity will lead to runoff in the streams or not.

With this view the water in the streams mainly consists of fresh rain or snowmelt water, i.e. water that has been in the catchment for only a few hours or days. The water has not been exposed t chemical processes in the soil and a large part of the pollutants that are added with the precipitation will reach the stream directly.

An early mathematical model for flood forecasting, built on this theory, is the so-called unit hydrograph. By comparing the water flow in a stream with the precipitation intensity in the catchment, an area-mean infiltration capacity is calculated. This infiltration capacity is used together with measured lag between precipitation and runoff to calculate the runoff in the stream for different possible precipitation intensities.

Robert E. Horton formulated this view on runoff formation during the early decades of the 20th century. He was mainly working in arable land and in dry areas in southern USA. In such areas overland flow from a large part of the catchment may dominate the streamflow in connection with large storms. Crusts may form on the soil surface during dry periods leading to low infiltration capacity.

Is overland flow really formed on Swedish till soils?

But how does it actually look in our climate and in the landscape typical for Sweden? How many have seen water flowing on the soil surface over extended areas in forested areas? Even if the discharge in streams has increased substantially after a heavy or long-lasting storm, it may only be on footpaths or in more or less permanently wet areas that water is splashing around your feet when you are walking in the rain. On footpaths we have caused compaction of the soil and thereby reduced the infiltration capacity. In the wet areas it is mostly the groundwater surface that has risen to the soil surface, so groundwater discharge prevents rainwater infiltration. It is seldom that the infiltration capacity of the soil actually limits infiltration.

Of course there are situations where the infiltration capacity is limiting. One such example is on rock-faces, but that overland flow will infiltrate after a short distance into cracks or into the soil below the rock-face. Another example is on some arable land in connection with heavy rain or snowmelt. Maybe occasionally during spring in some forest soils, when reoccurring melt- and freeze periods have formed an ice sheet in the upper soil layers. Still another example is on asphalt paving or other areas altered by man. But it is hard to believe that rainwater from the whole catchment can reach the watercourse through overland flow.

During the recent decades, extensive research has been conducted on runoff formation. It has primarily been done in Europe and North America in a climate that is similar to that in Sweden, but also in New Zeeland and in Australia. Many different approaches have been used to reveal the process. Examples are measurements of fluxes on and in the soil profile with troughs, analysis of the groundwater levels over extended areas, mathematical simulation of groundwater fluxes along hillslopes, studies of relationships between land surface topography and soil moisture and investigations of the origin of streamflow by means of water chemistry and isotopes.

Maybe overland flow is only formed on wet soil

Investigations in Eastern USA during the early 1960's found that precipitation intensity could best be related to streamflow if it was assumed that only a small part of the watershed contributed to the discharge, but on the other hand that all precipitation falling on that part contributed to runoff. This so-called active area was assumed to consist of moist areas close to the stream. The idea developed into a view that the flow events in the streams originated from overland flow from areas with the groundwater surface at or above the soil surface, i.e. *saturated areas*. The extent of these saturated areas was assumed to vary with time as the groundwater level varied in the area. With this view a certain storm would cause a larger streamflow when groundwater levels were high, i.e. during generally moist conditions. The role of groundwater in runoff formation would then be passive. The groundwater was looked upon as a regulating mechanism deciding the proportion of precipitation that would generate stream runoff, but groundwater would not actively contribute to the flow increase. Rainwater still dominates the streamflow.

Other investigations have shown that flow events in watercourses can be explained to a large extent by increased groundwater flow to the stream or to the saturated areas close to the stream. The presence of such large subsurface stormflows has been demonstrated, directly by troughs or indirectly by water chemistry or isotopes. According to this view a significantly larger area than that saturated to the ground surface is contributing to the flow events. Water is infiltrating over a large part of the catchment and presses groundwater into the streams, whose discharge is often dominated by groundwater.

Obviously two theories for runoff generation in our climate have developed. Both deviate substantially from the traditional view of infiltration-excess, but they are also quite different from each other. The difference can partly be explained by differences in geology, topography and climate in the areas where field investigations have been performed. As investigations are conducted on only few sites, while applications are made all over the world, it is important that the suggested models are critically reviewed and the conditions are carefully scrutinised. The question about the subsurface contribution to the flow events has thus been the subject of an intensive debate in the international literature.

The view that has developed in Sweden

In Sweden a new view has developed since the late 1960's which also stresses the active role of groundwater for runoff generation. In our till landscape the groundwater surface largely follows the soil surface. The groundwater is normally shallow, maybe a few metres below the ground surface on hilltops. In the lower parts of the hillslopes and in certain flat areas the groundwater is found only a few decimetres below the soil surface. In mires and streams the water surface coincides with the groundwater surface, which consequently lies above the soil surface.

By considering whether water is flowing into or out from the groundwater zone, the terrain can be divided into *recharge-* or *discharge areas*. High areas are mainly recharge areas, while low lying areas are discharge areas. Mires and streams are parts of the latter.

In Sweden the infiltration capacity is generally larger than rain or snowmelt intensity. Water that is delivered to recharge areas will therefore infiltrate. If the distance between the soil surface and the groundwater surface is small, the groundwater level will rapidly rise in response to infiltration. Due to a number of factors, the discharge from the discharge areas increases when the groundwater stage rises. The slope of the groundwater surface increases and the thickness of the groundwater zone increases. In addition the hydraulic conductivity of the soil is often much larger close to the soil surface than in deeper layers, resulting in a large increase in groundwater outflow even when the groundwater rise is small. The discharge in the stream will therefore increase quickly in response to increased water input.

Rain and meltwater on discharge areas can, on the other hand, not infiltrate as the pores of the soil are already filled. That water will therefore run off directly to the stream as overland flow together with the discharging groundwater.

With this view, flow events in streams consist of both rainwater that has fallen on discharge areas and groundwater discharged in these areas due to infiltration in the recharge areas. The flow events are often dominated by groundwater.

What has been sketched here is a hypothesis for runoff generation in Swedish till terrain. Still much research remains to quantify all the mechanisms involved and to connect them with the characteristics of each area and climate. But the hypothesis has shown to be a powerful tool for understanding and predicting water occurrence, flow and chemical composition in different parts of a catchment and in the streams. In this book we will treat water flow through a catchment based on this hypothesis. Our idea is to focus on the main principles for the water flow. The picture is therefore quite general, based on a generalized till landscape. Applications to specific catchments must always consider local geological circumstances.

2. Water turnover at a catchment scale

The turnover of water within a land area is characterised by the continuous flow of water from hills to valleys, from small to large watercourses. For some time water may be stored as snow or in soil, bedrock or watercourses. These storages are filled and emptied depending on seasonal variations. The source for the flow is precipitation and the output is by evaporation or runoff.

In this chapter we will show the role of different water storages in the water flow path from precipitation to its appearance in streams. The mechanisms governing storage and flow will be dealt with in more detail in later chapters, where many of the terms we introduce here will be defined in more detail.

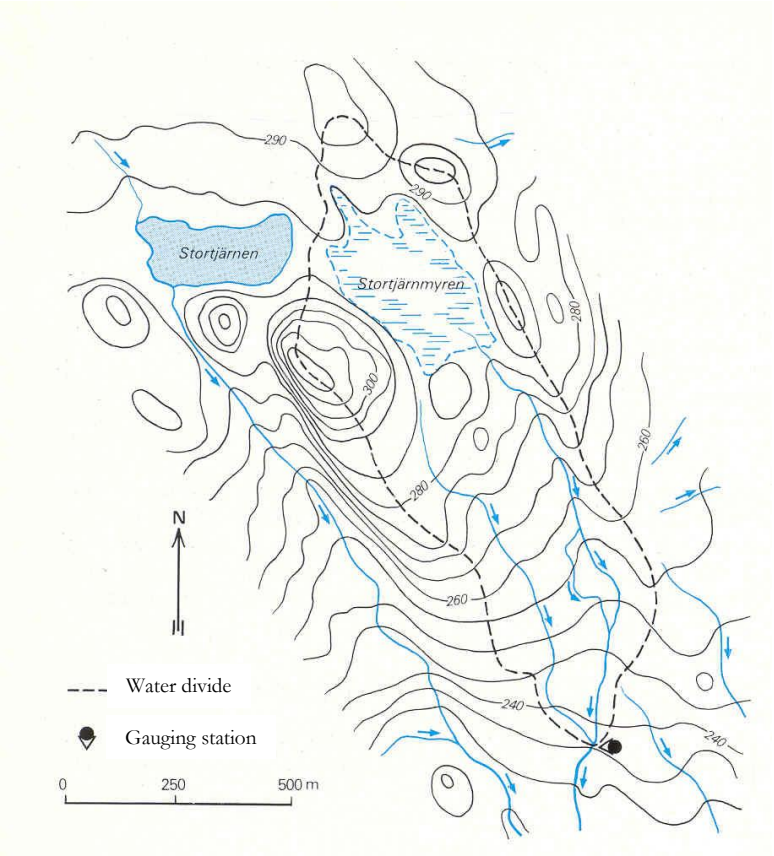


Fig. 1 Svartberget catchment, Västerbotten, Northern Sweden

The catchment is delimited by a water divide

Upstream from any point in a watercourse a *catchment* may be defined as the total area within which precipitation may contribute to the flow at that point. The catchment is delineated by a *water divide*. Precipitation that falls outside the water divide may contribute to the flow in another stream or to the flow in the same stream, but further downstream. A catchment defined at a point far downstream in a watercourse will include all possible catchments defined upstream from that point. There are no points in a landscape that do not belong to a catchment (Fig. 1).

The water divide constitutes the boundary between catchments. Water is always moving along slopes. Water divides plotted on a topographical map will therefore cross elevation lines at a right angle. One distinguishes between surface water divides and groundwater divides. A surface water divide appears in the terrain as a more or less distinct ridge, on each side of which water falling

on the surface would run in different directions. In many cases the terrain is rather flat and it is difficult to exactly determine the position of the water divide. The groundwater divide is similar, but follows a line along the ridge of the groundwater table. The closer the groundwater table is to the soil surface the better is the agreement between the surface and the groundwater divides. The position of a groundwater divide may change with the groundwater level.

In the till landscape of Sweden, where the soil cover is shallow on hilltops and the bedrock has rather few cracks, the surface and the groundwater divides coincide fairly well and we do not normally have to consider the difference between them.

No water can disappear

Precipitation falling on a catchment may temporarily be stored within the catchment, evaporate or run off. This is illustrated by the water balance equation

$$P = E + R + \Delta S$$

P is precipitation
E is evaporation. *(In this text the word evaporation is normally used synonymously with evapotranspiration to denote the total flow of water vapour from land and water to the atmosphere. In some discussions, however, when transpiration is discussed specifically, evaporation is used for the vapour flux from only free water (and snow), thus excluding the flow from transpiration. We apologize for this and will edit the text to make a consistent use of these terms.)*
R is runoff
ΔS is change in storage, i.e. the change Δ of the storage S
The units in the equation are normally given as volume per time and area, e.g. mm per year (month or day).

The unit mm (in fact mm water) denotes a volume of water per unit area, 1 mm = 1 litre/m^2. The runoff in a watercourse is normally given as *discharge* expressing the volume flow per unit time (m^3/s or litre/s). Dividing discharge with the catchment area gives the *specific discharge* (litre/(s km^2 or mm per time unit, where 1 litre/(s km^2) = 31.5 mm/year). The term runoff often, but not always, denotes the specific discharge.

The precipitation may be in liquid form as rain, dew and fog drips, or solid as snow and hail. Evaporation can take place from wet surfaces (wet vegetation, pools, lakes or watercourses), from snow and from bare ground. It can also take place via the stomata of leaves and needles, which is then called transpiration. Runoff can take place as surface runoff or groundwater runoff.

The basis of the water balance equation is that no water can disappear. Water that enters the catchment (P) must be stored (ΔS) or leave ($E + R$). The change in storage in the area, which is positive when the storages are replenishing and negative when emptying will have a damping influence on the runoff. The water in a rainstorm with one or a few hour's duration is partly stored and gives rise to runoff with a number of days' duration.

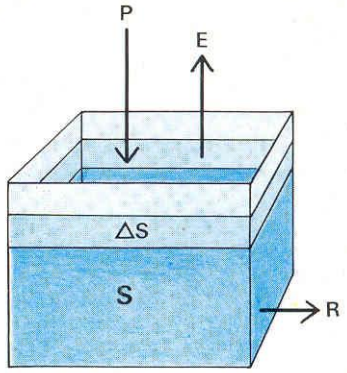


Fig. 2 The water balance equation. The precipitation (P) falling over a catchment during a certain time will be partitioned between evaporation (E), runoff (R) and change (ΔS) of the water storage (S). When looking at areas that are not complete catchments, such as small areas or lakes, inflow from surrounding land has to be added to the precipitation.

The water balance equation can be applied to areas of any size, from the River Göta älv catchment to a home garden. For small areas one usually looks at the water budget of a layer of the soil, for instance down to 1 m depth. The runoff is then often the downward flow from the layer, a flow that eventually will recharge the groundwater.

Different ways to describe the transit time of water

When talking about different reservoirs it is often of interest to know how long water stays in the reservoirs or how fast it turns over. In reservoirs with through-flow, e.g. a lake, the water particles have different *ages*. Some particles have just entered the reservoir and are young, while others have been there for a longer time. The water in the reservoir has therefore an age distribution. The time a particle has used to come from the inlet of the reservoir to its outlet is called the *transit time* (or travel time). The flow paths of the water particles through the reservoir are mostly of different length and the particles mostly move with different mean velocities. Therefore the travel of the different water particles through the reservoir will take different times and the water particles that leave the reservoir will have a transit time distribution. The *turnover time* for a reservoir is normally given as the ratio between the total volume and the total through-flow. The *residence time* is often used to denote the mean time a water particle stays in the reservoir.

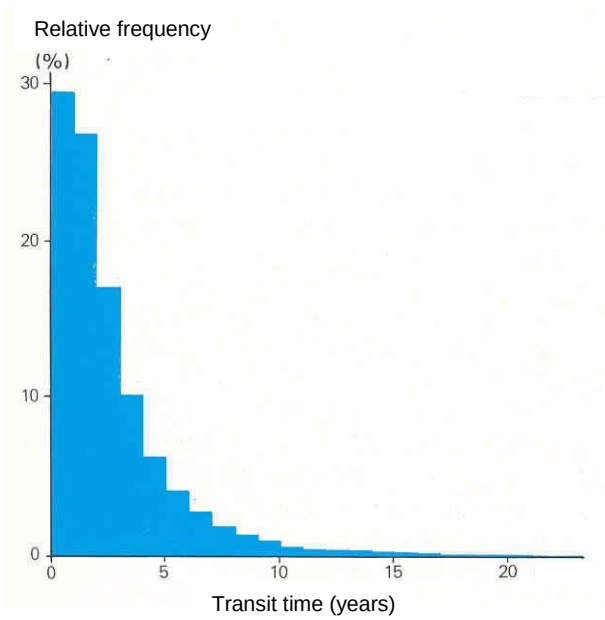


Fig. 3 Transit time distribution for water in River Råne älv catchment, Norrbotten. The nuclear bomb explosions in the atmosphere during the 1950's and 1960's added large amounts of radioactive tritium (^3H) to the atmosphere. After the test ban treaty in 1963 the tritium content of precipitation has gradually decreased. By comparing the tritium content of precipitation and river water over a long time period the residence times for the water particles in the catchment were calculated.

The diagram shows that 29% of the water particles that leave the catchment have been there one year or less, 27% from one to two years etc. A very small fraction of the water has been in the catchment more than 10 years.

Along the path of groundwater through soil and rock its content of dissolved matter will change through slow chemical reactions with its surroundings. The transit time for groundwater is therefore important for its quality at the point of outflow. The turnover time is important for the water quality in lakes. A small lake with a large drainage area may get its water replenished a couple of times per year, while a large lake with a small drainage area may have turnover times of decades.

Temporary storage in different reservoirs

There are a number of larger or smaller reservoirs where water is occasionally stored during its movement through a catchment. We will now follow the flow path for water from precipitation via the different reservoirs to the stream.

Interception

A part of the precipitation will never reach the soil surface, but will be caught on the leaves and branches of the trees. This is called *interception*. The storage capacity in the tree canopy is 0.5 – 2.5 mm water. The capacity differs depending on tree species and leaf area. Among our tree species the Norway spruce has the largest interception capacity. In forested areas 20 – 40% of a summer's precipitation will be transferred back to the atmosphere by evaporation from the interception storage. The reservoir is normally emptied (the trees are drying up) within a couple of hours after a storm event. Even though a part of the intercepted water is evaporated during the storm most of it is evaporated after the rain has stopped. The total amount of water evaporated from intercepted precipitation is therefore to a large degree dependent on the number of drying occasions, i.e. the number of storms. As shown in Fig. 4 intercepted water is flowing along the sloping spruce branches and drips at the periphery. That is why a dense spruce gives a good shelter if you are caught by a storm. Among our species beech, which has smooth bark and branches directed upwards, collects the intercepted water like a trough and a substantial flow will occur along its stem or drip from the branches.

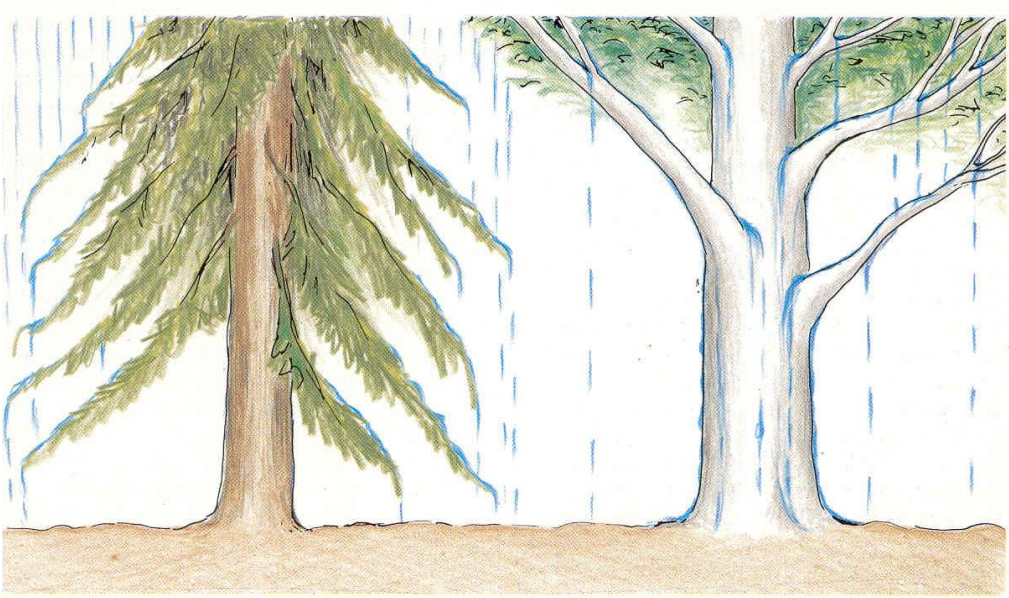


Fig. 4 Around 2 mm of a rainfall may be caught by the forest canopy, so called interception. When the interception storage is full all subsequent precipitation will reach the ground surface by dripping, stem flow, or rainfall between the canopies. The large turbulence that develops when the wind is blowing over the rough canopy surface provides an efficient transport of water vapour to the atmosphere. The

trees dry up after every rainfall even if the weather is not very suitable for evaporation. About one third of the summer precipitation in a central Swedish coniferous forest never reaches the ground surface due to interception.

Snow cover

In large parts of Sweden the snow cover has a decisive impact on the yearly turnover of water and gives the runoff its characteristic yearly variation. The density of snow normally increases with time. The density of newly fallen snow is about 100 kg m^3, while the density of old snow is about 400 kg/m^3. The snow cover represents a large water reservoir. A 0.7 m deep snow pack with a density of 300 kg m^3 contains 210 mm water. The snow cover is also a large cold store, which requires a great deal of energy to melt. During a warm day in spring 10 – 15 mm will melt. The melting mainly takes place at the snow surface.

Soil water

The water that is stored in the soil above the groundwater surface is called *soil water*. The upper part of the soil water reservoir, where the majority of the roots are located is called the *root zone*. In forest soil the roots are normally concentrated in the upper half meter. Ninety per cent of the root biomass may be located in the upper 10 – 20 cm of the soil. The amount of soil water that the trees can utilise in a till soil may be of the order of 20 – 30 mm water per 100 mm soil depth. The storage of water in the root zone is important for the water economy of the vegetation. In Sweden the early summer often has rather low precipitation and the plants have to rely on stored water in the root zone. The storage in the root zone is usually at its maximum after the snowmelt period in spring and in the autumn when water uptake by plants is small.

Let us estimate the turnover time for water in the root zone in a Swedish forest soil. Suppose the yearly precipitation is 800 mm and the evaporation of intercepted precipitation is 150 mm per year. The water supply to the soil surface, and therefore the infiltration is 650 mm per year if all water infiltrates. In a 0.5 m thick root zone with a mean water content of 30% the total water volume is 0.5 x 0.3 = 0.15 m or 150 mm. The turnover rate of this root zone, i.e. its volume divided by the flow, is 150/650 = 0.23 years, or 2.8 months.

Groundwater

The size of the groundwater reservoir in the soil is determined by the pore volume relative to the total soil volume (30 – 60 %) and the depth to the bedrock. The water content in the primary rocks may amount to a few parts of a per cent of the total rock volume. If the depth of the groundwater zone in the soil is one metre and the pores make up half of the volume the groundwater reservoir will contain 500 mm of water. With the same infiltration as in the soil water example above and a transpiration by the vegetation of 300 mm per year, the flow to the groundwater will be 350 mm per year. The turnover rate is then about 1.4 years. Different parts of the groundwater reservoir are renewed at different rates. Therefore different water particles can have very different ages. The mean age of groundwater in till soil is probably higher than the turnover rate, because deep groundwater moves substantially slower than shallow groundwater. A substantial part of the groundwater consists of water particles that have been in the groundwater zone for a long time. This gives the groundwater a high mean age. There is, however, shallow groundwater with a comparatively high velocity, where all water particles are young. These young water particles constitute only a small part of the total groundwater volume and have little effect on the mean age of the groundwater, but the large flow they represent makes on the other hand the turnover time short.

Surface water

The *surface water reservoir* is made up by water in lakes, pools and watercourses. In a catchment without lakes the surface water reservoir is small. In connection with snowmelt or large storms pools may be formed in the lower parts of the catchment. This water, the surface of which is normally a part of the groundwater surface, may represent a reservoir of up to 5 – 10 mm when the water supply is large. When the supply has decreased the groundwater level decreases and the most pools vanish. The water in the watercourses represents only about 1 mm over the entire catchment (one litre per m^2 of the catchment) with a variation from maybe 10 mm to close to zero. The turnover time may vary from only a few hours in a small stream to a couple of days in a large river. If the watercourse contains lakes the total reservoir will be substantially larger and thus also the turnover time.

Hydrological new-year in the autumn

To be able to calculate the evaporation from the water balance equation, measurements of precipitation and runoff must be available, as well as an estimate of the storage. From the overview of the different storage components given above it seems nearly impossible to accurately measure the change in storage between different times. If the time interval chosen is long enough (a number of years) the change in storage may be neglected. In order to minimize the error, water balance calculations are usually done for so-called 'hydrological years', where the new-year is chosen at a time when the storages are as equal as possible between years, giving a small change in storage from year to year. A good choice is when the storages are small. In our climate this is in early autumn.

Runoff variation

Largest amount of precipitation during summer – largest water supply during spring

Almost all over Sweden precipitation is largest in July and August and smallest in late winter and spring. Regarding the runoff generation we are primarily interested in the *water supply*, i.e. rain or snowmelt. Due to storage of snow during winter, the yearly variation in water supply is different from that of precipitation. In most of Northern Sweden more than 40% of the yearly precipitation falls as snow, in Central Sweden about 30% and in Southern Sweden about 20%. In the southern part this snow will melt now and then, but in the central part and especially in the northern part of the country, the snow will melt during a few intense weeks in spring. In much of the country we therefore have the largest water supply in spring.

Most water evaporates

The effect of water supply on runoff is largely dependent on evaporation. This follows more or less the temperature variation, with the highest rate during summer. During late autumn and winter the evaporation rate is very small. In most of the country more than half of the precipitation evaporates.

Regular yearly variation of runoff in the North and South

The runoff in our watercourses varies with season, but the variation is different in different parts of the country. In the North the yearly runoff is dominated by water from the snowmelt. The spring flood of the mountain streams occurs in June or July, while the spring flood of the forest rivers in the North occurs already in May. The mountain streams of the North have only one flood period per year, and after the spring flood the discharge decreases continuously until next year's spring. In addition to the spring floods, the forest rivers and the mountain streams of Central Sweden also have an autumn flood. This flood is normally much smaller than the spring flood and occurs between September and November. Like the mountain streams the forest rivers have their lowest runoff just before the start of the spring flood.

In Southern and Central Sweden the low flow usually occurs in July although this is often the month with most rain. The reason is that evaporation is largest during this month. The small rivers in the southernmost part of Sweden have maximum runoff during December and January and therefore have a yearly variation almost opposite to the mountain streams.

In the small rivers in Central Sweden both a spring and an autumn flood are common. The low evaporation in combination with the autumn precipitation causes an autumn flood that may be larger than the spring flood. In Fig. 5 we can compare the long term mean monthly runoff from a small river outside Uppsala with daily runoff from 1981 and 1982. The spring flood in 1981 was distributed over a number of flow events from January to May. Additional flow events were caused by a rainy summer. The dramatically high flow rate in August was due to heavy rains. A rainy autumn gave large flow from the end of October to the beginning of December. The mean runoff during 1981 was the second highest during the 24-year period of observations.

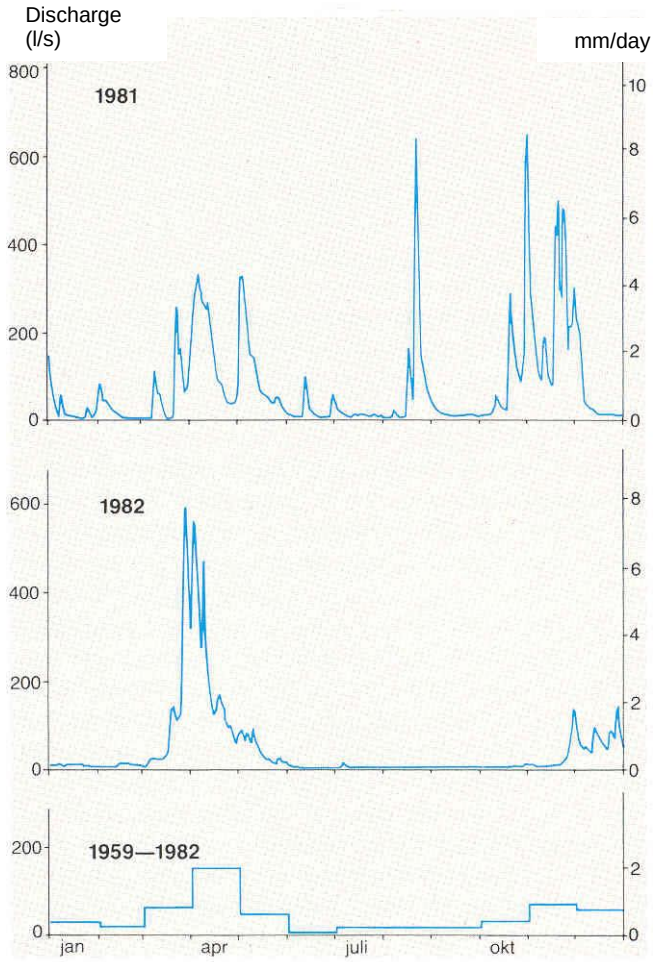


Fig. 5 Daily runoff during two years and long term mean values of monthly runoff in the river Uppsala-Näs-bäcken, Uppsala. The catchment is 6.8 km². In this part of Sweden the snow storage is irregular, resulting in many runoff events some winters. Other winters may have long periods with little flow ending with a strong spring flood. Normally the runoff is very small during summer, even if it rains, but very large rainfalls may give high flow also during this time. During autumn the evaporation is small and the rains often give considerable runoff.

The daily mean values, shown in the diagram, do not show the maximum flow, since the flow varies within the day. During spring flood in 1982 the highest momentary flow was 720 l/s.

The runoff in 1982 was concentrated to the snowmelt in March and April and to the autumn flood in November and December. During the summer the stream had nearly dried up. The mean runoff during 1982 was close to the mean for the 24-year period.

The large difference between the runoff distributions during the two years is typical for Central Sweden, which has unstable winters, large evaporation during summer and often heavy rains during the autumn.

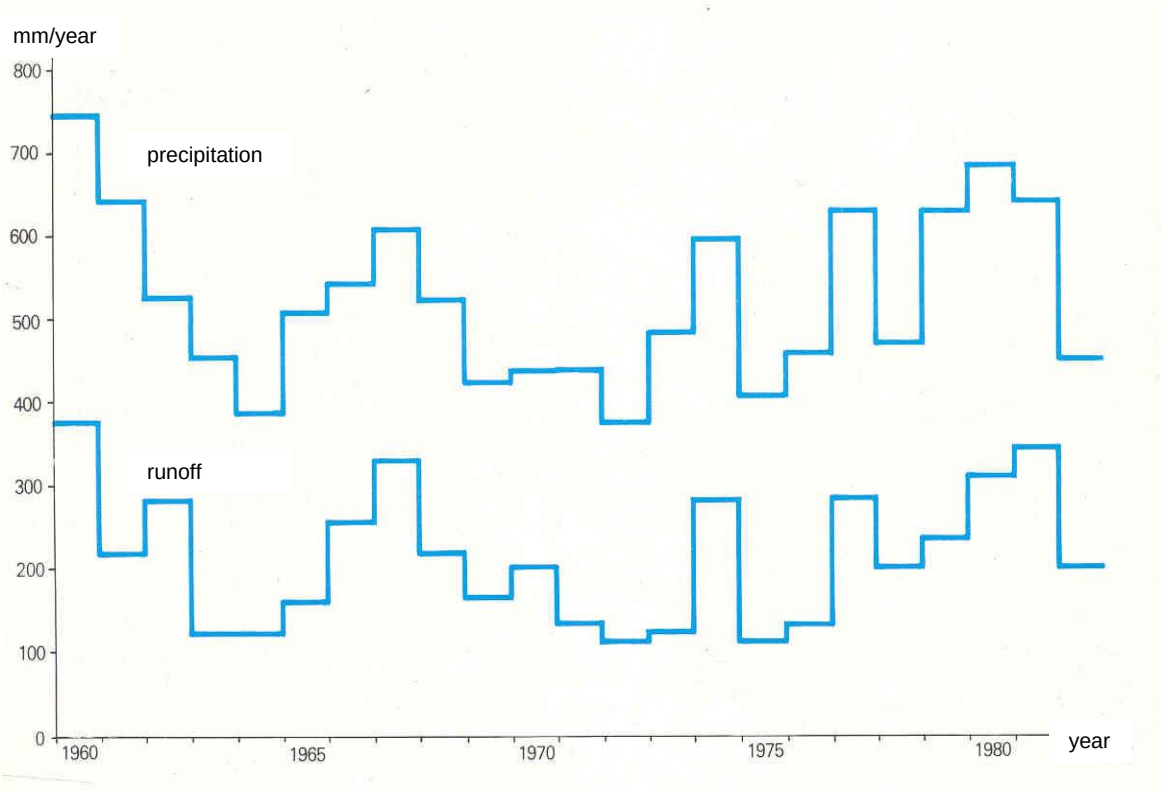


Fig.6 Yearly precipitation and runoff in River Uppsala-Näs-bäcken, Uppsala. The runoff follows the variations of the precipitation, giving large relative variations in the runoff and thus in water availability.

Large variations in yearly runoff

The precipitation varies a lot from year to year (Fig. 6). During the 23-year period shown the yearly precipitation varied between 370 and 740 mm in Uppsala, while the runoff varied between 110 and 380 mm. In the figure it is seen that the variation in yearly runoff is well connected to the variation in yearly precipitation. If the change in storage between years is neglected the evaporation is found as the difference between the two graphs in the diagram. (Actually the yearly evaporation is larger than this difference due to systematic errors in the precipitation measurements. It has been found that the precipitation gauges used by SMHI (Swedish Meteorological and Hydrological Institute) give too little precipitation. The raindrops and especially the snowflakes have a tendency to pass by the gauges. In addition some water evaporates from the gauges before they are read. Based on, among other things, the number of snowfall and rainfall events, one tries to correct the measured precipitation amounts. Without such corrections the estimated precipitation over Sweden is 15 – 25% too low. The larger the

proportion of precipitation that falls as snow, the larger this error is. The precipitation values presented in Fig. 6 are, as those given by SMHI on routine basis, not corrected.)

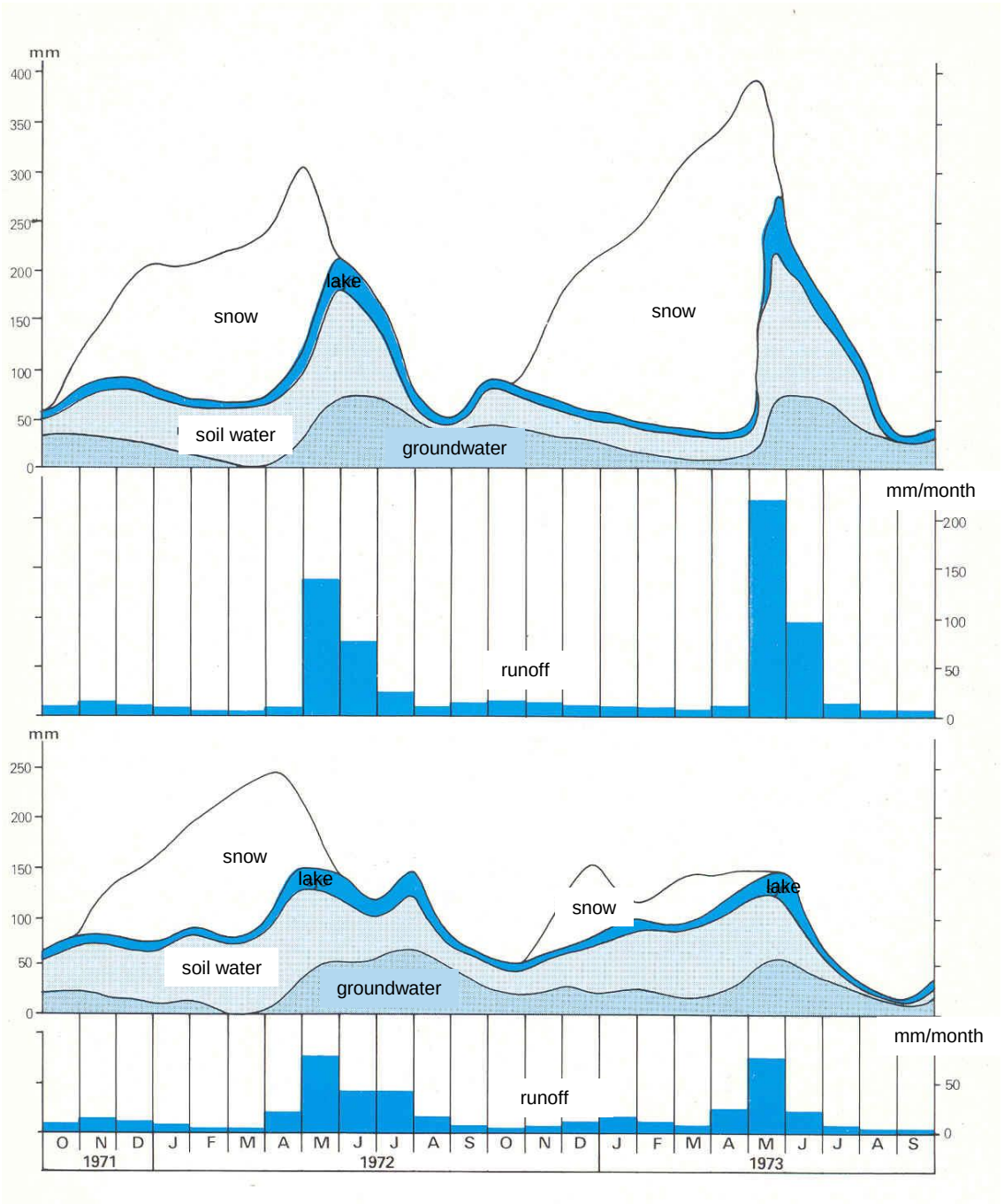


Fig. 7 The terms of the water balance equation were measured for a 10 year period in a number of so called representative catchments in order to get basic data on the flow and storage of water in small catchments. The figure shows storages and monthly runoff during two years in two catchments, Lappträsket (River Råne Älv, Norrbotten, Northern Sweden) and Kassjöån (Jämtland/Medelpad, Central Sweden). The storages are shown relative to arbitrary reference levels (the smallest values over the observation periods). From the graphs one can see the changes of the different storages and the change in total storage. The absolute magnitude of the storage can be seen only for the snow storage, which is empty in summertime. The soil water storage in Lappträsket, for example, is not zero in September 1973 but it had its lowest value during the period at that time.

The seasonal variation of runoff, with a dominant spring flood and a smaller autumn peak, is typical for the forest areas of Northern Sweden.

Since the variation in runoff is similar to the variation in precipitation, and the runoff is much smaller than the precipitation, the relative variation in yearly runoff is much larger than that of the precipitation. The runoff from an area is a measure of its water resources. For a sustainable use of the water resources, it is not the size of the groundwater and lake reservoirs that matters, but it is their filling rate that determines how much water one can extract. The large relative variation of the yearly runoff shows that the water resources vary a lot from year to year. We experience these variations during dry years with lowered groundwater levels and occasional problems with water supplies. The reason is the variation in precipitation. When it increases again the groundwater reservoirs recover.

Large changes in storage during a year

The different reservoirs have a characteristic yearly variation. This is most evident in areas with large snow accumulation and stable winters (see upper part of Fig. 7). The snow storage continuously grows during the whole winter, while the groundwater and lake storages are successively emptied. The soil water storage variation during the observed winter periods is irregular due to sudden small melt periods during the winters and soil water flow induced by soil frost growth. During spring, snowmelt decreases the snow storage rapidly while groundwater, soil water and lake storages increase and achieve their largest volumes. During the summer these storages again decline. When the evaporation is low during the autumn the rain fills the storages before the next emptying period in winter. In Southern Sweden, where melt periods and rain are common during winter, the storages may be kept at a high level due to frequent water additions.

3. Water's occurrence and flow in soil and bedrock

As a background to the continued discussion of water's journey through the watershed, we are going to describe the processes that control water's storage and flow in soil and bedrock. We will be relatively detailed, because we believe flows in nature are best understood in terms of the basic physical principles. We hope that the few mathematical expressions that appear do not intimidate the reader. Those equations express, in a compressed form, what we try to illustrate in text and figures.

Water – a fantastic liquid

Water has a number of special characteristics that influence its behaviour in nature. Despite its low molecular weight, water is a liquid at normal temperatures in nature, while similar compounds are gases. Compared with other liquids, water has an unusually high melting and boiling point, high latent heats of melting and vaporization, as well as a high surface tension. The cause of these extreme characteristics is water's molecular structure.

A water molecule consists of two hydrogen atoms and an oxygen atom, H_2O. The hydrogen atoms are bound to the oxygen atom so that the tree atoms form the points of an isosceles triangle, with the oxygen atom at the top. The angle at the point formed by the oxygen atom is 104.5°. As a result, one side of the atom (the oxygen side) gets a somewhat negative electrostatic charge, while the other side gets a somewhat positive charge (the hydrogen side). Because the charge is unequally distributed over the molecule, it is called a dipole.

Two water molecules attract each other so that the oxygen atom of one is turned towards the others' hydrogen atoms. It is this attraction between the molecules, the hydrogen-binding, which gives water the special characteristics mentioned above. This is also the reason why water is such a good solvent, and that it adsorbs (sticks) to solid surfaces.

In ordinary ice, the oxygen atom in every water molecule is surrounded in a regular pattern by the oxygen atoms of four neighbouring molecules. This structure, which takes up a great deal of space, is held together by hydrogen-binding. When ice melts, the molecules can pack closer together than in ice, with its strict geometric demands. Therefore water at 0°C has a higher density than ice at the same temperature. Thus ice floats on water. It has been calculated that only about 15% of the hydrogen-bonds break when ice melts to water at 0°C. Raising the temperature above 0°C breaks more hydrogen-bonds and the packing can be even tighter. Water has its greatest density, and therefore is heaviest, at +4°C. Thereafter the heat-movements of the molecules get the upper hand, and density decreases. Even at 100°C, though, there are strong attractions between water molecules. It takes a great deal of energy to eventually overpower these bonds and transform liquid water into water vapour.

The fact that the solid phase, ice, floats on the liquid phase, is perhaps the most remarkable feature of water. If water behaved like most other liquids, then in Scandinavia and other regions with subfreezing temperatures for sustained periods, lakes would freeze to the bottom and only thaw in a thin surface layer during the summer. In those regions, higher life would simply not be possible in water.

Water Content and Porosity

Soil is comprised of solid material, liquid and gas. Both mineral grains and organic matter make up the solid component. The liquid component includes water as well as dissolved substances. The gas component is a mixture somewhat different from that of the atmosphere. One can think of the solid phase as the soil's 'skeleton'. The construction of that skeleton determines the size and shape of the pore space, which in turn determines the behaviour of liquid and gas in the soil. Since water and air compete for the pore space (that is the space between the solid particles), one decreases when the other increases.

The term *porosity* refers to the proportion of the total soil volume that is comprised of pore space. For spherical mineral particles of uniform size, the porosity varies between 48 and 28% depending on how the particles are packed, but independent of the size of the particles. A bucket of apples and a bucket of peas which have both been shaken to achieve the densest packing possible will both contain about 30% air by volume. If one mixes large and small particles, the porosity will be smaller, since the interstices between the large particles will be partially filled by the smaller particles. In a soil, the particles are not spherical, which gives a larger porosity than if they were spherical. Typical values of porosity for a mineral soil are between 30 and 60%, while for peat the porosity is over 90%.

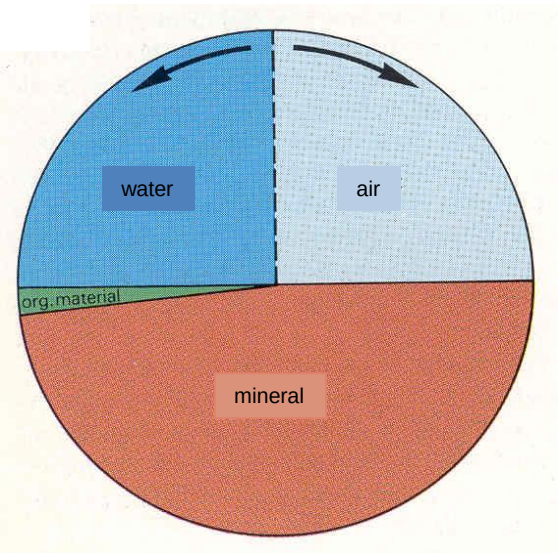


Fig. 8 About half of a well sorted soil consists of solid material. Water and air compete for the interstitial volume
Porosity = the interstices' portion of the total soil volume
Water Content = water's portion of the total soil volume

The size of the pores depends partly on the size and shape of the mineral particles, and partly on how they are arranged to build the soil "skeleton". The term *texture* usually refers to the size of the soil particles. For example, in a sandy soil, the particle size distribution and porosity are related to the texture. In a fine grained soil, though, the mineral particles have a tendency to stick together in larger units, aggregates. In between these aggregates are pores which are much larger than could have formed between individual mineral particles. Therefore the soil can feel coarse, even though it is very fine-grained. The formation of aggregates gives soil what is termed *structure*. The structure is particularly evident in the superficial soil layers due to biological activity and influence of soil frost. (An important reason for ploughing and other measures on agricultural soil is to give soil a structure which is conducive to the growth of crops.) Worm holes, the channels left by decayed roots, and drying cracks are examples of pores that contribute to the soil's structure.

The relative proportion of water in the total soil volume (solid material and pores) is called the *water content*, and is often expressed as a volumetric percentage. In other situations the expression "water content" can refer to the weight of water relative to the weight of the solid soil. This is expressed as a weight percent. The following expressions describe the relationship between volumetric percentage and weight percentage.

$\theta = w \cdot s / \varrho$

θ = water content in volumetric percent
w = water content in weight percent
s = dry bulk density, which is the ratio between a soil sample's mass and volume
ϱ = density of water

If all pores are filled with water, then the soil is *saturated*. The water content at saturation is identical to the porosity. The water storage, S, in a soil layer of a given thickness is often expressed in terms of mm of water, i.e. as litre/m^2. If θ is the average water content between layers z_1 and z_2, then $S = \theta \cdot (z_2 - z_1)$. (Here θ should be expressed as a fraction, not as a percentage.)

The *groundwater level* is the level of the water surface which is found in a hole dug in the soil, or in a perforated tube driven into the soil. This is the level where the water pressure is the same as that of the atmosphere. Beneath the groundwater level, in the *saturated zone*, or the *groundwater zone*, the water pressure is greater than the atmosphere's. All pores are completely filled with water. Above the groundwater level, in the *unsaturated zone*, (also called the *vadose zone* or *soil water zone*); the pressure is less than the atmosphere's. There is both water and air in these soil pores. (In the discussion of runoff generation, we refer to the saturated area (cf. page 5). This is the portion of the catchment where the soil is saturated all the way to the soil surface, i.e., the area where the groundwater level is at, or above, the soil surface.)

The root zone plays a key role in the transmission of water to the groundwater zone. The root zone is where plant uptake determines how much of the infiltrating precipitation returns to the atmosphere via transpiration, and how much water flows further down to the groundwater zone. This is also the zone where some of the most rapid changes occur in the water chemistry. The unsaturated zone between the root zone and the groundwater level is called the *intermediary* zone.

Water's binding to the soil

Soil has a tendency to retain water even when it is freely drained. If a small tank with a tap at the bottom is filled with gravel and water, almost all of the water will drain when the tap is opened. But if the tank is filled with fine sand, or a mixture of fine sand and gravel, only a few drops will drain when the tap is opened. The water retention quality of gravel differs markedly from that of sand or the sand-gravel mixture.

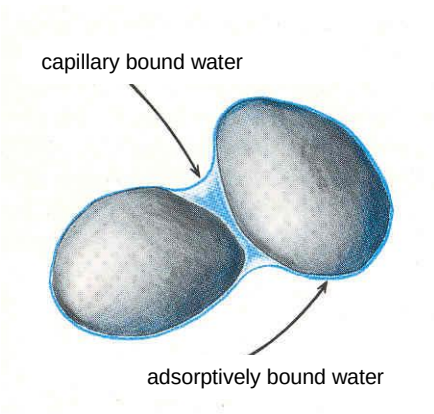


Fig. 9 Adsorption binds water to mineral particles in a very thin layer. In the pores between mineral particles, water can be held by an interaction of liquid, gas and mineral (or organic) particles in what is called "capillary binding" caused by water´s surface tension. Capillary binding can only occur if there is an interface between liquid and gas, and therefore not when the soil is saturated.

Water binds to the soil through both adsorption and surface tension. *Adsorption*, or the adhesion of water molecules to soil particles, results from the electrostatic forces between the dipolar water molecules and the charged surfaces of the soil particles.

A water molecule within a body of water is attracted in all directions to the neighbouring water molecules. For a molecule at the water surface, the forces of attraction to the molecules in the air are less than to the water molecules to the side and below the molecule. Therefore the water surface tends to hold together. This is the *surface tension* which gives a water drop its rounded form. The surface tension is influenced by dissolved substances. If the attraction between the molecules of the dissolved substance and water molecules is stronger than between the water molecules, then the solution's surface tension is greater than that of pure water. This is the case with a salt solution.

Water's *capillary* rise in a narrow glass tube results from adsorption and surface tension. These give rise to what is termed the capillary forces in the tube. We will now see how water is held in the pores of the soil by capillary forces.

The water surface bends upwards where it meets something like a glass wall. The adsorptive forces between the glass wall and the water are larger than the forces of attraction between the air and the glass wall. Therefore water wets the wall and the angle of contact between the wall and the water is a pointed tip. In a similar fashion, a bulging water surface is formed in a narrow glass tube that is pushed down into water. The bend in the water surface becomes larger the smaller the tube is. The bend shows that the water under the surface has a lower pressure than the air above. The pressure at the water surface outside the tube is that of the atmosphere. The water is therefore pushed up into the tube by the greater pressure on the free water outside the tube, until the original pressure difference is balanced by the pressure from the water column that has been pushed up. The capillary rise in a circular tube is described by the following relationship

$$h = \gamma \cdot \cos \alpha \,/\, (\varrho \cdot g \cdot r)$$

h = capillary rise
γ = the liquid's surface tension
α = the angle of contact between the meniscus and the wall (0 is complete wetting, i.e. the wall of the tube is tangent to the meniscus.)
ϱ = the liquid's density
g = acceleration due to gravity
r = the radius of the tube
For water and mineral soil (as for water and glass) the angle of contact is close to 0. This gives, for h and r in cm,

$$h = 0.15/r$$

This expression gives the relationship between the pore radius and the capillary binding force which the pore generates. The *binding force* (a negative pressure in the water) means that a compensating force is required to drain a pore of its water, since the water is held in the pore. The binding force is caused, as explained above, by the interaction between the molecules of the water, the wall and the air. One of the prerequisites for the existence of this binding force is an interface between water and air. If this interface does not exist, as is the case in the groundwater zone, then there are no capillary forces, and the water is not held by capillary forces in the pores. (The adsorptive forces, which bind the water to the mineral particle, do however, still exist.)

Water's binding described by the pF-curve

Let us return to the tank filled with gravel and sand. In both the sand and the gravel, the water bound by adsorption accounts for only a very small volume of water. The finer the grains of the soil are the greater is the significance of the adsorptively bound water, since the surface area of particles per unit volume increases. The layer around every particle bound by adsorption is said to be about seven molecules thick, and every water molecule is ca $3 \cdot 10^{-7}$ mm in diameter. If we assume that the particles are spherical, that gives a content of water bound by adsorption in sand (0.6 mm particle diameter) of 0.001%. The percentage is even smaller in the gravel. The tiny volume of water that, despite everything, remains in the gravel after it has been drained is in large the part that bound by capillary forces where the mineral particles are in contact with one another (cf. Fig. 9). After draining the sand and the sand and gravel mixture, practically all of their pore volumes are filled with water held by capillary forces.

The assumption of spherical particles can be a reasonable one in very coarse-grained soils. A corresponding calculation for clay (0.0002 mm diameter), on the other hand, gives too small a proportion of water bound by adsorption (3.3%). In fact, clay particles are not at all spherical, but rather flat, which gives a significantly larger surface area per unit volume. The proportion of the water volume bound by adsorption is perhaps 20%. Because of clay particles' large surface area, the clay content of soil is of decisive importance for the amount of water bound by adsorption.

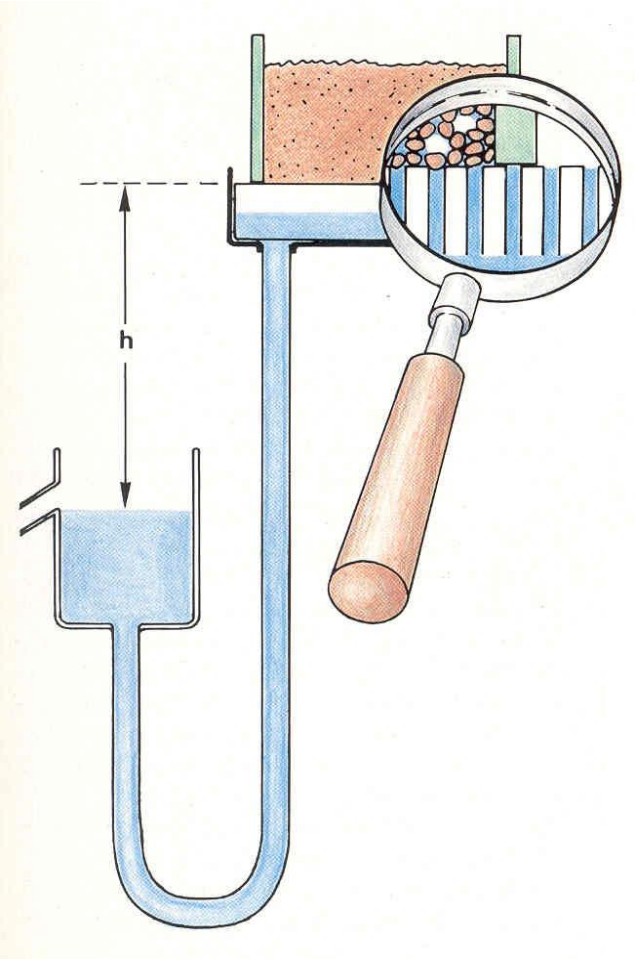


Fig. 10 Device for determining the relation between binding pressure and water content in a soil sample. The sample is placed on a ceramic plate that is in contact with a reservoir of water. When the water pressure in the reservoir of water is lowered by lowering the reservoir, and thus the free water surface in the reservoir, water is sucked from the soil sample through the pores of the ceramic plate. When the soil sample stops draining, its binding pressure = -h. The corresponding water content can be determined by weighing.

The ceramic plate cannot admit air since the pores of the plate cannot be drained by the moderate suctions generated by the apparatus. The pores of the plate which are not in contact with the water in the soil sample are blocked by the capillary forces that are generated at the interface between water and air.

To empty a pore of water, one must apply a suction that is larger than the binding forces associated with that pore (eq. on p. 20). In the laboratory, suction can be applied to a soil sample using the apparatus in Fig. 10. The more the free water surface is lowered, the more water is drained from the sample and its water content decreases. By measuring how much water is drained as the suction increases, one can determine the relationship between binding forces and water content in the soil sample. This relationship is referred to as the water retention curve, or the pF-curve, and is a very informative soil attribute.

In a dry soil, the water is bound tightly. The pressure in the water is a large negative number, corresponding to perhaps a water column of several hundred meters. The concept pF has been introduced to express the water's pressure in a manageable fashion for both large and small negative pressures. By pF, one means $^{10}\log(-\psi)$ where ψ = water's pressure expressed in terms of cm of a water column. pF 1 and 4 thus denote respective water pressures of -10 and -10,000 cm.

The terminology for the pressures of soil water and groundwater can be confusing. We have assigned the atmosphere a pressure of zero. Groundwater's pressure is then positive, and that of soil water is negative. To make it easier to speak of negative pressures, we sometimes use the word suction. A large suction means that the pressure is a large negative number.

The largest suction that the apparatus in Fig. 10 can generate is the pressure of the atmosphere, which is equivalent to a 10 m column of water (pF 3). To drain the soil sample further in the laboratory, one increases the pressure in the air on the upper side of the soil sample, or one can also take advantage of the equilibrium that is established between the negative pressure in the water and the vapour pressure in the surrounding gas.

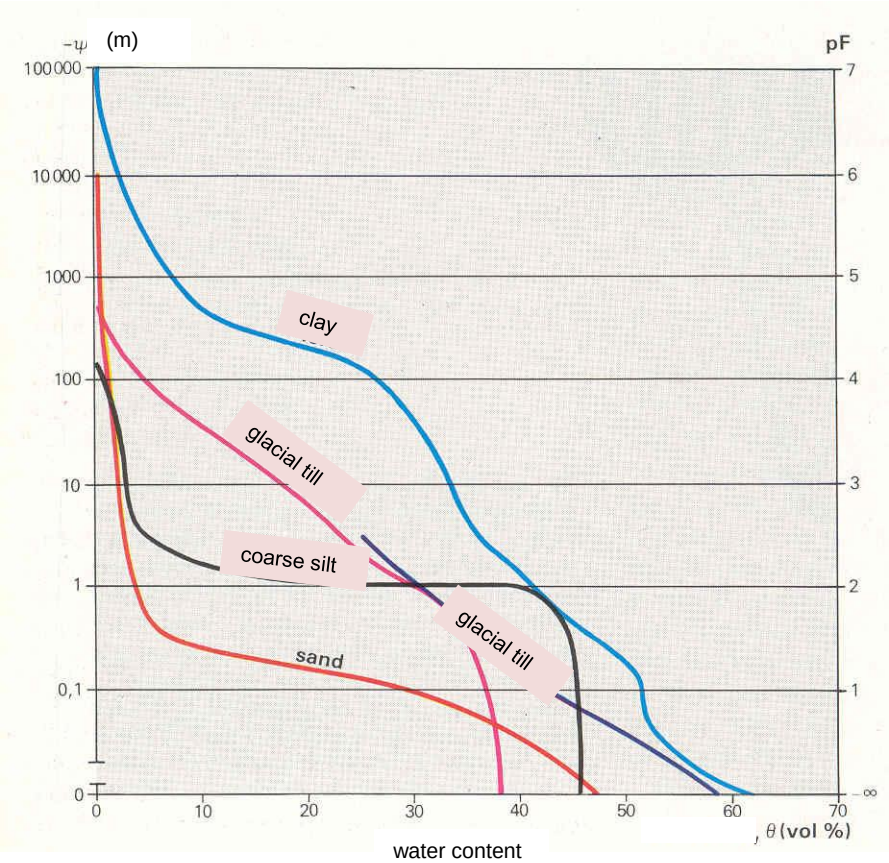


Fig. 11 A soil's water retention properties are illustrated by the water retention curve or pF-curve. In the coarse-grained sand, a large portion of the pore volume is already drained once a small suction has been applied, while fine-grained clay retains much water even when a large suction is applied.

The well-sorted, coarse silt has pores of a very uniform size. These drain at a specific suction where the water content diminishes drastically. In the till and the clay, there is a range of pore sizes, which is why the water content decreases gradually as the pressure becomes more negative.

From the water retention curve, one can learn something of the pore size distribution in a soil. More clay in a soil increases the proportion of small pores, and means that the water is more tightly bound at any given water content. Sorted soils have a relatively uniform pore size so that much of the water is held within a narrow pressure interval. This means that many of the pores empty at a certain suction, creating a plateau in the pF-curve (cf. the coarse silt in Fig. 11). This situation results from the plateau in the texture. Smaller plateaus in pF-curves can also result from the structure of the soil.

When making a pF-curve, thin soil samples are used. This means that one can ignore the pressure differences that exist at different levels in the sample. In a 'tall' soil sample with a large vertical extent, this pressure difference cannot be ignored. Water in the upper portion of the sample experiences a more negative pressure than the water in the lower portion of the sample. When the free water surface in the reservoir of the apparatus is level with the sample's lower surface (h = 0 in Fig. 10), the suction at the upper surface of the sample is equivalent to the height of the soil sample, and that suction can drain larger pores.

When an initially saturated soil sample is allowed to drain freely until the flow of water stops, the pressure in the water at the lower surface of the sample is equal to that of the atmosphere. This level corresponds to the free water surface. At that time, the suction at any level in the soil sample is equivalent to the distance to the lower surface (assuming that there is no evaporation from the upper surface of the sample. The reason why so little water drains from the tank with sand or sand and gravel was that the forces holding the water in the pores (the binding pressure) was greater than the height of the tank, so that not even the uppermost pores drained.

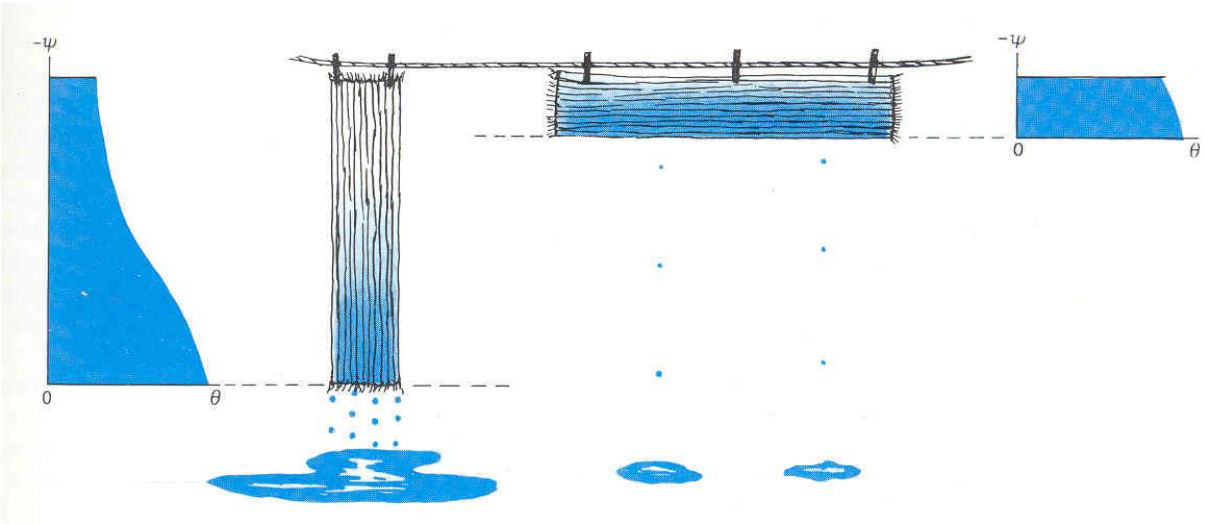


Fig. 12 When the dripping stops, the mat on the left is drier than that on the right. At that point, the suction at any level in the mats is identical to the distance to the mat's lower edge, where the water pressure is the same as atmospheric pressure. In the upper portion of the left-hand mat the suction is great, and the water content low. In the right-hand mat, most pores are still filled at the modest suction found even at the upper edge of the mat.

The groundwater level influences the soil water content in a similar fashion. When there isn't any vertical flow, the suction at a given level is the same as the distance down to the groundwater level. (This applies when the distance is relatively small; up to about a meter in till soil, and just several decimetres in sand.) Shallow groundwater therefore gives high soil water content. Among other things, this means that the groundwater level rises rapidly and far even with just a small addition of water.

In the above example, the "free water surface" can be replaced with "groundwater level" and "soil sample" by "soil above the groundwater level". When there isn't any vertical drainage, i.e. at drainage equilibrium, then the soil water's suction at any particular level is equal to the distance to the groundwater level. The location of the groundwater level is thus of great significance for the water content in the unsaturated zone. This is the case for relatively shallow groundwater. If the groundwater level is deep, equilibrium is never established between the soil water and the groundwater level. The drainage of water from the soil stops because the capacity of the soil to transmit water is practically zero when the suction reaches a certain level. In a sand soil, this occurs already when the suction is about 0.5 m, in a moraine this occurs at 1-2 metres of suction, and in a silt or clay, at about 3 metres of suction. These values denote the greatest depth at which the groundwater level influences the soil water content near the soil surface. (In the apparatus in Fig. 10 the situation is different, since the soil sample drains through a tube that is filled with water, and thus does not lose its ability to transmit water.)

Plant available water

Two concepts which are often used to describe the hydrologic status in the root zone are field capacity and wilting point.

By *field capacity*, one means the water content in a previously saturated soil after free drainage, such as might result from lowering the water table. This is the greatest water content the soil is capable of holding against the force gravity. During, and shortly after a large rain event, the water content can exceed the field capacity. After a period of drainage, the downward flow essentially ceases, and the water content remains almost constant at the field capacity. (Continued decreases in the soil's water content after that results primarily from water uptake by plants and evaporation from the soil surface.)

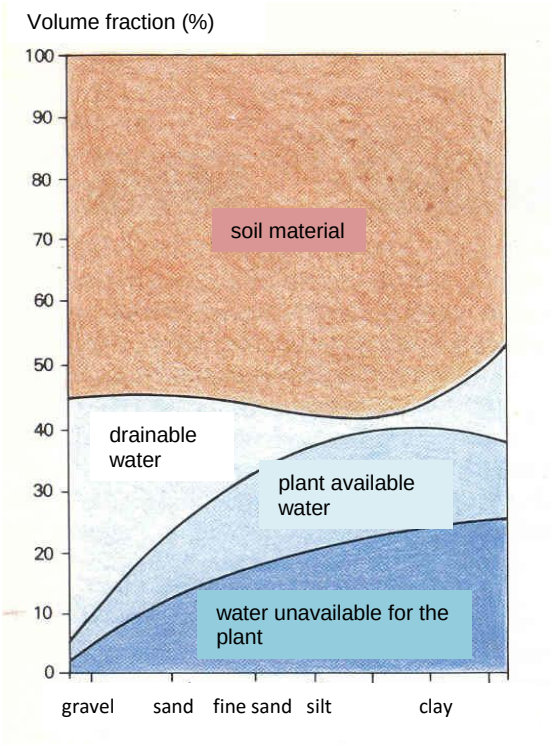


Fig. 13 During dry periods, plants must take advantage of the plant available water in the root zone, i.e. the difference between the water content at field capacity and the wilting point.

The diagram shows the water retention qualities of sorted soils with different particle sizes. The increase in the drainable water volume of the finest soils results from the existence of large pores created by the soil structure.

From the discussion of water retention, one can determine that the field capacity depends on the depth to the groundwater level (for example in Fig. 12). The deeper the groundwater level, the more water drains from the root zone (within certain bounds). In general one can think of the groundwater level of being up to a few meters depth. A more precise definition of the field capacity is the "soils water content a pF 2.0", i.e. at a suction of 100 cm. This definition is applicable to individual soil samples. In a root zone above shallow groundwater, the equilibrium water suction at any level is equal to the distance to groundwater, and therefore varies with depth in the root zone. Thus the definition of field capacity by reference to pF 2.0 is not applicable.

One way to determine the field capacity of the soil in the field is to generously water the soil surface, and then measure the water content after a period of time. The field capacity is the soil water content after it has fallen to a relatively constant value, most likely after a few days. (See page 32 for more on this.)

The wilting point is the water content at which the water uptake by plants ceases as the soil dries out. Roots create suction to draw water up into plants. At the wilting point, the water remaining in the soil is so tightly bound that the plant's suction is not sufficient to pull out any more water. The wilting point is usually assigned a value of pF 4.2, which corresponds to a suction of 150 m. The wilting point is largely controlled by the amount of adsorptively bound water in the soil. The finer grained a soil is, the larger the water content at the wilting point.

The difference between field capacity and the wilting point corresponds to the volume of water that plants can make use of during dry periods. This is usually referred to as the plant available water and can be determined from the pF curve. The amount of plant available water varies markedly between different soils. For the sand soil in Fig. 11, this amounts to just 3% (15 mm in a 0.5 m root zone). In that soil, much of the water has already drained before reaching field capacity. For the coarse silt depicted in the figure, 40% of the soil volume consists of plant available water (200 mm in 0.5 m), but in the clay shown in the figure, the plant available water is relatively small, 23%. Here most of the water is still held in the soil at the large suction corresponding to the wilting point. It should be noted this is an extremely stiff clay, unaffected by soil processes. The plant available water in most Swedish clay soils is considerably larger, which makes them particularly favourable from the standpoint of agriculture.

In gardens (and in flowerpots) one can manipulate the soil's particle size distribution, and thereby create a soil that can sustain plants through extended periods of drought by means of their large reservoir of plant available water.

The driving forces behind the flow of soil water and groundwater

In the discussion of water retention, we have primarily considered equilibrium conditions, i.e. situations where any flow has ceased. Thus we have ignored another force that can be said to contribute to holding water in the soil, namely the force of friction. The force of friction that is generated by flow in a soil pore is proportional to the velocity of the flow. (This is the case as long as the flow is laminar, i.e., it occurs in layers without mixing in the pore. When there is turbulent mixing, the forces of friction increase as the square of velocity. This applies to the aerodynamic drag on a car or the bed friction in a stream). The proportionality between velocity and the force of friction forms the physical foundation for Darcy's Law, which expresses the relationship between driving forces and flow in the soil. This relationship, which we will consider

in some detail, is the basis for almost all calculations of water flow in the soil water and groundwater zones.

Groundwater flow can, as we shall see, be treated as a special case of soil water flow (the pressure is greater than the atmosphere's and the water content is equal to the porosity). For this reason when discussing flows, we often let the word soil water refer to both soil water and groundwater.

The flow of soil water and groundwater is driven by the forces of gravity and pressure. The driving forces are balanced by the frictional forces generated by flow. Since the frictional forces are proportional to the velocity of flow in a pore, flow is proportional to the driving forces.

A good tool for understanding and describing soil water and groundwater flow is the concept of potential. In physics, the term potential refers to the work (= force · distance) that is required to move a body to a given position or state.

A spring potential is, for example, the work needed to stretch a mechanical spring to a particular position. Gravitational potential is the work required to lift a body from a certain reference elevation to a new elevation. In a body attached to the stretched spring, as in the raised body, there is, by definition, a force that strives to move the body towards a lower potential. The decline in potential energy which occurs in the body when that force is allowed to work is equal to the sum of the increases in the body's kinetic energy and the frictional energy (heat) that is generated.

The *soil water potential* is defined in a similar way. The *total potential* in the soil water is the sum of the gravitational potential and the pressure potential. There is also an osmotic potential, but one usually does not need to consider that.

A body's potential is usually expressed as energy per unit mass, i.e. J/kg. When speaking of water flow the units of energy per unit volume (J/m^3), and energy per unit weight (meters of a water column) are used. The latter terminology, which is very practical, is arrived at by dividing the term J/m^3 by $\varrho \cdot g$ where ϱ = water's density (kg/m^3) and g is the gravitational acceleration (9.81 m/s^2).

The *gravitational potential* for soil water is the amount of work required to move a unit mass of water from a reference level to the level where that water is. Expressed in terms of meters of a water column, the water's gravitational potential is its elevation above the reference level.

Pressure potential is defined as the "work required to transform a unit mass of water from atmospheric pressure to the pressure found in the water". This is simply the pressure of the water compared to that of the atmosphere. The pressure potential is thus negative in the soil water zone, zero at the groundwater level, and positive in the groundwater zone. Pressure potential in the unsaturated zone is often called the *binding potential*. Pressure potential at a point is the height above (or negative distance below) that point where a free water surface in equilibrium with the water at the point would stand. In the unsaturated zone the water pressure can be measured with a *tensiometer*. If the water tank in Fig. 10 is removed, so that only the tube remains, the water level in the tube will adjust to the pressure in the soil sample. This is the principle of a tensiometer. The water level in this case is lower than the measurement point in the soil, and thus the height is negative (pressure potential <0). In the saturated zone pressure is measured with a tube that is open at the top and has an intake only at the soil level being measured, called piezometer.

The soil water's total potential at a point is thus the sum of the point's height above a certain reference level (gravitational potential) and the elevation of the water level in a measuring device relative to the measurement point (pressure potential). The total potential is therefore also the elevation of the water level in the measuring device relative to the reference level.

The *osmotic potential* arises when a substance is dissolved in a liquid. A solution that has different concentrations of the dissolved substance in different portions of the solution will eventually have the same concentration in the entire solution. The migration of the dissolved substance's molecules from the areas with high concentration to areas with low concentrations is called molecular diffusion. If one divides the liquid with a semi-permeable membrane that does not allow the dissolved molecules to pass, but only the solvent, then the solvent will move through the membrane from the side with the higher concentration of solvent to the side which has the lower solvent concentration, i.e. from the side with the lower concentration of the dissolved substance to that with the higher concentration. The counter-pressure that one must apply to stop the migration of the solvent is the solution's osmotic potential. The osmotic potential is of significance for water uptake by the roots, which function as such a membrane. If the soil water has a high salt content, then the osmotic potential is low (a large negative number) and the plant can have difficulty in generating a sufficiently large suction to extract water from the soil, even if water is available. For flow within the soil, though, the osmotic potential is not of practical importance, since the effect requires the presence of a semi-permeable membrane.

Calculation of saturated and unsaturated water flow

Darcy's law was formulated in 1856 by the Frenchman Henri Darcy. It was originally intended for groundwater flow, but has since been shown to apply even to unsaturated flow. The law states that flow between two nearby points in the soil is proportional to the total potential difference between those points.

$$Q = - K \cdot A \cdot d\phi / dx$$

Q = water flow (m^3/s)
A = cross-sectional area of the soil layer (m^2)
K = the soil's *hydraulic conductivity* (m/s)
ϕ = the water's total potential (m)
x = distance (m)
$d\phi / dx$ = the change in total potential per unit distance, *potential gradient* (m/m, i.e. unitless)

In the equation, Q is the water flow, i.e., the volume of water that has moved in a unit time across a cross section of the soil with an area of A. The term can also be used to give the flow of water per unit area of soil Q/A. This has the units $m^3/(s \cdot m^2)$ = m/s, or in the case of vertical flow, for example, litre/ (day $\cdot$ m^2) = mm/day.

The minus sign in the equation is a result of flow occurring in the direction of decreasing potential, which means that Q is positive (directed in a positive direction along the x-axis) when the potential gradient is negative.

Table 1 *Approximate values of saturated hydraulic conductivity for some soils.* (From Fagerström & Wiesel, 1972).

Soil	*Hydraulic conductivity at saturation (m/s)*	
Fine gravel		$10^{-1} – 10^{-3}$
Coarse sand		$10^{-2} – 10^{-4}$
Medium sand		$10^{-3} – 10^{-5}$
Fine sand		$10^{-4} – 10^{-6}$
Coarse silt	Coarse till	$10^{-5} – 10^{-7}$
	Sand till	$10^{-6} – 10^{-8}$
Fine silt	Silt till	$10^{-7} – 10^{-9}$
	Clay till	$10^{-8} – 10^{-10}$
Clay	Till clay	$<10^{-9}$

Note. These values are valid for the conductivity caused by soil particle distribution. In fine textured soils the saturated conductivity is often larger in the upper soil layers, due to soil structures like dry cracks, root channels, worm holes, etc.

For unsaturated flow, the expression implies that flow tends to occur downwards (towards lower gravitational potential), and to drier regions (towards lower pressure potential). These forces combine when water infiltrates into a dry soil, but they oppose each other when water evaporates from the soil surface.

The soil's hydraulic conductivity

The hydraulic conductivity is a measure of the soil's ability to transmit water. This depends on the soil's pore size distribution and the composition of the pore system, as well as on the soil's water content. The conductivity of a soil is greatest when the entire pore volume is filled with water, i.e. under saturated conditions. The hydraulic conductivity at saturation is sometimes called the permeability. This latter word more correctly denotes a soil's capacity to transmit a liquid or gas, for example oil or air. This value is a property of the soil that does not depend on the properties of the liquid or gas. The permeability has the unit m^2.

The soil's hydraulic conductivity is a key factor when it comes to calculating the flow of water in the soil. The hydraulic conductivity can be determined from Darcy's Law by measuring the flow and potential gradient. This can be determined on a soil sample in the laboratory as well as for an entire soil layer in the field. One normally determines the saturated conductivity, that is, the hydraulic conductivity under saturated conditions. In the field this can be done by means of so called pumping tests where one relates the amount of water pumped from a well to the lowering of the groundwater level around the well.

Gradient

Gradient is a mathematical term which describes how the size of something changes in space. Temperature gradients in the x-direction, for example, define how many degrees the temperature (T) changes when one moves one meter in the direction of the x-axis. If the temperature increases when moving in the direction, the change (dT) is positive, and the temperature gradient (dT/dx) is positive. If the temperature decreases, the gradient is negative.

The potential gradient of soil water describes, in a similar manner, how soil water potential changes in a specific direction in the soil. There is a force working on water that seeks to move it from high to low potential. If the potential gradient in the x-direction is negative, i.e. if the potential decreases with an increase in x, then flow occurs in the direction of the x axis. One usually defines this as a positive flow in relation to the x-axis. In order for a negative potential gradient to generate a positive flow, one must set a minus sign in front of the right side of Darcy's Law. In simple applications of the equation to saturated flow, though, the direction of flow is easy to keep track of and the minus sign can often be omitted. In the case of unsaturated flow though, the direction of flow might be what is searched for, which is why the minus sign is good to have in the equation. In flow modelling the minus sign is also necessary.

Let us see how the conductivity depends on the pore size and water content. The velocity of water flow in a narrow tube is defined by the following relationship

v = constant · $r^2 \cdot d\phi / dx$
v = average velocity
r = the radius of the tube

The velocity of the water for a given driving force thus increases rapidly with the radius of the tube. This is because friction's ability to retard the flow decreases as the radius increases. The flow of water (Q) through the tube is given by the velocity (v) times the cross-sectional area (A)

$$Q = v \cdot A$$

Since both v and A increase with r^2, the flow of water will increase as a function of r^4. Thus a tube with a radius of 1 mm conducts as much water as 10,000 tubes with a radius of 0.1 mm. Assume that the soil consists of narrow, parallel tubes with very thin walls oriented in the direction of flow. If the tube diameter increases ten-fold, then the total flow across the cross-sectional area of the soil increases by a factor of 100. (Water flow through every tube is 10,000 times greater, but a given cross section of the soil contains 100 times fewer tubes.)

This example calculation implies that the saturated hydraulic conductivity in a real soil increases quickly with the size of the pores, and therefore with the particle size in the soil (Table 1). In addition to the particle size, the structure also is of great significance. A water-filled worm hole, or a water filled crack transmit, as demonstrated above, much more water than the corresponding cross-sectional area of "ordinary" soil pores.

A prerequisite for a pore transmitting water is, of course, that the pore contains water. This requires, in turn, that the water pressure is not below the binding potential of that pore. As a soil

dries, it empties progressively smaller pores of their water. Thus the conductivity decreases rapidly, partly because the water content, and thus the water-filled portion of the soil's cross-sectional area, decreases. But also partly, and in fact primarily, the conductivity decreases because the largest pores drain first. These pores have contributed to a large portion of the total flow.

A sand soil and a clay soil can have approximately the same porosity, perhaps 50%. Thus they have approximately the same cross-sectional area for water flow at saturation (if we assume that all water contributes to flow, that is, ignoring water bound by adsorption). The saturated conductivity is larger in the sand soil because of its larger pores. After a large suction has been applied, the large pores have drained. The clay soil, with its many small pores now contains the most water and has a larger conductivity (cf. differences between two sand soils in Table 2).

Table 2. *Measured hydraulic conductivity (K) and water content (θ) in two sand soils under saturated and unsaturated conditions, respectively.* (From Jansson, 1980.)

Pressure (m)	*Soil 1 (coarse) (Well sorted medium sand)*		*Soil 2 (fine) (Non-sorted sand containing fine particles)*	
	θ (vol %)	K (m/s)	θ (vol %)	K (m/s)
0	39	$2.4 \cdot 10^{-4}$	59	$0.7 \cdot 10^{-4}$
-0.25	10	$3.9 \cdot 10^{-6}$	36	$3.9 \cdot 10^{-6}$
-0.50	6	$1.1 \cdot 10^{-8}$	22	$1.7 \cdot 10^{-7}$

At saturation the conductivity is greatest in the coarser soil. The considerably larger pore volume in the finer soil cannot compensate for the contribution from the largest pores in the coarse soil. In both soils the conductivity decreases quickly with increasing tension, i.e. when the largest pores are emptied. At -0.5 m pressure there are few water filled pores in the coarse soil. The conductivity is now greatest in the fine textured soil.

It should be noted that in this case there are relatively small suctions. In sand soils like these the water movement is much reduced during dry periods, when the pressure can amount to minus many tens of meters of water and the water content becomes very small. This also makes water uptake by plants more difficult and the wilting point (pressure = -150 m) is seldom reached.

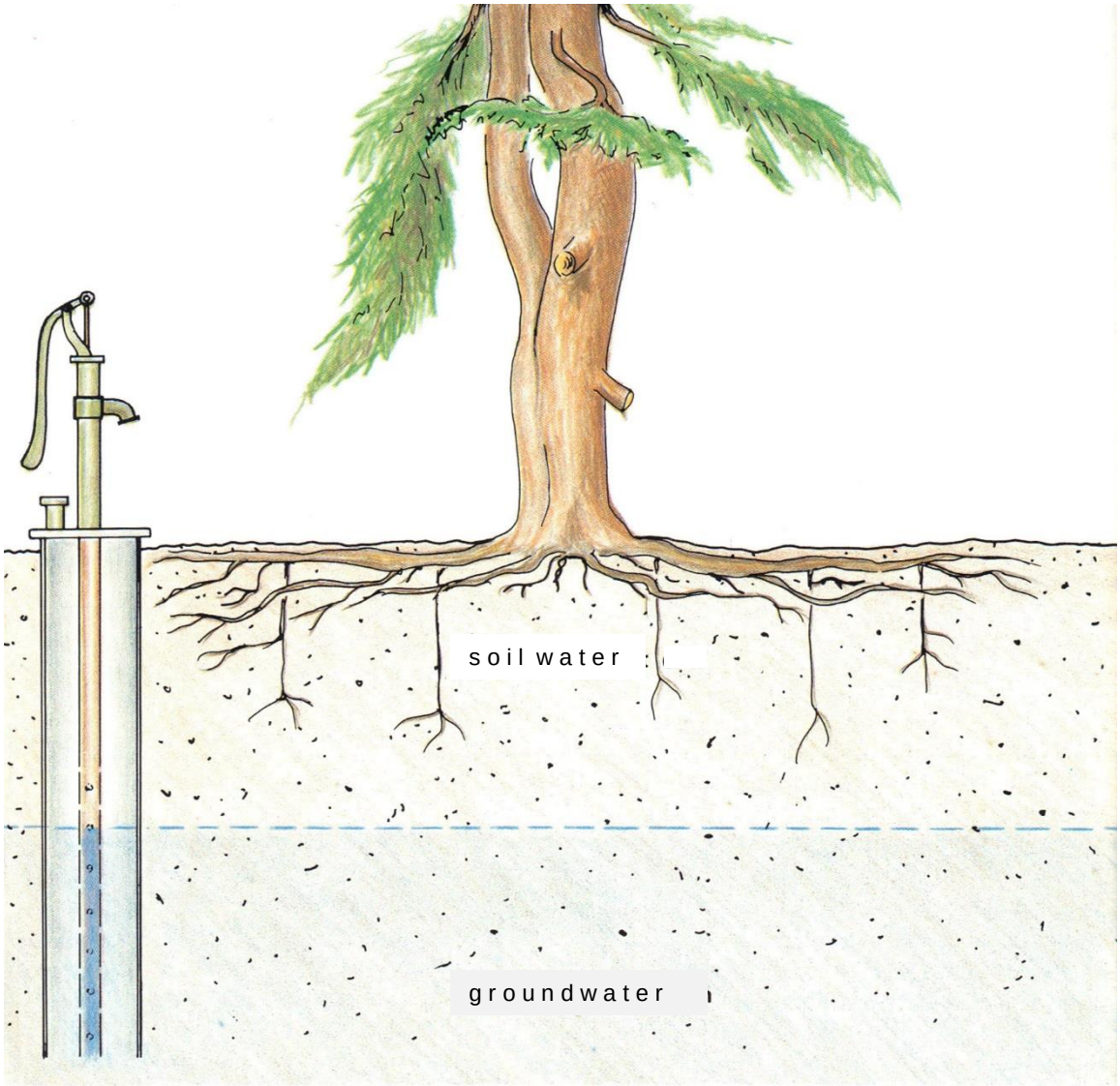


Fig 14 a. The groundwater surface is the level at which the water pressure equals that of the atmosphere. In the groundwater zone all pores are filled with water and the pressure is higher than the atmosphere's. In the soil water zone there are both water and air in the pores. The pressure of water is lower than that of the atmosphere.

The unsaturated conductivity is related partly to the saturated conductivity, which serves as the starting point when a soil dries. Partly the conductivity is related to the pF-curve, which determines how much water is left at a particular suction. This relationship is used when

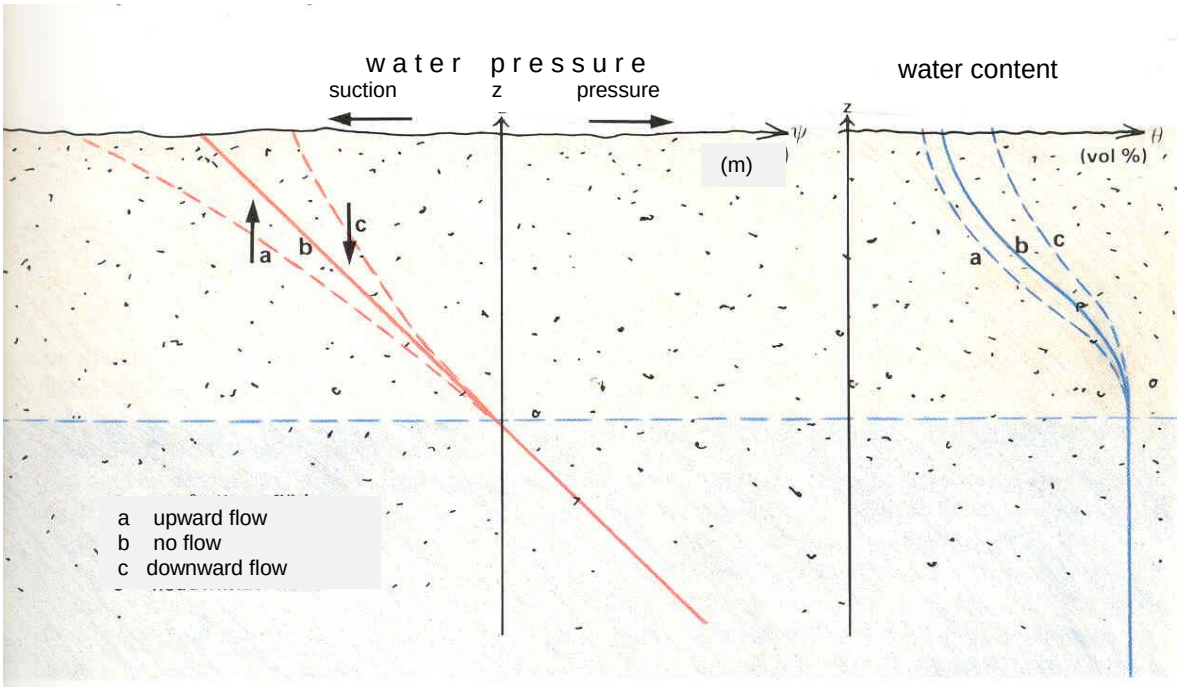


Fig. 14 b. Water pressure, water content and vertical water flow in an unsaturated soil.

(1) $Q = -AK\, d\phi/dz$
(2) $\phi = \psi + z$
(3) $dz/dz = 1$, i.e. $d\phi/dz = d\psi/dz + 1$
(4) b) $d\psi/dz = -1$, which gives $d\phi/dz = 0$,
i.e. $Q = 0$, no flow
(5) a) $d\psi/dz < -1$, which gives $d\phi/dz < 0$,
i.e. $Q > 0$, flow upward
(6) c) $d\psi/dz > -1$, which gives $d\phi/dz > 0$,
i.e. $Q < 0$, flow downward

(1) Water flows from higher to lower total potential.

(2) Water's total potential = pressure potential (pressure) + gravitational potential (elevation).

(3) Gravitational potential, i.e. elevation, always increases at a rate of one meter for every meter moved upwards.

(4) In line b) the pressure decreases at a rate of one meter for every meter moved upwards (the line has a slope of –45°). The pressure potential decreases just as much as the gravitational potential increases. The total potential is therefore constant and no flow occurs.

(5) In curve a) the soil surface has dried and the pressure in the soil water zone decreases faster with elevation gain than in b). The pressure potential's decrease with elevation prevails over the increase in gravitational potential. The total potential decreases upwards, and the flow of water is upwards.

(6) In curve c) water has been added at the soil surface. In this case, pressure decreases more slowly with elevation gain than in b). The pressure potential's decrease with elevation gain does not fully compensate for the increase in gravity potential. The total potential increases upwards and the flow of water is downwards.

One cannot determine the vertical direction of water flow in a similar fashion from the change in soil water content with depth. It is the pressure and not the water content that determines the flow direction. One can, however, calculate the pressure from the water content by means of the soil water retention curve.

calculating the unsaturated hydraulic conductivity at different suctions, since direct measurements of the unsaturated hydraulic conductivity's magnitude often do not exist.

The differences in how conductivity changes with water pressure mean that the drainage of a sand soil and a clay soil proceed in different ways. In a sand soil, the drainage essentially ceases at a certain suction. In a clay soil the drainage is slower, but it decreases more gradually. Even at very great suctions, flow can still occur in clay. It is this difference that means that the concept of field capacity is better defined in a coarse-grained soil than in a fine-grained soil.

Water pressure and vertical flow in unsaturated soil

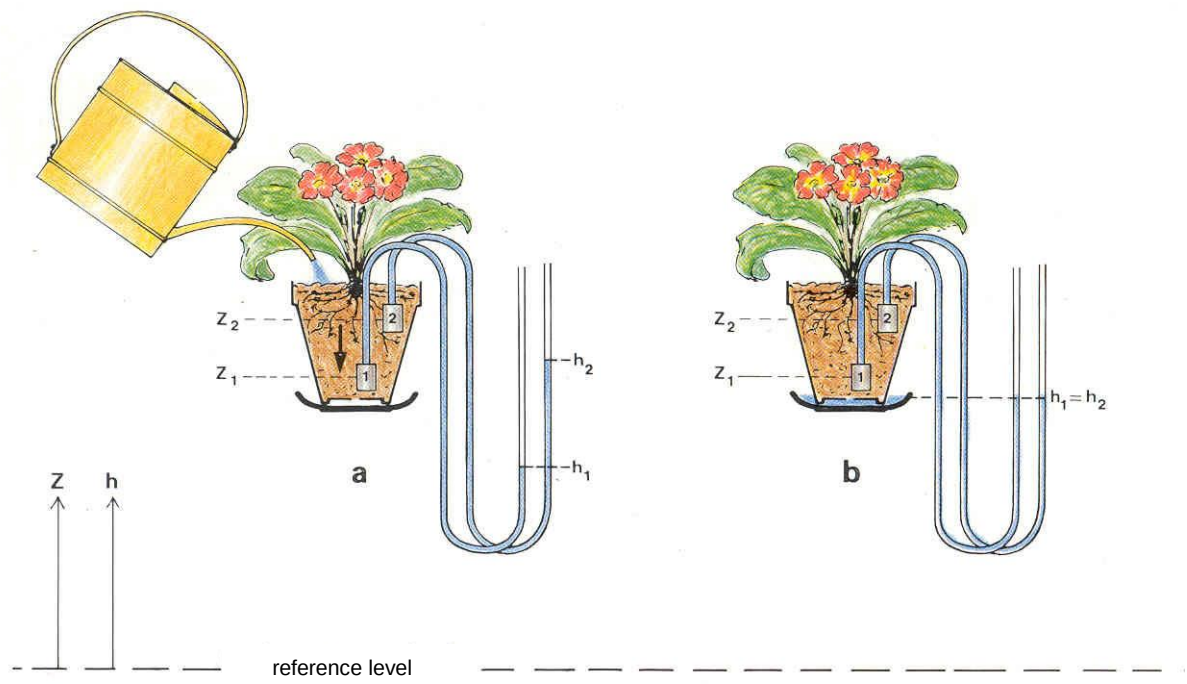
Let us consider some applications of Darcy's Law.
Consider Fig. 14 which shows a soil with a horizontal soil surface and a horizontal groundwater level. When there is not vertical flow, the total potential is identical at all levels ($d\phi/dz = 0$). Since the gravity potential (= elevation) increases linearly with elevation, the pressure potential must decrease at a corresponding rate, so that the sum remains constant. In the unsaturated zone the pressure is negative and equal to the distance to the groundwater level (cf. page 23). In the saturated zone the pressure is positive, and identical to the depth below the water level. The pressure potential is a straight line with a slope of –45° in the diagram.

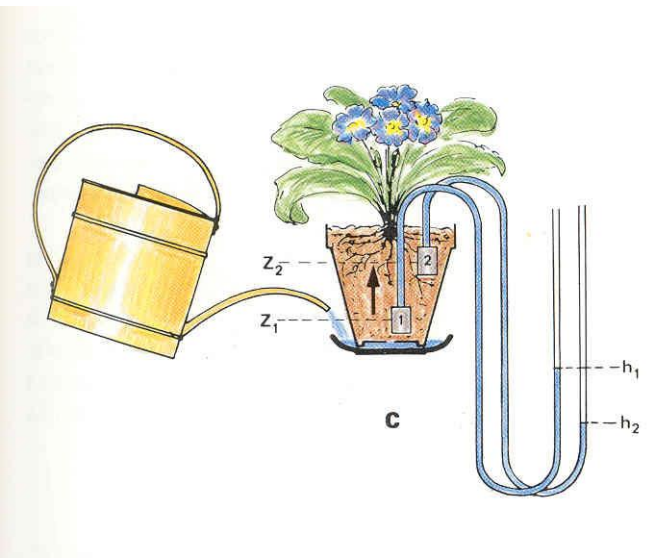
The water content varies in a different way. Below the groundwater surface the water content is constant, and equal to the porosity. Above the groundwater surface, the water content varies with elevation in accordance with the soil moisture characteristic (drawn with a linear scale in the figure). Here the pores begin to empty at a suction that corresponds to say, some dm above the groundwater surface. This pressure, the air-entry pressure, plays a major role in soil water flow at conditions near saturation. It is first at greater suctions that the hydraulic conductivity begins to decline markedly.

If water is added to the upper soil layer through infiltration, then the soil profile is no longer in equilibrium. Water pressure at the surface increases (becomes less negative) which increases the total potential at the surface, and a downward flow begins ($d\phi/dz > 0$). If water is instead removed by evaporation from the soil surface or water uptake by plants, the water pressure at the surface decreases (becomes more negative). The total potential at the surface decreases and an upward flow will begin from deeper levels ($d\phi/dz < 0$).

The pressure forces in the unsaturated zone, primarily generated by capillary forces, can thus work downwards as well as upwards, and in other situations even sideways. The force of gravity always works downwards, as is well-known. (Even in the groundwater zone pressure forces can work in all directions, as we will see later.)

For every water pressure in the soil, there is corresponding water content in accordance with the soil moisture characteristic for the soil in question. The water content in the different flow situations has been drawn in Fig. 14 to represent the water content at equilibrium and an imagined extension of that curve.





The relationship between total potential at different levels and the direction of flow is also illustrated for a flower pot in Fig. 15. Note that the difference in total potential in the soil water is not proportional to the magnitude of the flow, since the conductivity depends on the suction. A specific potential difference in a soil under moist conditions will therefore give rise to a much larger flow than the same total potential difference will generate under dry conditions.

Fig. 15 Soil water pressure is measured with tensiometers. Here these consist of a ceramic cup that is connected to a free water surface. The height of the free water surface above or below the measurement point gives the pressure (or suction) at that point, i.e. pressure $\psi = h - z$ where h and z are heights above a reference level. Under unsaturated conditions, $h < z$ and the water pressure is negative relative to that of the atmosphere. Total potential = pressure potential + gravity potential $\phi = h - z + z = h$

The height of the free water surface thus denotes the total potential
a) The total potential decreases downward and the flow is downwards
b) The total potential is constant and there is no flow
c) The total potential decreases upward and the flow is upwards

A coarse layer diverts unsaturated flow

An example of how conductivity changes in different ways in different soils as the suction changes is the effect of a coarse layer in a fine-grained soil. If that layer slopes somewhat, unsaturated flow vertically downwards will be diverted to the side (Fig. 16). This is not because the coarse layer transmits the water to the side, but because the water cannot penetrate into the coarse layer. Instead the water flows in the fine-grained soil along the upper edge of the coarse layer. At the suctions that can exist in soils after a moderate addition of water, the coarse pores do not fill with water, which makes the conductivity small in the coarse layer and water flows in the fine soil with a higher conductivity. If the input of water increases significantly, so that nearly saturated conditions are created, the water runs downward into the coarse layer. See experiment in Fig. 17.

This principle has been employed when covering old garbage dumps with soil. To keep infiltrating rainwater from getting into the garbage and then transporting pollutants into the groundwater, the garbage has been covered by a coarse-grained layer. A fine-grained layer is then placed above this.

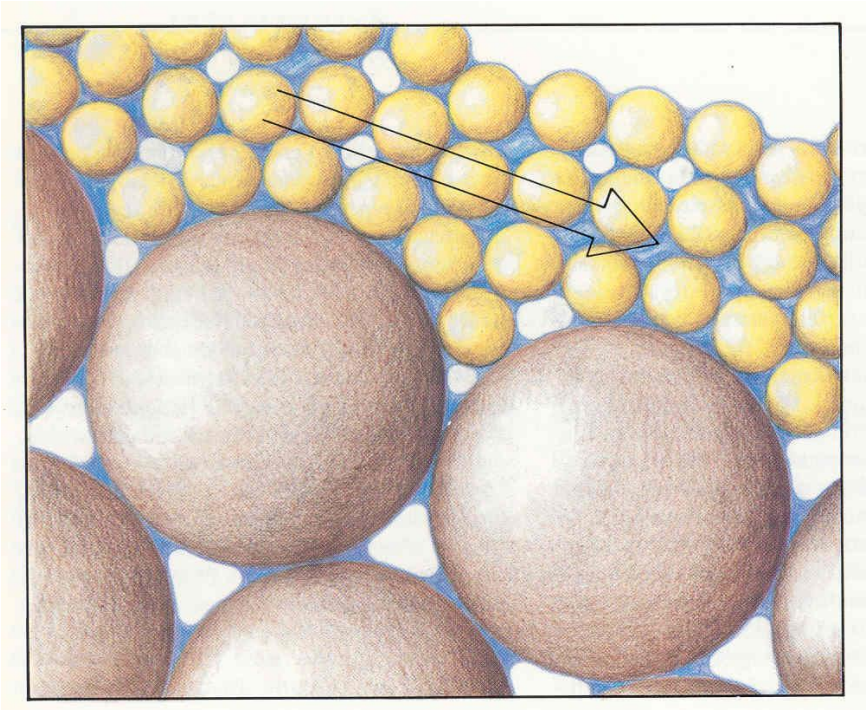


Fig. 16 Large pores transmit water much more efficiently than small pores. But pores that do not contain water don't transmit any water at all. When the soil water pressure is somewhat low, the large pores are emptied. In the coarse-grained soil there are few water-filled pores, but in the fine-grained soil there are many. A moderate flow of water from above, caused by, for example, infiltration, is diverted to the side into the lower portion of the fine-grained soil since the coarse-grained soil is not capable of transmitting the water. The large, unfilled pores serve to block the flow.

If the coarse layer is horizontal, there won't be any diversion of the flow. Instead a layer near saturation will build up in the fine-grained soil. When the pressure increases sufficiently, a sudden drainage of the water through the coarse layer will occur. This breakthrough can be very local, due to horizontal discontinuities in the soil and possible undulations in the upper surface of the coarse layer.

A hole that is dug in the unsaturated zone diverts unsaturated flow in a similar fashion. For water to come out into the hole, which can be treated as an extremely large pore, the water pressure must be equal to that of the atmosphere. The water content on the up-gradient side of the hole increases at first due to the flow of soil water. This increases the water pressure, thereby decreasing the potential gradient which drives that flow. If the cross-section of the hole in the direction of the flow is small, say 0.5 m, then all flow is diverted around the hole.

If the cross-section is large, say several meters, diversion will continue to the sides. In the middle of the hole, saturation can occur, at which point water will begin flowing into the hole. The flow into the hole, though, will be less than the flow across that same cross-section without a hole.

We have previously emphasized that flow occurs in the direction of diminishing total potential. This actually only applies to a homogeneous soil with the same conductivity in all directions. Fig. 16 and 17 are examples of situations where this is not the case. Because the conductivity of gravel is so much less than that of the sand (when the input of water is moderate) the small component of the potential gradient along the boundary of the sand layer is much more effective for transmitting flow than the larger component perpendicular to the boundary, into the gravel. Another example is groundwater flow in a sand soil that is overlain by clay. Even if the total potential gradient is directed up into the clay, the flow is still essentially parallel to the boundary between the two soil types.

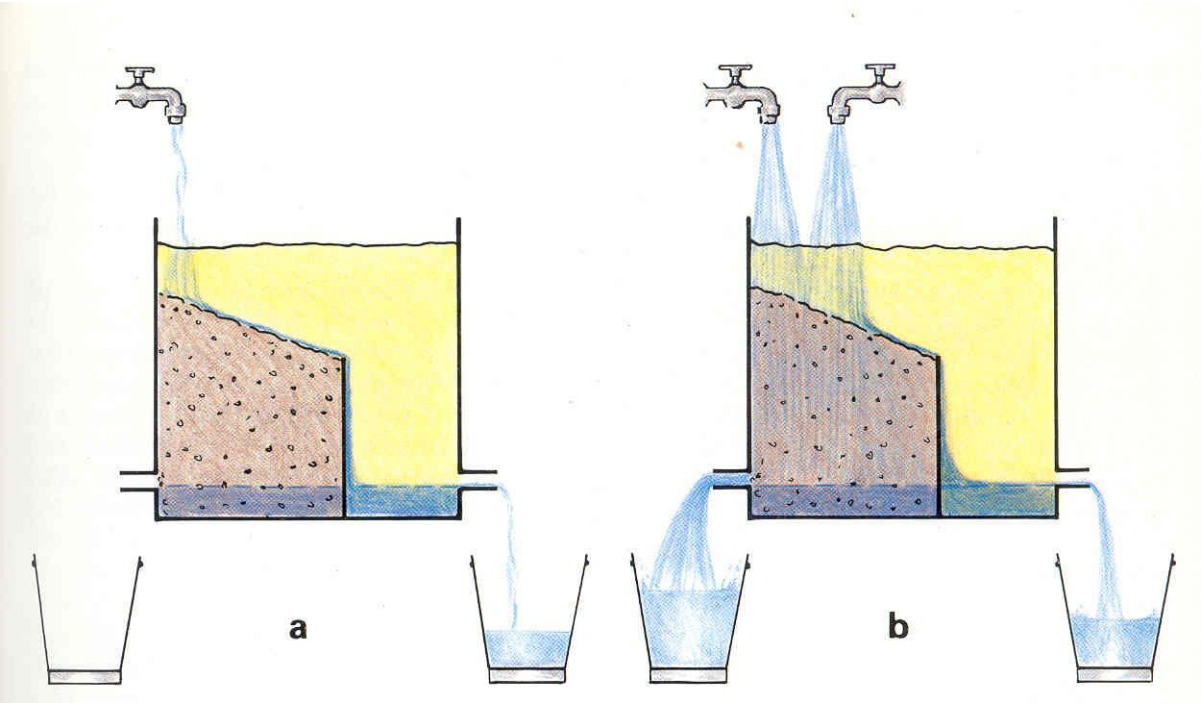


Fig. 17 When the input of water is small, the water pressure is rather low and the unsaturated conductivity is much larger in the sand than in the gravel. The flow is diverted laterally along the underside of the sand, a. When the input of water is large the soil is almost saturated and the large pores in the gravel fill with water. The conductivity of the gravel becomes large, and almost all of the flow goes through the gravel, b. Under saturated conditions the conductivity in the gravel is much larger than in the sand.

Water flow in saturated soil

When we have discussed unsaturated flow, we have primarily considered flow directed either upwards or downwards. These flow directions usually dominate in the soil water zone. Even in the groundwater zone vertical flow occurs, or flow with a vertical component. But here the horizontal component is usually dominant.
In the most common applications of Darcy's Law under saturated conditions, one assumes that the total potential gradient at every level is the same as the slope of the groundwater surface relative to the horizontal plane. The vertical differences in total potential are thereby ignored. The pressure at every level is assumed to be identical to the distance to the groundwater surface. This

means that one assumes that the flow at every level is horizontal. This simplification is called *Dupuit's assumption*, after its originator.

Water flow (Q) through a cross section with area A, if the x-axis is horizontal, is now given by

$$Q = - K \cdot A \cdot dh/dx$$

h = elevation of the groundwater surface (m)
dh/dx = the slope of the groundwater surface (m/m, i.e. unit less)
The hydraulic conductivity is a measure of the soil's capacity to transmit water expressed per unit area of cross-section. Often it is practical to define the capacity to transmit water per unit width of the cross-section. This is called the transmissivity, T (m^2/s).

$$Q = -K \cdot A\ dh/dx = -K \cdot m \cdot b \cdot dh/dx = -T \cdot b \cdot dh/dx$$

m = the thickness of the soil layer in question
b = the width of the soil layer in question

The transmissivity thus expresses the capacity of an entire soil layer to transmit water. It is related to the conductivity by the expression $T = K \cdot m$. As we shall see somewhat further on, K often changes a great deal with depth. In such cases T is the sum of the contributions by every soil layer to the hydraulic conductivity,

$$T = \int_{z1}^{z2} K dz$$

If we know the transmissivity, we can determine the total flow of groundwater per unit width directly from the slope of the groundwater surface.

Determining the direction of groundwater flow

As we stated earlier, the magnitude of the hydraulic conductivity is a central question when calculating groundwater flow. Due to the variation of this quantity by many orders of magnitude between different soil types, and even in the same soil type due to differences in the structure of the soil, it is difficult to estimate a reliable value for what the conductivity should be in a natural environment. It is therefore difficult to make precise calculations of the amount of flow. The direction of the groundwater flow, however, is simple to determine since the flow follows the slope of the groundwater surface. From at least three observation points of the groundwater table at a reasonable distance from each other it is possible to determine the slope of the groundwater surface, and thus the direction of groundwater flow.

Figure 18 shows the results of a detailed mapping of the groundwater surface in a small area of forested, glacial till soil. The groundwater level has been determined along a grid network with the help of perforated plastic tubes that have been driven into the soil. The groundwater's direction of flow is perpendicular to the contour lines of groundwater level. A comparison with the topography of the ground surface shows that the groundwater table essentially follows the slope of the ground (except for in the streams themselves that are incised into the soil). The groundwater thus flow perpendicular to the contour lines of the ground surface. This is the usual situation in glacial till terrain, even in the areas where the groundwater level does not lie as near to the soil surface as in this case. The deeper the groundwater level lies, the more local divergences there will be from the topography of the ground surface, but the principal direction of flow will

coincide with the principal direction of slope. If one does not know the direction of the groundwater flow, the best guess is often in the direction of the hillslope's fall line. (In other geological formations, for example in a glaciofluvial esker, this does not apply.) Another observation that can be made in Fig. 18 is the relationship between the streams and the groundwater level. The groundwater flows towards the streams, the surfaces of which can be thought of as a part of the groundwater surface. Even this is typical for streams in our climate zone, and not just in till soils.

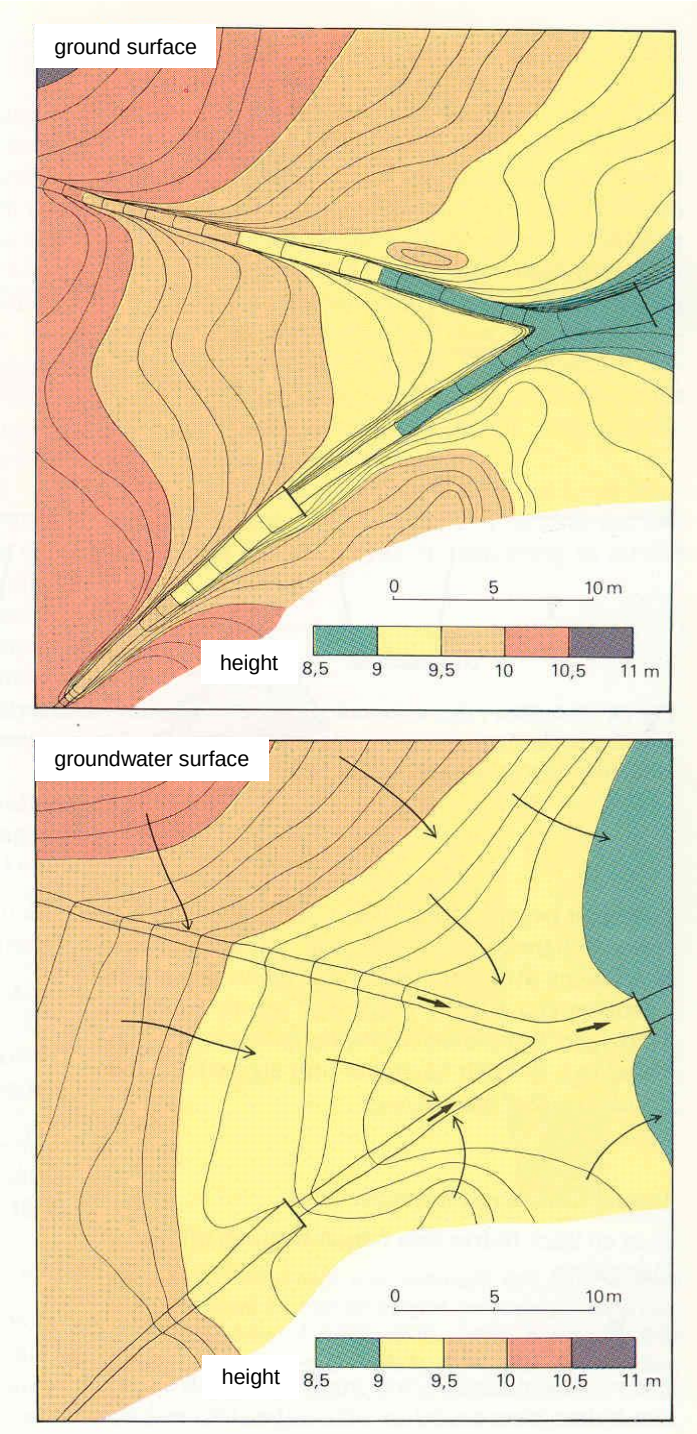


Fig. 18 Topographic contours (above) and the groundwater surface (below) near two streams that join. The Kloten area, Västmanland, September 1983.

The groundwater level is very superficial in this till hillslope. The flow of groundwater, which occurs perpendicular to the groundwater contours, follows, to a large extent, the slope of the soil surface.

Normally groundwater flows towards streams. In one of the stream channels and below the confluence, there is a streamflow gauge (the straight line across the stream in the figure) that dams the water level. This increase in the level of the stream leads to a groundwater flow from the stream into the adjacent slope in the vicinity of the gauge. Calculations show that this leakage is of no significance for the determination of streamflow.

The groundwater surface in Fig. 18 was defined with the help of observation tubes constructed so that water can enter along the entire length of the tube

that is below the groundwater level. Two tubes driven to different depths at the same point will thus have the same water level. This is the groundwater level, defined as the level where the pressure in the soil water is the same as that of the atmosphere.

If one instead has tubes that are perforated only at a specific level, one can get different water levels in tubes at the same location, but with water intakes at different depths. The water level in each of these tubes shows the total potential of the groundwater at the level of the intake. In the middle of a hillslope, the total potential in the groundwater is often the same at all levels, and the flow is essentially parallel to the groundwater surface, so Dupuit's assumption applies well. Near the ends of the hillslope, though, the assumption is less appropriate. At the upper end of the hillslope, near the groundwater divide, there is a downward component of flow. The total potential decreases with depth. At the foot of the hillslope, the situation is reversed, and the total potential increases with depths, so there is an upward component of flow. An observation tube with an intake one or two meters under the bottom of a stream often has a water level that is several cm higher than the surface of the stream itself, which shows that groundwater is flowing from the bottom of the stream into the stream.

The velocity of soil water and groundwater flow

There are at least three different sorts of "velocity" that are used when characterizing the flow of soil water and groundwater. Darcy velocity, particle velocity and the pressure wave propagation velocity. It can sometimes be difficult to keep from confusing these different "velocities".

Darcy velocity

Darcy's Law can be written

$$Q/A = -K \cdot d\phi/dx$$

The term Q/A refers to flow, i.e., the amount of water passing a cross-sectional area of soil per unit time, and it has the dimension velocity (units of m/s). This is also often called the velocity and is represented by v, which is the customary symbol for velocity. Sometimes this is called the Darcy-velocity (v_d). It is important to note that this "velocity" does not refer to the velocity of a water particle (v_p), i.e. the velocity at which a perfect tracer would move.

Water particle velocity

A stream is one example of where the entire cross sectional area contributes to the flow of water. There the following expression holds

$$v = Q/A$$

where v = the average water particle velocity. In the soil, only a portion of the cross-section contributes to flow, which we can term the A_e (= effective area). The water particle velocity v_p can be described

$$v_p = Q/A_e$$

If the presence of water bound by adsorption is ignored, then

$$A_e = \theta \cdot A$$

where θ = water content (expressed as a ratio, not a percentage). We get the following relationship

$$v_p = Q/(A \cdot \theta) = v_d/\theta$$

Since the water content is less than one, the particle velocity is greater than the Darcy velocity. In the case of groundwater flow, the water content is equivalent to the porosity, p and
$v_p = v_d / p$

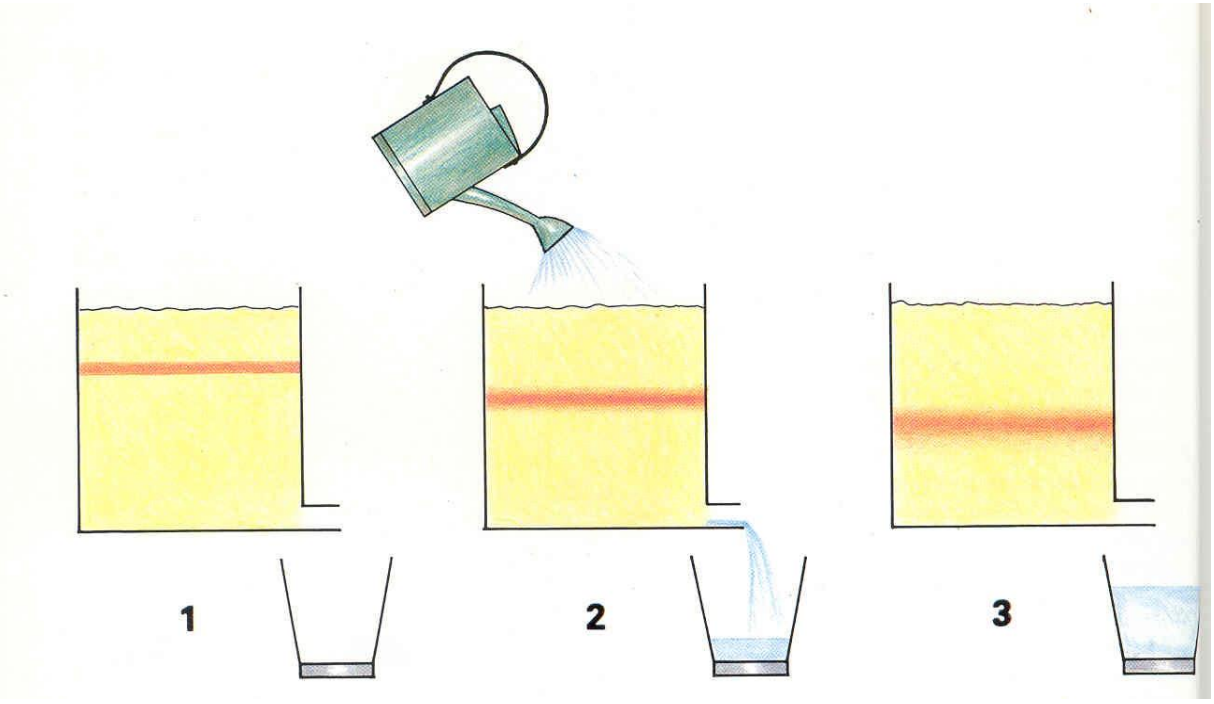


Fig. 19 Piston flow in the unsaturated zone. Disregarding a certain spreading of the dye, the labelled layer of water maintains its identity while moving through the soil. The "groundwater" that is created in the bucket during the experiment consists of other water particles than those which infiltrated at the surface.

This relationship applies to groundwater flow in a sorted, reasonable coarse-grained soil (e.g. sand) without structure. In this case the water bound by adsorption that does not participate in the flow is a small portion of the total porosity. Since the soil does not have a structure, there is no interconnected system of large pores. All water particles therefore participate in the flow on equal terms. An individual water particle's velocity does indeed depend on the size of the different pores. But seen on a length scale that is larger than the size of the pores, the water particles attain the same average velocity in the course of their journey through larger and smaller pores, $Q/(A \cdot p)$.

In a fine grained soil, the water bound by adsorption cannot be ignored when calculating velocities. In order to account for this, the porosity in the expression for particle velocity is often replaced by the *effective porosity* or *kinematic porosity* defined as v_p / v_d . A comment on the terminology is needed here. In some literature, the term effective porosity refers to the drainable volume fraction of the soil, i.e., the fraction that is "effective" for groundwater extraction, which equals porosity minus field capacity. That porosity, called specific yield (see page 60), should not be used when calculating particle velocity of groundwater. Below the groundwater level, as is commented on page 20, there are no capillary forces. All water, except for that which is bound by adsorption, participates in the flow.

In a soil with structure, there can be integrated systems of larger pores, such as cracks, worm holes and root channels. Under saturated conditions, these can result in a rapid transport of water particles. For some of the particles, the velocity will therefore be greater than $Q/(A \cdot p)$ and for others, those that don't get into the larger pores, the velocity will be less. A tracer will move at very different velocities, depending on whether it is moving in the large or the small pores.

Pressure wave propagation velocity

The two preceding equations described the relationship between the Darcy velocity and the particle velocity. The third velocity, the pressure wave propagation velocity, cannot be related to the others in any simple way. It is always much greater than the particle velocity in unsaturated, as well as saturated flow. Thus a groundwater level at 2 m depth can begin to rise a few hours after it has begun to rain, even though it may take months before the infiltrating water particles reach the groundwater. An increase in the groundwater level and an increase in its slope increase the outflow of groundwater from a discharge area, but these will be different water particles than those which caused the groundwater level to rise. The pressure propagates rapidly while water particles move slowly.

One can perhaps make a comparison to a long garden hose that is attached to a water faucet. It takes a goodly while to fill an empty hose before it begins to spray water out the far end, and this delay is determined by the particle velocity. But when one opens the faucet on a hose that is already filled, the hose begins to spray water immediately. Here it is just the pressure wave propagation velocity that determines how fast the discharge responds.

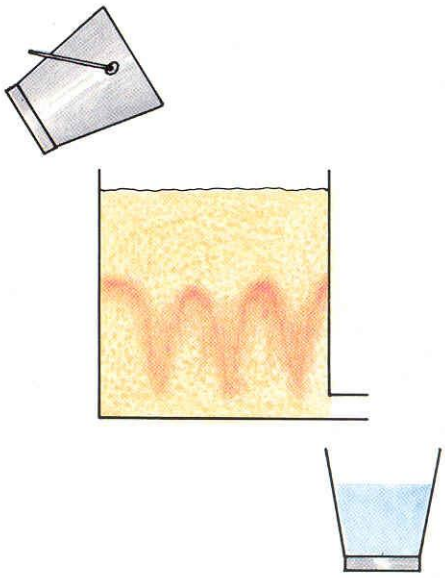


Fig. 20 Here the soil is not homogeneous and there has been a large amount of infiltration. Under the almost saturated conditions that were created in the soil, systems of larger pores can transmit water and the percolating water particles move at different velocities. This situation can also arise through local breakthroughs in a coarse horizontal layer. This is not piston-flow.

The Sausage-filling Principle

Water's chemical transformation in the unsaturated zone depends to a large extent on the residence time of water there, which in turn depends on the velocity of the water. The equation for water particle velocity (page 38) is based on the participation of all water particles in the flow on equal terms, i.e. they all have the same average velocity. No particles slip by the others by having found conditions particularly conducive to rapid flow. This type of flow is usually called *piston flow*. The whole water volume moves through the soil without changing the internal ordering of water particles within that volume. For vertical flow in the soil water zone, that

means that the water particles at a particular level all have the same age. In Fig. 19, piston flow in the unsaturated zone is visualized by a coloured water layer, which moves downward uniformly when there is a moderate input of water at the ground surface. If the input of water is large, as in Fig. 20, the water particles at one level have different average velocities, and one no longer has piston flow. The cause for this can be in-homogeneities in the soil, for example a vertical system of larger pores that fill with water when the infiltration rate is high, or horizontal lenses of coarser material, through which local breakthroughs can occur (cf. page 33). There are different opinions among experts as to how much of the water participates in the flow under different conditions.

The soil frost growth and disappearance

In a frozen soil layer the water is totally or partially frozen. The soil frost grows with a velocity that depend on the net soil heat flux at the freezing front, i.e., on the difference between the heat loss by conduction through the frozen soil to the soil surface and the heat gain by conduction through the unfrozen soil from the warmer layers below. The growth is favoured by large energy loss at the soil surface making the soil surface cold. Low air temperature and negative net radiation are the meteorological conditions. Snow has a low conductivity for heat and decreases the energy losses from the soil surface and therefore the soil freezing. The consequence of a few dm snowfalls on frozen bare soil is that the growth of the soil frost stops, and in some cases the frost starts to melt.

The soil frost depth depends on the heat conducting properties of the soil matrix, and also on the soil water content. The larger soil water content the more heat has to be conducted away for a given soil layer to freeze. In soils with high water content the soil frost front grows slowly downward. The different frost depths in different soils demonstrate the connection between water content and soil frost depth. The deepest soil frost is found in the coarsest soils because their low water holding capacity result in low water contents at the tie of freezing.

The melt of the soil frost is mainly by heat flow from above when the snow cover has disappeared and the sun can heat the soil surface up. When the soil surface has been heated up the soil frost is also melting by heat conduction from below. At this point in time the temperature gradient in the frozen layer decreases and consequently also the heat conduction from the frost front.

The soil frost depth in an area can vary a lot depending on differences in the energy balance of the surface, the soil type and water content (see Fig. 21). The radiation loss from the soil surface is smaller in a forest than in the open, because of heat radiation from the forest canopy. (The difference between forest and open ground can be compared with the difference in ice formation below bridges and on open water. The heat radiation from the bridge reduces the heat loss from the water and the ice cover is formed later and the ice for skating is weaker.) Below dense forest stands the snow depth is small, which may over rule the effect from the heat radiation from the canopy to the soil surface and consequently give large frost formation there. The air temperature close to the soil surface can also vary a lot from place to place, for example will cold air be collected in low-lying parts of the terrain (the silt soil clear-cut in Fig. 21). The frost depth in the soil also varies a lot between years because of differences in weather and snow cover. During the winter after the one shown in Fig. 21 both sites on till soil were free from soil frost and the frost depth on the silt soil clear-cut was maximum 40 cm.

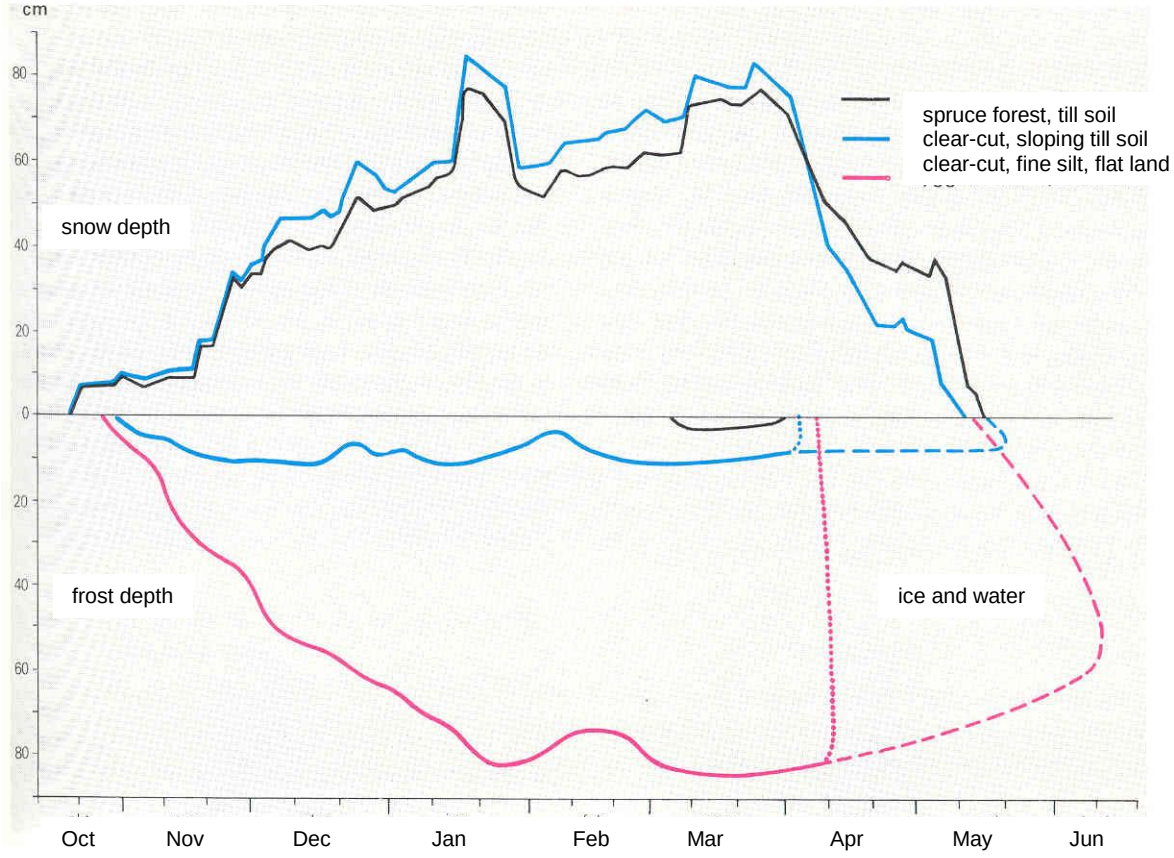


Fig. 21 Snow depth and frost depth at different places in the Svartberget Research Forest, Vindeln, County of Västerbotten, during the winter of 1982/83.

The snow depth is roughly the same at the measurement sites in the forest and on the clear-cut, but with a slower accumulation in the forest. The melting is considerably slower in the forest.

The depth of soil frost on the open area is much larger than in the forest, where the soil remains essentially unfrozen during the entire winter. The difference in soil frost depth between the two clear-cuts results partly from the fact that the topographic situation of the fine-silt clear-cut has a tendency to gather cold air, and partly from the low water content in that soil created by the large depth to the water table.

The depth of the soil frost has been measured indirectly, with a pressure sensor – a method that is based on the drastic decrease in soil water pressure (i.e. increased suction) when the soil freezes. The input of meltwater starting in the beginning of April that increases the soil water pressure in the entire frozen layer makes this measurement method unusable. At that point there is ice as well as water with a relatively high pressure in the soil. The melting of frost during this period has been estimated with experience from other measurements (dashed lines).

The published measurements of soil frost in Sweden were almost exceptionally performed on open ground. The appearance of soil frost in forested till soils is relatively unknown. The frost depth is probably small and often discontinuous over an area.

Water movement during soil frost formation

All water in the soil will not freeze at 0°C. The freezing point of water is lowered when the soil water pressure decreases. When heat is conducted away from soil at 0°C the water in the large pores freezes first. In the same way as when the soil is drying the pressure of the remaining water decreases, and therefore its freezing point. The more fine textured a soil is the more unfrozen

water can it holds at a given temperature below 0°C. A clay soil that was originally saturated with water can contain 5% unfrozen water at –5°C, while a corresponding sand perhaps only 0.5%.

The lowering of the water pressure that the frost formation causes, leads to a pressure gradient, and therefore a flow from the underlying soil water to the frost front. The soil frost sucks water from the surroundings that freezes if the heat conduction away from the frost front is effective enough. The effect of this redistribution of water in the soil profile depends on the unsaturated hydraulic conductivity of the soil and on its water holding properties. In clay soils so called ice-striped soil frost appears. The water in single pores starts to freeze and suck water from the surrounding unfrozen soil. Horizontal ice lenses form in the otherwise unfrozen soil. The ice lenses are perhaps a few mm thick and some cm long, but also dm-thick lenses may form. In a coarser soil the water freezes in the whole pore system and a homogenous (but normally not massive) soil frost develops. In the case of medium sized pores in the soil its unsaturated hydraulic conductivity is favourable and a substantial water flow can take place from the underlying soil- and groundwater towards the frost front. The ice content of a frozen soil may therefore be higher than its water content before frost formation.

An effect of this upward water flow and freezing can be seen on some roads in springtime. When water freezes its volume increases. If a large flow takes place from below and that water freezes at the frost front the volume of the total soil increases considerably and the soil surface is heaved. The heaving during the winter is successive and even and is seldom observed. The thawing of the frozen soil in the spring is on the other hand fast. The soil sinks unevenly and pits are formed in the road surface. The upward directed water flow, and therefore the frost heaving is extraordinary large in fine silt with short distance to the groundwater surface. Roads on such soils are notorious for frost heaving.

Evaporation and transpiration

Water is lost from soil and water surfaces and vegetation to the atmosphere by *evapotranspiration*. Evapotranspiration is the sum of two partial flows. *Evaporation* takes place from wet surfaces (for example wet leaves, snow, lakes, and watercourses) and soils, while *transpiration* takes place via the stomata of the plant leaves. Normally the word evaporation is used synonymously with evapotranspiration.

The evaporation as well as the transpiration is driven by meteorological conditions. The atmosphere "demands" water from the moist surfaces it is in contact with. When the plants open their stomata to collect carbon dioxide for its growth moist surfaces are exposed to the atmosphere and are forced to release water. The water flow through the pants is much larger than what is needed for plant growth. Perhaps is the large transpiration an important way for the leaves to avoid over-heating. The plants have some ability to regulate the stomata openings and by that the water flow. If the water flow to the atmosphere for a long time is larger than what the plants can take up from the soil the plants will wilt.

For evapotranspiration to take place energy to transform the water from liquid phase to gas phase, an unsaturated atmosphere, and a transport mechanism to bring the water vapour into the free atmosphere is needed. The energy is mainly taken from the net radiation, i.e., the difference between incoming and outgoing radiation. In some situations even the energy that is stored in warm vegetation, soil, water and air is used. The energy storage delays evapotranspiration during the year and during the day compared with the net radiation. In the morning, for example a large part of the net radiation is used to warm up the soil, which only marginally contributes to the

evapotranspiration. In the afternoon the stored energy is an addition to the net radiation and the evapotranspiration is relatively large.

The flow of water vapour into the free air is driven by a gradient in vapour content. The air just above the evapotranspiring surface has higher moisture content than the air at higher levels. The resistance of the air restricts the transport of water vapour. This resistance that is called *aerodynamic resistance* decreases with increasing wind velocity because the turbulence will increase and the vortexes lift air with high moisture content and replace it with air that have lower moisture content. The higher the vegetation is the more turbulence it creates. The aerodynamic resistance over forest is therefore smaller than over grass.

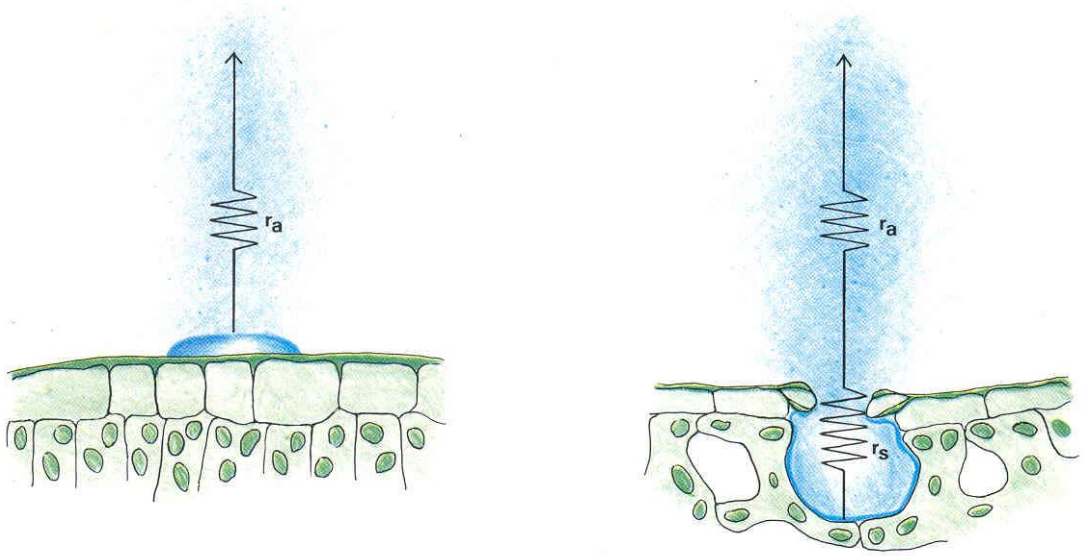


Fig. 22 In analogy with Ohm's law the flow of water vapour can be written as, flow = concentration difference in water vapour divided by resistance (cf. I = U / R for electrical current.) When intercepted water is evaporated the water vapour has only to overcome the resistance between the surfaces of the leaves and the atmosphere, the so-called aerodynamic resistance (r_a). At transpiration the resistance in the stomata is added. The total effect of all stomata coupled in parallel is called surface resistance (r_s).

Water that evaporates from wet surfaces only has to overcome the aerodynamic resistance, while water that is lost by the plant transpiration in addition has to overcome a resistance in the stomata. For dry forest this so-called *surface resistance* is 10 – 30 times larger than the aerodynamic resistance. This means that the evaporation from wet forest is many times larger than the transpiration from dry forest under similar climatic conditions. From a grass covered surface, where the aerodynamic resistance is about equal to the surface resistance, the evaporation from wet grass is about two times as large as transpiration from dry grass (Fig. 23).

The evapotranspiration is said to be *potential* if the water supply is good and the flow is only restricted by the weather. Potential evapotranspiration can be calculated from net radiation, air temperature, relative humidity and wind speed. A good correlation is often found between calculated potential evapotranspiration and the evaporation from small open water filled containers. These values are however a bad measure of real evapotranspiration because plants have an ability to reduce their water loss, which the container cannot.

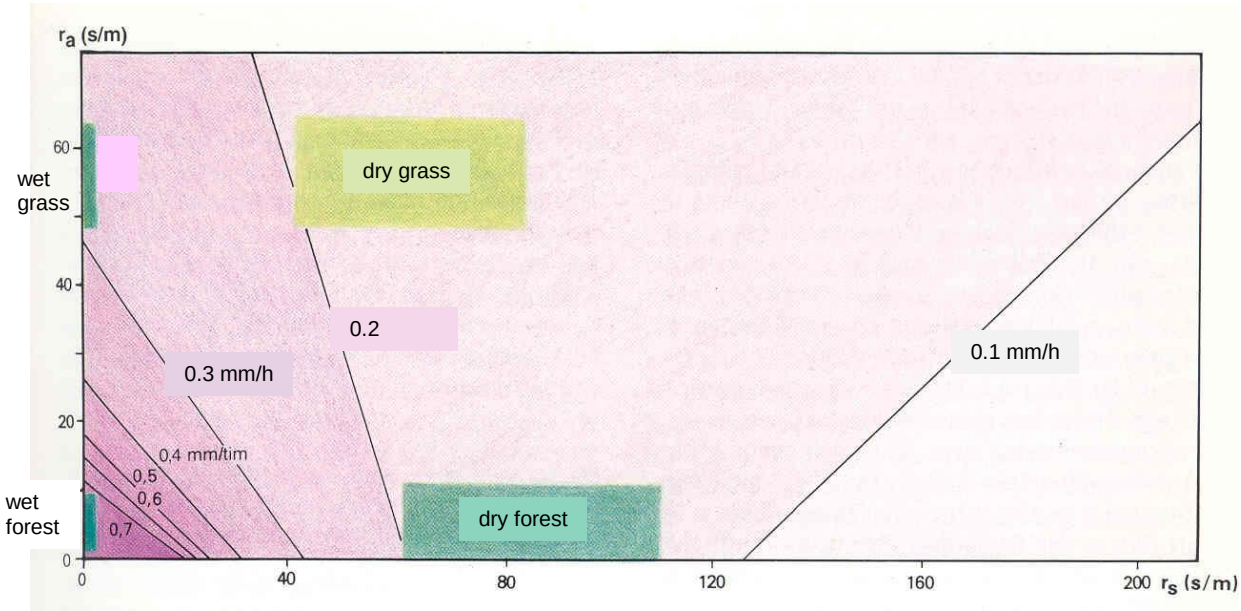


Fig. 23 The diagram shows how the evaporation intensity depends on aerodynamic resistance (r_a) and surface resistance (r_s) during a summer day.

At a given weather the transpiration from a dry forest is about equal to that from dry grass. Evaporation from wet forest is on the other hand much larger than from wet grass. The reason for the high intensity of the evaporation from the wet forest is that the forest creates intense vortexes that effectively transport the water vapour away. The aerodynamic resistance (r_a) is small. In the diagram the same summer weather was assumed for wet and dry vegetation. But directly after a rain the weather is often less favourable for evaporation (the vegetation is cold, the air moist and the radiation small). The effective transport of water vapour from the evaporation surface still makes evaporation from wet forests rather large, maybe as large as transpiration from a dry forest a nice summer day.

Plants can actively reduce their water loss

Plants can reduce transpiration by closing their stomata. The resistance is large when the relative humidity is low and increases with solar radiation.

The resistance is also influenced by the water availability in the root zone, and increases with decreasing water content because the water then is held at lower pressures and the flow towards the roots decreases. The sensitivity to decreased water availability is different for different plant species, but the pressure potential where many plants wilt is fairly constant, -150 m (the wilting point). It has also been shown that the pressure potential when the stomata start to close depends on the potential evapotranspiration. At high potential evapotranspiration the reduction starts already when the available soil water is high, but when it is low the reduction starts at lower water availabilities (Fig. 24).

During and after rain when needles and leaves are wet, water for evapotranspiration is almost exclusively taken from the wet surfaces because the total resistance for evaporation is less than for transpiration.

Trees first use water stored in leaves, trunks and stems when they transpire, but when the storage in the plants decreases the potential gradient between the soil and the plant increases and the water uptake increases. The water uptake therefore lags behind transpiration.

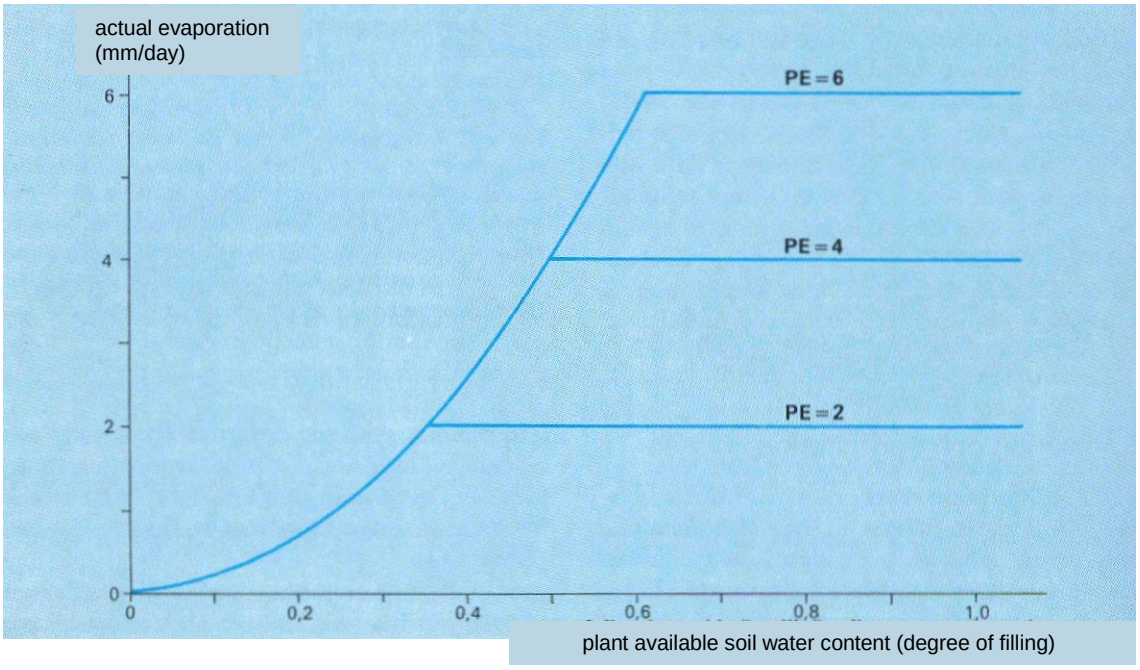


Fig. 24 Actual transpiration from dry vegetation under different potential evaporation, PE, as a function of soil water content. In the diagram the wilting point is represented by the available storage 0, and the field capacity by the available storage 1.

The transpiration depends on the weather, i.e., the potential evaporation, and on the soil water content. When the potential evaporation is small enough soil water can flow to the plant roots also when the water content is low. When the potential evaporation is high the soil water content is limiting.

The water uptake from the root zone depends on water availability at different horizons and on root distribution. At equal pressure potentials on different horizons the water uptake from each level is often quoted to be proportional to the root density in each layer. When the pressure potential is different between horizons uptake from where water is most easily extracted dominates. The root activity decreases strongly with soil temperature, and the water uptake ability is less good in a cold soil as compared with a warmer soil with the same pressure potential. One effect of this is that coniferous trees may desiccate during spring when the atmosphere starts to demand much water while the water rich soil is still too cold.

4. Water in recharge areas

In Chapter 2 the catchment was used as the unit. We saw how much water that was input to the area from precipitation and discussed where it ended up. During a certain time period some water was accumulated, some was lost by evaporation, and some was discharged as runoff in streams. Flows and storages were explained per unit area of the catchment (cf. Fig. 7). Differences between catchment sub-areas were not considered. Nor did we try to explain the mechanisms that govern water storage and flow during the water's journey from rain to stream.

In this and the coming chapters we emphasise instead the differences within a catchment. Based on the physical principles presented in Chapter 3 we will follow the water from recharge- to discharge areas.

The most common landscape type in Sweden is coniferous forested till terrain on solid rock. We use this landscape type to discuss the water flow through a catchment. Examples from other landscape types are used when we want to discuss specific processes and when we, as in the last chapter try to apply the views presented here on certain practical problems.

Groundwater is formed in the recharge area

Recharge areas are the parts of a catchment where there is a recharge of groundwater, that is, where groundwater is formed. In these areas water percolates from the soil water zone to the groundwater zone.

From the groundwater flow direction recharge areas can be defined as "areas where the upper groundwater has a flow component directed into the groundwater zone". For this definition to give groundwater recharge also for a sloping groundwater surface the flow direction is compared with the groundwater surface, not with the horizontal plane. This means that the flow resultant at the surface must incline more than the groundwater surface. In practice the slope of the groundwater surface is often small compared with all possible flow directions and a sufficient definition is "areas where the upper groundwater flow has a downward component".

The existence of groundwater recharge can be proven by measurements of the total potential at different depths in the groundwater zone (page 40). A decreasing total potential with depth indicates recharge. The groundwater flow in a hillslope is often more or less in parallel with the soil surface. The downward directed potential gradient is small in such case and may not be detectable.

In Swedish till slopes the groundwater surface, as earlier stated, more or less closely follows the soil surface. The surface of streams is the lowest part of the groundwater surface. According to the principles for groundwater flow this means a groundwater flow from high to low positions in the landscape with discharge around and in the streams. We know, therefore that there is a groundwater flow through a catchment to the streams, but we don't know the size and importance of this flow for runoff generation. There is no practical possibility to measure the total groundwater discharge into the streams of even a small catchment. Calculations with Darcy's Law may indicate, but not show the importance of the groundwater flow, since the hydraulic conductivity of the till soil and the bedrock is not enough well known. The few measurements of conductivity that do exist show large variation with depth and between different points of measurement which means that calculations of flows and their variation with time will be uncertain.

Because we can neither measure, nor with enough accuracy calculate the groundwater flow in a catchment we are referred to indirect methods to judge the validity of the hypothesis for runoff formation that was sketched in the introduction.

Water input

In Sweden rainfall normally has low intensity. A "rainy" day, when you would get properly wet without a raincoat, the rainfall intensity may be 2 – 4 mm per hour. The showers during summer afternoons often have higher intensities, maybe 10 mm per hour, and the intensities of heavy thunderstorms maybe 30 – 50 mm per hour. High intensities are rare and seldom last more than a couple of minutes to less than one hour. From Fig. 25 we can see that the intensity exceeds 40 mm per hour during 10 minutes only once per year, statistically. The same intensity is exceeded during 35 minutes only once per decade. The data is from observations in Gothenburg, West Sweden, but the conditions are similar in other parts of the country.

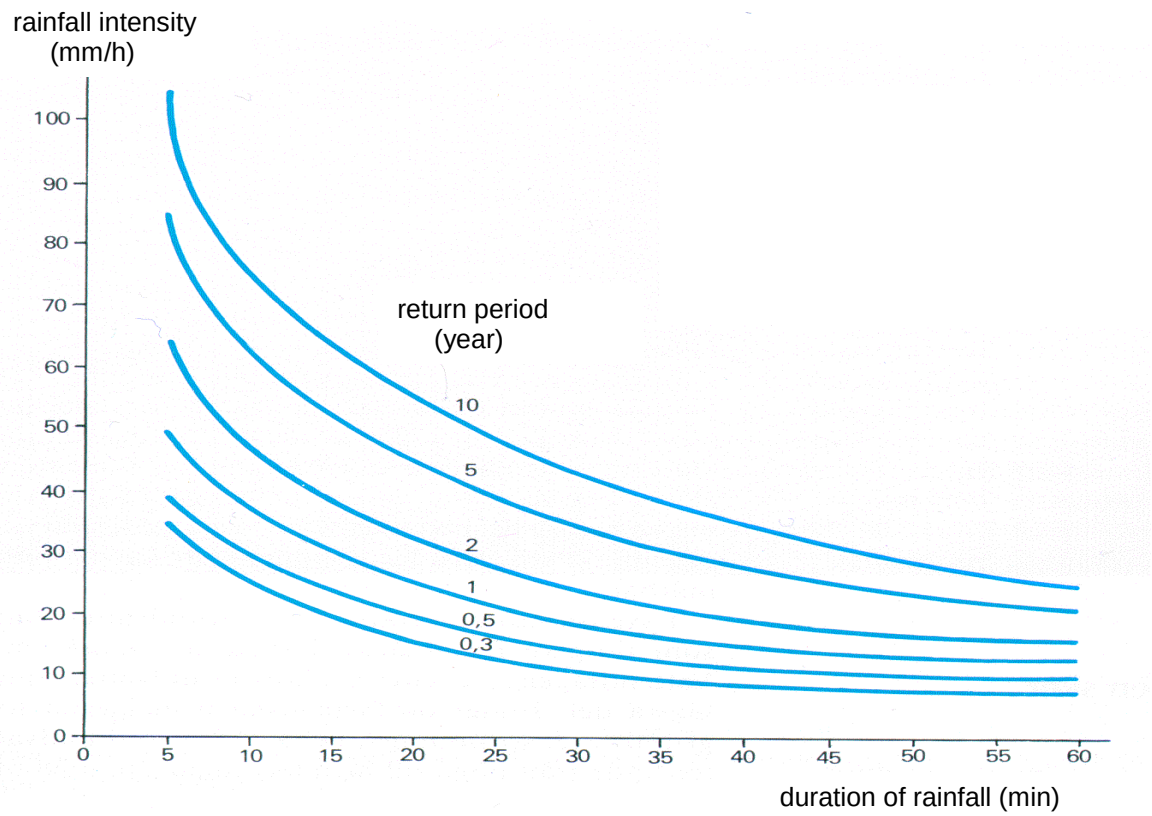


Fig. 25 The diagram shows the probability for rain of a certain intensity and duration (observations during a 10-year period in Gothenburg). Rains with high intensity have long return period, i.e., they are sparse and when they appear they have short duration. For example it happens statistically two times per year that the intensity exceeds 30 mm per hour during a 10-minute period (the return time is 0.5 years).

Consequently, Fig. 25 shows the intensity of the heaviest rains. The diagram is an important basis for the calculation of maximum runoff from impermeable surfaces for example when the dimensions of urban storm water pipes are designed. From the figure we cannot read anything about the intensity of the frequent storms and these are our focus when we discuss runoff generation. The share of rain with different intensities is shown in Fig. 26. As is seen most rain falls with low intensity. For example 50% of the rain has less intensity than 2 mm per hour and 80% with less intensity than 6 mm per hour. On the other hand we can also see that the rare

heavy rains contribute a relatively large part of the total rain. Thus about 6% of the total amount fell with intensity larger than 25 mm per hour. Even if these storms are infrequent and last short time, they bring large amounts of water. Also this diagram, which shows data from Östersund, Central Sweden, is in principle valid for other parts of the country.

Earlier we showed that snowmelt contributed a large proportion of the total water input and that this input is often more effective for the runoff generation than the same amount of rain (page 12). In comparison with rain snowmelt has low intensity.

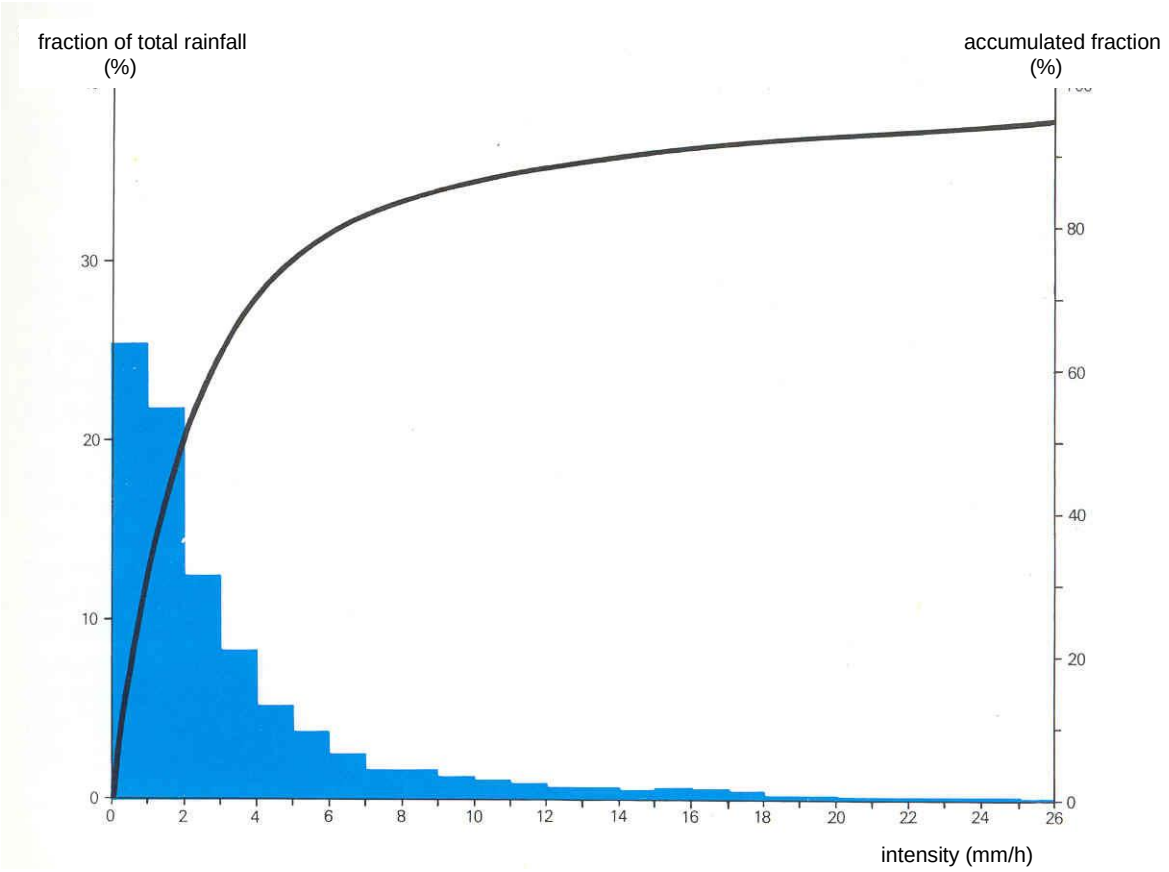


Fig. 26 Rainfall intensity distribution for a nine-year measurement period in Östersund, Northwest Sweden. The staircase curve shows how large part of the rain that has fallen with a given intensity (left scale). The smooth curve shows how large part that has fallen with intensities lower that a given value (right scale). Most of the rain falls with low intensity and for example almost half of the rain had intensities less than 2 mm per hour.

Snowmelt is determined by the energy input to the snow pack. This is largely the sum of solar radiation, heat conduction from the air, and if the air has high humidity condensation of water vapour (energy is released when water is condensed). Favourable condition for snowmelt is therefore sunny weather with warm and moist air and strong winds. The wind and the turbulence it causes transport heat and water vapour from the air mass to the snow surface. Often the air is relatively dry and evaporation takes place from the snow pack instead of condensation. The water loss that is due to evaporation is small, but the heat it consumes constitutes a relatively large part of the available heat and the melting rate decreases. Some energy may be added to the snow pack from rain, but that is normally negligible. A rain of 20 mm at an air temperature of +4°C will for example melt only 1 mm water. A rain storm may on the other hand have a large effect on the runoff, because the soil is already wet and the evaporation is small.

Experience shows that the air temperature is a good index for energy input to the local snow pack. For each degree above 0°C about 2 – 5 mm snow melts per day. This figure, the so-called degree-day-factor, is larger for open areas than for forestland. The larger solar radiation and the higher wind speed on open areas give a faster snowmelt there.

The snowmelt intensity is usually less than 10 mm per day. During some warm days it may be as large as 15 mm per day and in rare cases up to 30 mm per day have been measured. The snowmelt has a characteristic daily variation with maximum intensity during the afternoon and often no melt during the night. If we assume that all melt occurs during the day, that is, during 12 hours, the melt intensities given above would be 0.8, 1.3 and 2.5 mm per hour, respectively.

Water entrance into the soil

Infiltration denotes the entrance of water into the soil. Infiltration is a flux and is given in for example mm per hour. The *infiltration capacity* of a soil is the maximum infiltration rate that a specific soil can have. (The term infiltration coefficient is sometimes used to denote the relative share of the precipitation that forms groundwater. Plants take up a large part of the water that infiltrates for their transpiration and therefore the groundwater recharge is far less than the infiltration. The term infiltration coefficient is therefore misleading as the infiltration denotes the water that penetrates the soil surface. "Coefficient of groundwater recharge" would be a better term.)

If the intensity of rain or snowmelt is larger than the infiltration capacity of the soil the surplus is accumulated on the soil surface and eventually overland flow is formed. It is useful to distinguish between two types of overland flow:

Hortonian overland flow is the surface runoff that is formed when the infiltration capacity is exceeded in areas where the groundwater level is below the soil surface, i.e., mainly in recharge areas.

Saturated overland flow is formed on areas where the groundwater surface is at the soil surface. No infiltration can occur on such areas and all rain or snowmelt will form overland flow together with the discharging groundwater. On such areas it is meaningless to use the term infiltration capacity since discharge takes place and not infiltration.

A prerequisite for all overland flow is that the top layer of the soil is saturated. Only then can free water exist on the soil surface. But in the case of Hortonian overland flow the saturated layer is thin and there is an unsaturated horizon below the soil surface and the groundwater surface. In contrast to saturated overland flow some of the water will infiltrate where Hortonian overland flow takes place (as long as the surface is not completely watertight). If nothing else is stated overland flow will be used for Hortonian overland flow in the following text.

Determination of infiltration capacity

The infiltration capacity of different soils has been determined in the field with mainly two different methods, with infiltrometer and with overland flow measurements. A ring infiltrometer is a stump of a pipe with a diameter of 20 cm or more that is driven into the soil surface some cm. Water is added to the pipe until a few cm of water stand on top of the soil in the pipe. The infiltration capacity is given by the rate of water addition that is needed to keep the water level at

the same position. The water flow below the cylinder will not be totally vertical because a pressure potential gradient is formed from the near saturated area below the pipe and out to the drier soil around it. The spread of water laterally leads to overestimations of the infiltration capacity. To minimize the error another pipe with considerably larger diameter is often placed around the inner pipe and the water level is kept equal as in the inner pipe. The lateral spread from the inner pipe will then be less. From Darcy's Law it is possible to correct the infiltration capacity with respect to lateral spread.

Another way to estimate infiltration capacity is to collect and measure runoff from natural surfaces with small well-defined areas, so called runoff plots. Surfaces of may be 1 –100 m^2 are contained within artificial water divides. Water is added either by natural precipitation or by artificial rain simulators. Few such experiments are reported from Sweden, but internationally it has been a common method.

Factors determining the magnitude of the infiltration capacity

Water content

The process of infiltration involves the interaction between gravity and suction to move water downward from the soil surface. As infiltration proceeds, the forcing pressure gradient decreases and the infiltration capacity declines. After some while, may be after some hours, the infiltration capacity reaches a constant rate equal or somewhat smaller than the saturated hydraulic conductivity. When the infiltration capacity for a certain soil is given it is usually this constant rate that is referred to.

A very dry, fine textured soil may have low infiltration capacity due to change in structure when wetted. In such soil the infiltration capacity is highest at intermediate water contents.

Saturated conductivity

The final infiltration capacity is thus dependent on the saturated hydraulic conductivity of the soil. This in turn is dependent on pore size distribution (texture) and aggregate formation (structure). The large structural pores do not normally contribute to the infiltration in our soils, because they are seldom water saturated at the pressure potential prevailing under natural infiltration. On the other hand they contribute to the infiltration capacity, that is, infiltration under unlimited water addition.

Vegetation, worms and other biological activity in the soil favour conductivity due to structure. Compaction of the soil due to passage of humans, animals and vehicles decreases the conductivity. In this respect agricultural- and forest machines play an important role. The surface pores can also be compacted by high intensity raindrops and splashes in more extreme climates.

Soil frost decreases the hydraulic conductivity of the soil because some part of the soil pore volume is filled with ice, which cannot take part in the water flow. The decrease depends to a large extent on the water content when the soil freezes. The larger it is the lower will be the hydraulic conductivity and therefore the infiltration capacity. On the other hand soil frost leads to a loosening of the soil structure and an increase of the infiltration capacity during the frost-free period. (Infiltration in frozen soil is further discussed later.)

Storage capacities and layering

The comments on infiltration capacity above mainly refer to non-layered soils. In a layered soil the infiltration capacity may decrease with time in a more complicated way. During a limited time the infiltration capacity depends on an interaction between conductivity and storage capacities in the different soil layers. The size of the final infiltration capacity depends to a large extent on the conductivity of the least permeable layer, but during natural conditions the duration of the infiltration may be insufficient for this layer to be limiting.

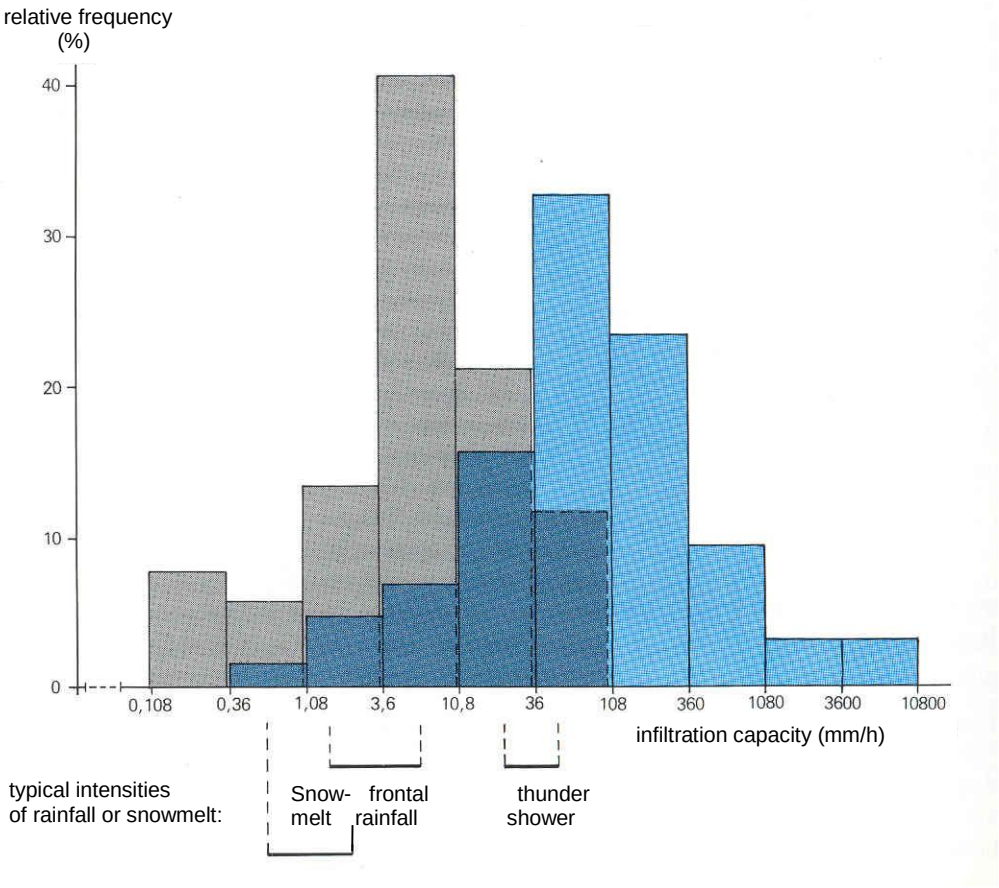


Fig. 27 Results from Swedish infiltration measurements in clay soils (black bars, 52 experiments) and till soils (blue bars, 64 experiments). In most of the till soils the infiltration capacity is large enough to infiltrate all rain and meltwater. In many clay soils the infiltration capacity is too low and Hortonian overland flow may occur.

Normally both conductivity and porosity decreases with soil depth. An intense rainstorm can infiltrate and temporarily be stored in the upper soil layers. The percolation to greater depth can then continue for a long time with a much lower intensity than the rain had. The deeper soil horizons will in such case not limit the infiltration capacity.

Vegetation plays an important role for this storage. Already in the interception, that is, the storage of water on trees, bushes, dwarf shrubs, herbs, grasses and mosses contributes to decrease the intensity of rain at the soil surface. In the soil an important storage takes place in dead organic material, both non-decomposed litter and decomposed humus.

The slope of the soil surface does not influence on the infiltration capacity. But in a slope the possibilities for storage on the soil surface and the upper soil layers is limited. Overland flow or lateral water flow in the upper soil layers will be favoured on the expense of infiltration. Terracing

sloping arable land increases the possibilities for storage on the soil surface, increasing the infiltration (but not the infiltration capacity) and decreases the overland flow.

Measurements of infiltration in Sweden

Infiltrometer measurements in Sweden are summarised in Fig. 27. Some were made on the natural soil surface, that is, the humus layer and others are made on the mineral soil surface after the humus layer has been scraped away. The latter generally gave lower infiltration capacities. The high infiltration capacities measured on the clay soils are probably influenced by structural cracks. A comparison between rain intensity and measured infiltration capacities shows that Hortonian overland flow should be rare on till soils but can occur on some clay soils.

In order to collect possible overland flow in five forested till slopes close to Uppsala one metre long collecting troughs were installed in the mineral soil about one cm below the humus layer in pits with tanks for water collection. During two seasons' snowmelt, about 15% of the snowmelt from their recharge areas was collected in two of the troughs. In the other troughs the collected amount was much lower (< 1%). During two summer- and autumn periods the troughs collected between 0% and 3% of the rain. A shorter observation period on a sloping meadow clay soil gave comparable result. Overland flow, as a water flow on the soil surface, was not observed during the numerous visits to the area during rain and snowmelt. In the case water was collected in the troughs it trickled out from the humus layer. In another trough on heavily cultivated grass covered sandy till overland flow was observed during one snowmelt period. In that case an ice sheet had formed on the soil surface due to frequent melt and freezing periods (see also page 54).

On page 34 it was stated that a pit diverts unsaturated water flow in the soil. A saturated flow, on the other hand, has a tendency to concentrate to a pit. In the study with the troughs the occurrence of saturated flow on the soil surface or in the surface soil layer was investigated. The disturbance that the pits induced should therefore lead to an over estimation of the flow. Probably this effect was small in this experiment.

Infiltration into frozen soil

We have already seen that water can move in frozen soil. Even if the soil was saturated when it froze it can be permeable for water. The conductivity in such cases is very small because the unfrozen water only exists in the smallest pores and around the soil particles. It decreases rapidly with the temperature because the water in successively smaller pores freezes when the temperature decreases. Saturated frozen soil may have a relatively high conductivity at 0ºC, but be practically impermeable at only some tenth of a degree below zero. An example is an experiment where the conductivity of a saturated fine textured soil was observed to decrease from 10^{-7} to 10^{-11} m/s (10 to 0.001 mm per day) when the temperature was lowered from 0.0ºC to –0.4ºC.

The situation is often different when water is infiltrating frozen soil because soils are normally not saturated when they freeze. If the water content was low enough when the soil froze there is a "spare" pore space for infiltration. Infiltrometer- and overland flow experiments have shown that the infiltration capacity of frozen soil is often enough for meltwater to infiltrate.

Overland flow experiments on silt soil in Alaska showed the influence of soil water content on infiltration capacity. Normally all meltwater infiltrated on the plot. In one spring out of three, overland flow was observed. In that case a heavy rain shower came just before the soil started to

freeze the autumn before. The depth of the soil frost was about 1.6 m during all winters (despite differences in water content).

It is thus not the frost depth that determines the infiltration capacity of frozen soil but the water content of the surface soil layers at freezing and also the redistribution of water during the winter season. The water content of the surface layers is probably highest at the lower parts of hillslopes because the depth to the groundwater is smallest there. If there is soil frost along the whole hillslope, all meltwater may infiltrate in the upper parts, but only a part of the water in the lower parts. The saturated overland flow would in such case get a contribution from Hortonian overland flow from areas just above the discharge areas.

During some winter seasons heavy ice formation takes place at the soil surface and in the surface soil layers. It is most frequent in southern Sweden when frost degrees occur after large water inputs by rainstorms or by snowmelt. If the remaining snow cover is thin a massive ice crust may form on the soil surface. The ice crust will be especially thick after repeated melt- and freezing periods. It is impermeable for water and additional water recharge generates overland flow. On arable land this type of ice formation may hurt winter crops.

Water storage in the upper soil horizons

The importance of water storage in the upper soil horizons is illustrated in Fig. 28. During the night totally 8 mm rain fell with "normal" low intensity. During the first hour all added water was stored in the moss layer. When the water content of the mosses had increased with 1.5 mm water it started to percolate into the litter and humus layer and their water content started to increase. The water storage in the moss layer continued to increase and at the largest rain intensity the water content of that layer was 3.5 mm higher than its initial value. When the rain intensity drastically decreased, the water content of the moss layer also decreased because the drainage to lower soil horizons exceeded input from above. Two hours after the start of the rain the recharge of the mineral soil started and after another two hours the effect of rainfall could be noticed at 20 cm depth in the mineral soil. The increase of the water content in the deeper soil layers continued long after the rain had stopped. Below 30 cm soil depth the increase in water content was small for the time period shown.

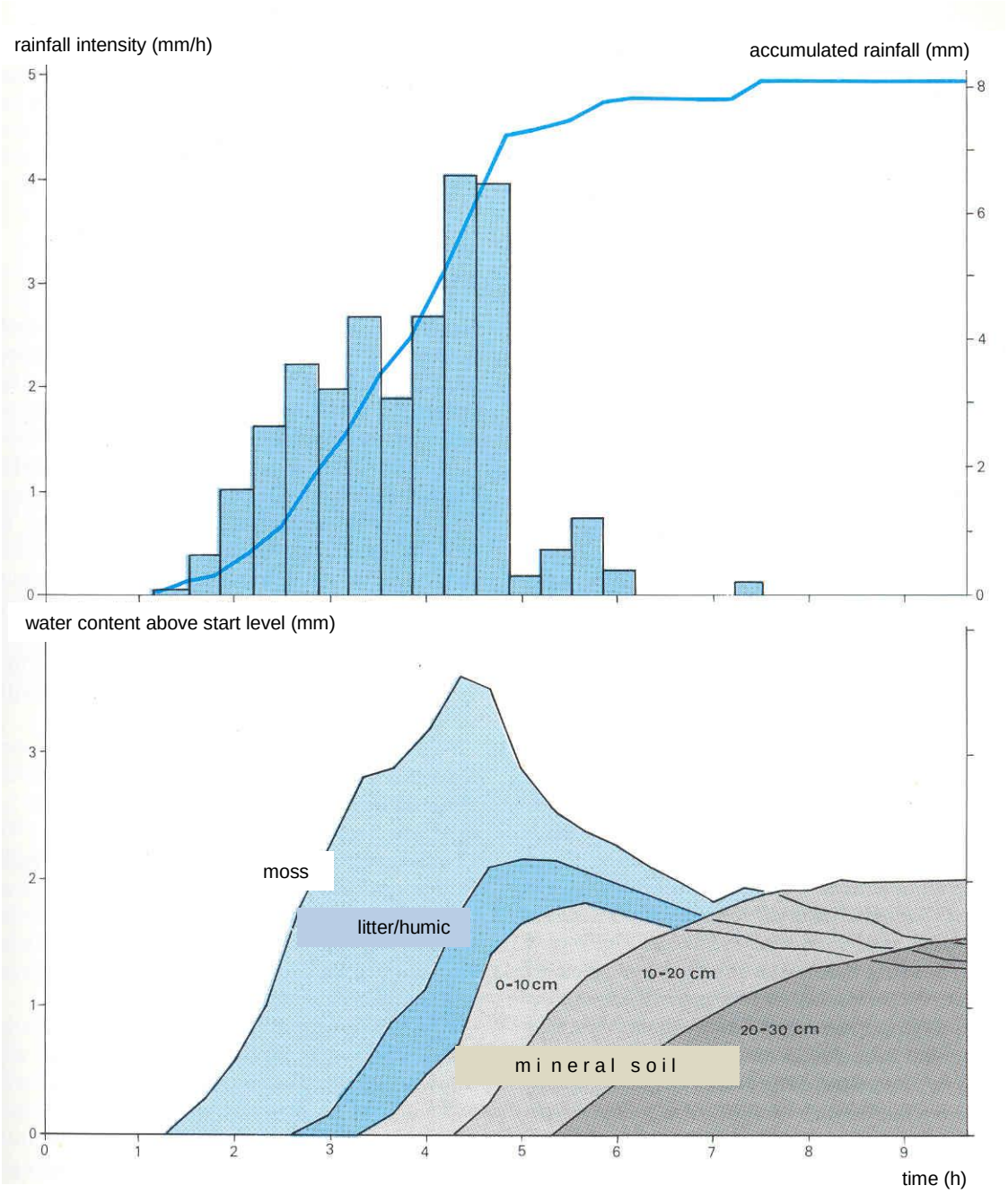


Fig. 28 A rainfall event's infiltration into a sand soil. (Jädraås, Gästrikland, Central Sweden, 28 September, 1975). The water content in the moss layer increases as soon as the rain starts. The water content increases with time deeper and deeper in the mineral soil. Due to the storage in the different layers the flow intensity decreases with depth. Denser layers that could limit the infiltration are not activated because of the short rainfall duration.

The depth to the groundwater is about 15 m on the site. Probably all water from this rain storm was stored in the root zone even after the measuring period and no groundwater recharge took place. On other places with shallow groundwater, but with all other conditions equal, some recharge of the groundwater would have taken place generating increased groundwater discharge.

The slope of the curves in Fig. 28 denotes the change in water content per time unit, that is, the storage in the layers (positive at recharge, negative at drainage). As long as the water content in the layers below is not changed the storage in a layer is equal to the input. When the water content of the layers below is increasing the storage is only a part of the input, which in this case

is larger than the slope of the curve. The input to a layer may be calculated if the storage in the layer above is subtracted from the input to that layer. The highest input intensity to the lowest layer was 0.7 mm per hour, which could be compared with the intensity of the rain and the input to the moss layer of 4 mm per hour. The maximum input intensity to the mineral soil was 3 mm per hour. The storage in horizons above has a damping effect on the percolation rate to the lower horizons. The demand for conductivity in a layer to cope with input from above is therefore decreasing with soil depth (if the rain is not very long lasting).

Observe that the curves in Fig. 28 show the increase in water content above its initial values for each layer. The water content in the mineral soil was about 25% at the beginning of the rain, which corresponds to 25 mm water in a 10 cm thick soil layer. The relative change in water content of the different soil layers was therefore small.

The time displacement between the different curves with depth shows the propagation velocity of the precipitation pulse. It is about 10 cm per hour in the mineral soil. Let us compare this velocity with the velocity of the water particles:

If we assume that all water particles took part in the flow the velocity of the water particles was

$v_p = Q/(A \cdot \theta)$

The flow (Q/A) at 10 cm depth was about 1 mm per hour. Given the water content $\theta = 25\%$, the particle velocity, v_p = 4 mm per hour. The three velocities discussed on page 38 were therefore

Flow (Darcy-velocity)	1 mm per hour
Particle velocity	4 mm per hour
Pressure propagation velocity	100 mm per hour

The further movement of water through the soil profile

The downward unsaturated flow of water in the soil profile is called *percolation*. The difference between the velocity of the water particles and the propagation velocity of the pressure is characteristic for percolation. In several studies the particle velocity of the percolating water has been estimated with natural and added tracers in the water. The investigations have been made on sand deposits with deep groundwater table. From relatively scarce observations particle mean velocities of about 0.1 – 0.2 m per month has been estimated. These are mean velocities from which, for example, the transport of a contaminant in the unsaturated zone may be estimated. At each single percolation event the velocities were probably larger (compare the velocities in the example above). The pressure propagation velocity in these thick unsaturated zones has been estimated to about 3 m per month.

These investigations, as well as that in Fig. 28, concern sandy soils with small water holding capacity. In till soils the water content is often larger and the particle velocity lower for a

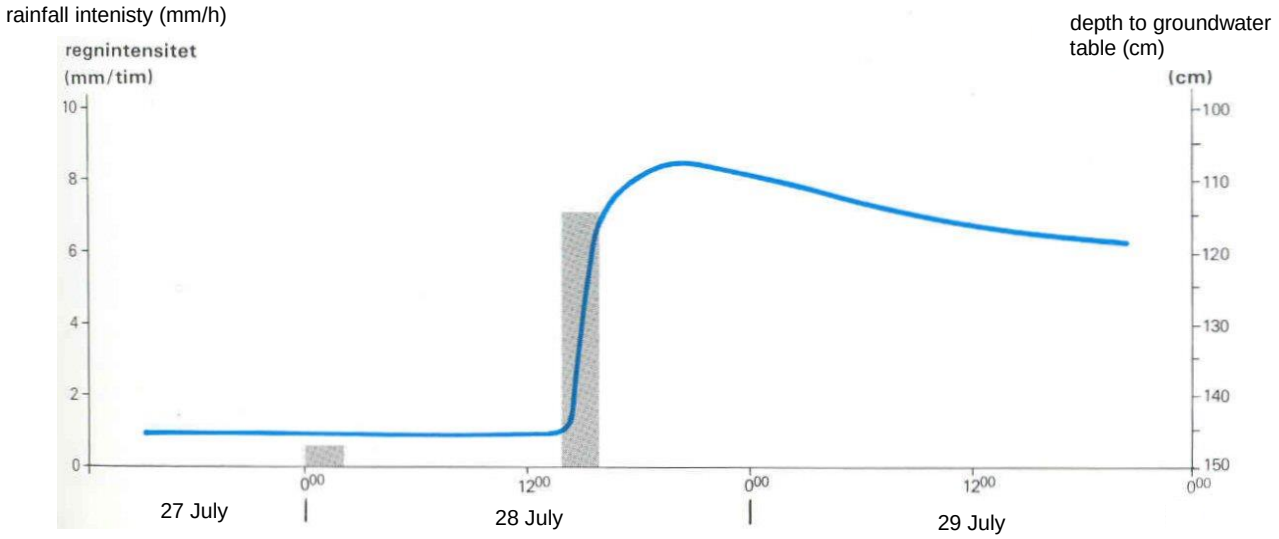


Fig. 29 The groundwater level can react fast and with large change even when the rain is relatively small. In the afternoon on the 28 of July, 1983 14 mm rain fell during two hours (7 mm per hour) on a till hillslope at Lagga, Uppland, Southern Sweden. Within a few hours the groundwater level had risen almost 40 cm, that is, 400 mm.

given flow. The pressure propagation velocity is on the other hand higher. Fig. 29 shows how the groundwater table in a till hillslope reacts on a rainstorm. The groundwater level at 1.5 m depth reacts within one hour after the start of the rain. At this storm the rainwater particles reached a maximum depth of about 4 cm. When the velocities of the water particles were calculated according to the equation on page 38, piston flow was assumed (cf. page 40). Water held by adsorptive forces was neglected, as well as possible flow in large structural pores. There are good reasons for these assumptions in sandy soils but less so in till soils, where soil structure can be important for the flow at saturated or near saturated conditions.

Water movement followed by a natural tracer

A case of piston flow in the unsaturated zone is shown in Fig. 30. Here the stable oxygen isotope ^{18}O (oxygen-18) was used as a natural tracer. First we give some details about this tracer, which we will refer to later.

About 0.2% of the water molecules in natural water contain ^{18}O-atoms instead of the most abundant ^{16}O-atom. The ^{18}O-concentration in precipitation has a yearly variation with low values during winter and high values during summer (Fig. 30, right hand curve). The variation is small and does not affect the density or flow characteristics of water, but it gives a natural labelling of precipitation water for different seasons and also for different precipitation events.

After soil water sampling in May 1982, current as well as last year's winter precipitation could be identified in Fig. 30 at 0.7 and 2.3 m depth, respectively. Later samplings showed how the ^{18}O-wave was successfully pushed downwards when new water was added at the top. (The picture is somewhat obscured by the transpiration that is using water from the root zone.) The identification in May of last year's meltwater at 2.3 m depth results in a particle mean velocity of 0.2 m per month.

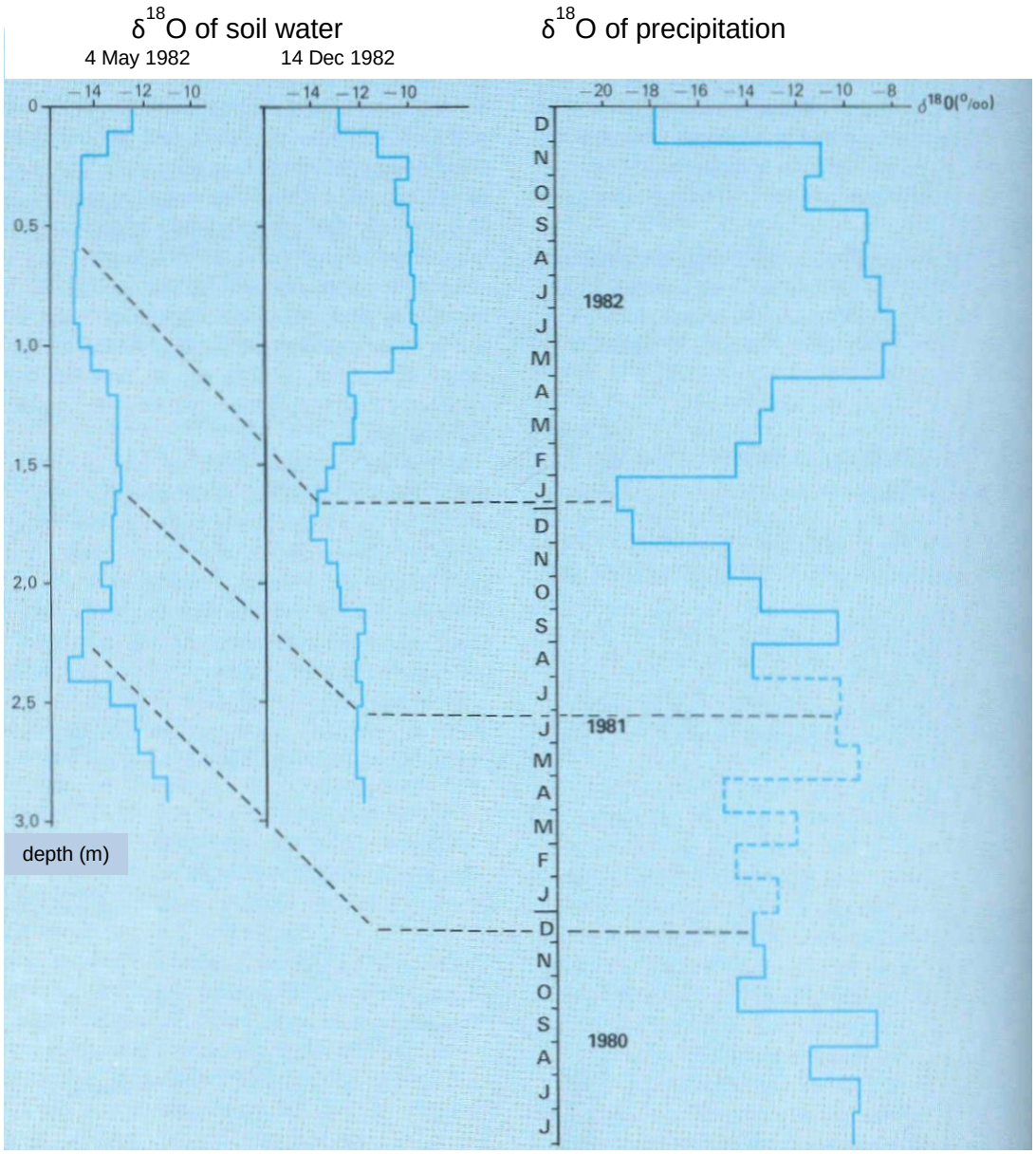


Fig. 30 The ^{18}O-concentration of soil water in a sandy soil profile in Uppsala, Central Sweden, at two occasions and the monthly mean ^{18}O-concetration in precipitation the years before. During the first half of 1981 the ^{18}O-concentration of the precipitation was estimated from nearby stations (hatched line).

The seasonal variation of the ^{18}O-concentration in precipitation can be traced in the soil profile. In May 1982 the winter precipitation with low ^{18}O-concentration was found about 0.7 m below the soil surface and last years' at about 2.3 m depth. During the summer and autumn until December 1982 the soil water particles were moved one metre downwards.

The notation δ^{18}O gives the deviation of the sample ^{18}O-concentration from the ^{18}O-concentration of an international standard water.

In connection with the ^{18}O-sampling the water content at the different levels was also determined (not shown in Fig. 30). The water content between two levels is according to page 19 given by the mean water content between the levels multiplied by the thickness of the layer. The mean water content between 0.7 and 2.3 m depth was 17.5% in May. This gives total water content of 280 mm between the two levels. The flow below half a metre depth can be regarded as downward during the whole year, because it is in a sandy soil with relatively deep groundwater table. The total percolation during the year is therefore equal to the water content between the two levels, that is, 280 mm. This is a reasonable value of the groundwater recharge during a year.

There was no overland flow at the site and all precipitation that is not evaporated will form groundwater (cf. yearly runoff in Fig. 6).

A similar regular variation in the ^{18}O-concentration of soil water with depth was found on till soil in the Kloten-area in Västmanland, Central Sweden. From a sampling in September the water from last year's snowmelt could be identified at 1.5 m depth.

Below the root zone water seems to move as piston flow

The fact that the yearly variation in ^{18}O-concentration of the precipitation can be observed in the soil water zone and that the groundwater recharge calculated as above gives a reasonable value indicate that the soil water moves as piston flow below the root zone. The whole water volume is successively moved down through the soil when water is added at the top. The soil water particles at a certain level have about the same age. If a considerable flow had taken place in large interconnected pores the yearly variation should have vanished when water from a certain infiltration event would appear at a certain level at different times. Or should the yearly variation have given a too low value of the groundwater recharge, since a large proportion of the infiltrating water would have reached the groundwater level almost directly.

Partly contradictory results have been reported from experiments with added dye as tracer. The soil surface was irrigated with dyed water and after the irrigation was finished the penetration of the dye was studied in pits that were dug. In sandy soil as well as in till soil an irregular dye pattern was observed indicating a variable percolation rate of the water particles. The experiments were, however, done with very high artificial rain intensity and their information on natural conditions is therefore limited (cf. the discussion on page 40).

Confined and unconfined aquifers

An *aquifer* is a geological formation that contains groundwater. Often a practical aspect is added and aquifer is used for a formation from which groundwater can be extracted. With this usage a clay soil, for example, is not an aquifer even if it, which is the normal case, contains groundwater. Two different aquifers are identified, uncofined and confined.

In an *unconfined aquifer* the groundwater surface (defined as the level in the soil where the water surface in an inserted perforated tube is located) coincides with the upper level of the groundwater zone. This is the most common type of aquifer in Sweden, for example the one that is shown in Fig. 14. It is also called aquifer with free groundwater surface.

A *confined aquifer* may exist where a sandy soil is overlain by clay soil. The clay soil is impermeable as compared with the sandy soil and functions as a cover above the sandy aquifer. The groundwater in the sand is often exposed to overpressure and the water surface in a tube, perforated only in the sand, will rise above the upper limit of the sand, up into the clay layer and often even up above the soil surface. The sand deposit is a confined aquifer (also called artesian). The groundwater surface is in this case a pressure level, called piezometric pressure head. It is, as well as in the open aquifer, the level at which the water in an observation well is adjusted.

In the valleys in Central Sweden there are often till aquifers below the surface clay layers. These confined aquifers, which pressure head is often above the soil surface, get their water from infiltration in the till hillslopes around the valley bottoms. The pressure head of the groundwater

in the confined aquifer is among other things determined by the level of the free groundwater surface in the till deposit in the surroundings of the clay soils. There is also a groundwater level in the clay layer, a free groundwater surface that under natural conditions often is lower than that in the underlying confined aquifer.

In the coming discussion we will only occasionally treat water flow in confined aquifers. We will keep to the free groundwater surface in the till deposits.

The *storage coefficient* of an aquifer denotes the relation between change in storage and water level change.

$$M = \Delta S / \Delta h$$

M = storage coefficient
ΔS = change in storage
Δh = change in groundwater level

The storage coefficient is equal to the volume fraction of water that is drained when the groundwater level decreases. In an uncofined aquifer, the storage coefficient is also called *specific yield*. From the earlier definition (page 39) we can see that the specific yield is the difference between porosity and field capacity.

The storage coefficient may be up to 20% in coarse sand and around 2 – 5% in till. If the storage coefficient is 5% and 20 mm water is added to the groundwater it will give an increase of the groundwater level of 20/0.05 = 400 mm or 0.4 m.

In a confined aquifer the storage coefficient is not determined by the water holding properties of the soil, but of the ability of the soil matrix to stand pressure changes in the water. Storage coefficients in confined aquifers are very small, often only parts of percentages. Small storage changes, caused by for example groundwater extraction, result in large changes in groundwater level (the pressure head of the groundwater).

Different ways to estimate the rate of groundwater recharge

The rate of *groundwater recharge* determines how large the groundwater extraction can be from an aquifer without a permanent decrease in the groundwater level. The amount of groundwater that can be extracted from a certain well is also dependent on the hydraulic and water-holding properties of the aquifer and on the construction of the well.

In Sweden the groundwater recharge in an area is somewhat smaller than the runoff from the area. Over a long time period the groundwater that dominates runoff must be recharged at the same rate as it is discharging to the streams.

Groundwater recharge does, however, by definition only take place in recharge areas. Expressed per unit area recharge area the groundwater recharge is equal to the specific discharge (runoff per unit area) for the whole catchment, or perhaps a little larger. This somewhat paradoxical relationship could be explained in the following way:

If there is no Hortonian overland flow on the recharge area the groundwater recharge over a long time period is equal to the difference between precipitation and evaporation on that area. The specific discharge from the catchment is equal to the same difference for the whole catchment, that is, recharge and discharge areas. The evaporation should be largest from the discharge areas, where the water availability is always good. The precipitation can on the other hand be regarded as equal in size over the two types of areas. The difference between precipitation and evaporation will consequently be largest on the recharge areas, which means that the groundwater formation per unit area of recharge area will be somewhat larger than the specific discharge from the whole catchment.

The total groundwater recharge in an area will increase when the groundwater level decreases due to for example large extractions or drainage. If the groundwater level decreases, the extent of the discharge area, that is, areas where no groundwater recharge takes place, will also decrease. The term *potential groundwater recharge* denotes the groundwater recharge that would take place in an area if the whole area was recharge area. The potential groundwater recharge is therefore the groundwater recharge per unit area recharge area.

The specific discharge gives a, perhaps somewhat too low, mean value of the potential groundwater recharge within a catchment from which discharge data exists. Neighbouring, relatively large catchments (some km^2) have about the same specific discharge because the variations in evaporation that exist within a catchment are smoothed out. The maps over the long term mean specific discharge shows this mean discharge from the landscape with their mosaic of different soil types and soil usage. When water supplies are planned the focus is more on the groundwater recharge in a certain aquifer and it can deviate substantially from the area mean.

The groundwater recharge in a point or in a given aquifer may be calculated in a number of different ways. Some principally different methods with different possible time resolutions and different demands on measurements are:

- Water balance calculations in the root zone. The water content in the root zone is calculated from for example daily values of precipitation and potential evaporation, the latter reduced to actual evaporation with the help of calculated soil water content. When the water content reaches the field capacity a volume comparable to additional input from precipitation is assumed to percolate. This method requires among other things knowledge of the water holding properties of the root zone.
- Percolation calculations with Darcy's Law. This requires a detailed knowledge of the unsaturated hydraulic conductivity and water holding properties of the soil in the root zone.
- Percolation calculations with natural or artificial tracers. The velocity of the water particles, together with measured water content will give the groundwater recharge (cf. the example on page 56). This calculation requires the percolation to be of the piston flow type.
- Calculation of horizontal groundwater balance within an area for a long time period. The groundwater recharge in an area is equal to the difference between the groundwater outflow from the area and the groundwater inflow to the area. The calculations require information on the topography of the groundwater surface and the transmissivity.
- Calculation of the groundwater storage from measured increase in groundwater level during all episodes with increasing groundwater level during a long period. The method requires that the groundwater flow is small during the episodes. The specific yield of the aquifer must be known.
- Long-term pumping from wells with defined recharge areas. The mean discharge at equilibrium is equal to the groundwater recharge.

This list was done to illustrate the groundwater recharge from different aspects. An evaluation of the different methods will not be done. In practise there is no choice; one has to accept the local conditions and the data that are available.

Groundwater recharge in a pine forest on sandy soil

In Fig. 31 the percolation in a pine forested sandy soil in Central Sweden is calculated by means of a model that combines the first two methods mentioned above for some summer periods. In spite of a number of rain occasions (not shown) during the periods, percolation and groundwater recharge only take place at a few occasions. This soil has an unusually well-defined field capacity. It is a sandy soil with relatively uniform pore sizes and there is a coarse sand layer below the root zone that acts as a valve for the soil water flow. When the soil is dry, that is, at low water pressure the hydraulic conductivity of the coarse layer is very small because nearly all pores are empty. When the pressure of the soil water increases to a certain value just below atmospheric pressure, the coarse pores are filled with water and the sand above is drained with a rate determined by the conductivity of this layer. The valve is therefore either completely closed or completely open. In a soil with finer texture the "valve" should open and close more gradually (cf. page 32).

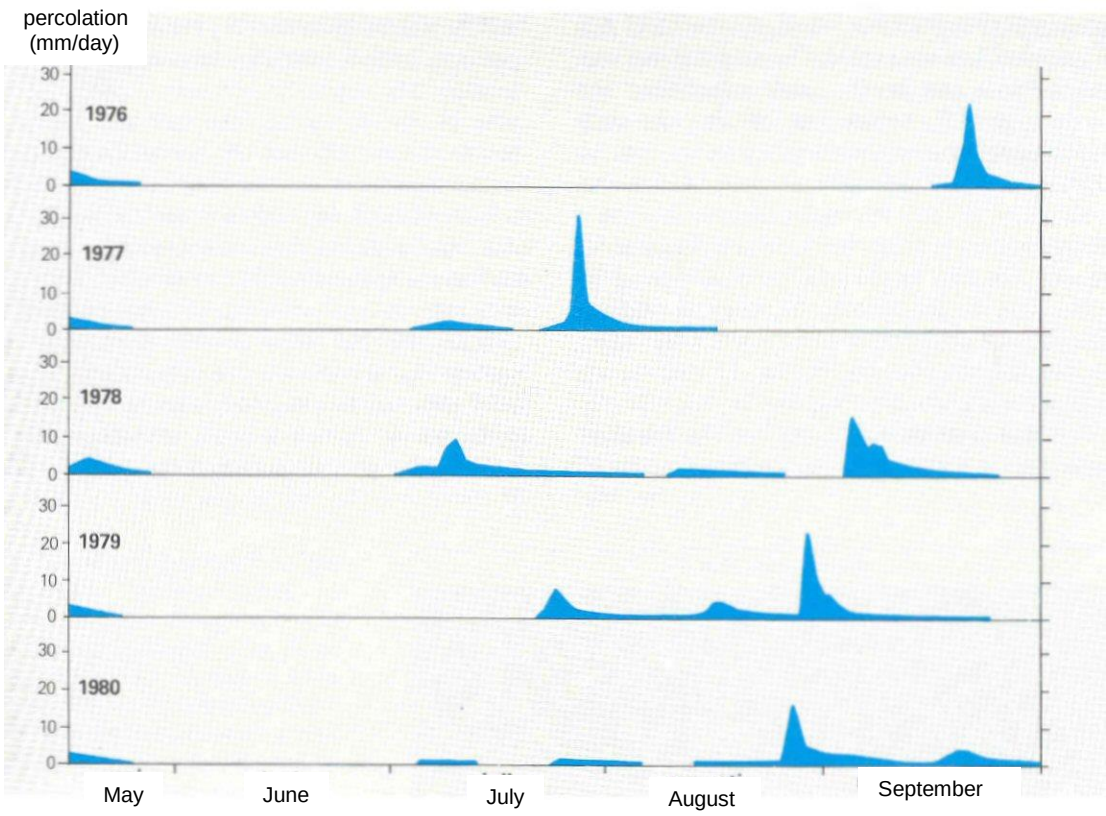


Fig. 31 Calculated percolation below the root zone in a pine forested sandy soil, Jädraås, Gästrikland, Central Sweden. In this well sorted soil the field capacity of the root zone is extraordinary well defined. The percolation takes place only at a few occasions when this field capacity is exceeded. At lower soil water contents all infiltrated water is stored in the root zone and a large part is lost by transpiration

The groundwater recharge during the summer periods shown in the figure varied between 69 mm (the dry year 1976) and 157 mm (1978). These periods also had the smallest and largest precipitation, 235 and 365 mm, respectively. In this case, the largest amount of rain coincided

with the largest groundwater recharge. Due to interception, water storage in the unsaturated zone, and the transpiration during the period it is not the total rain that falls during a period that determines the groundwater recharge. A few large storms will give a larger groundwater recharge than many small rains, even if the total amounts of rainfall are the same. It should be noted that the largest groundwater recharge during a year at this location, as well as in other places in Central and Northern Sweden, takes place during snowmelt. This period is not shown in Fig. 31.

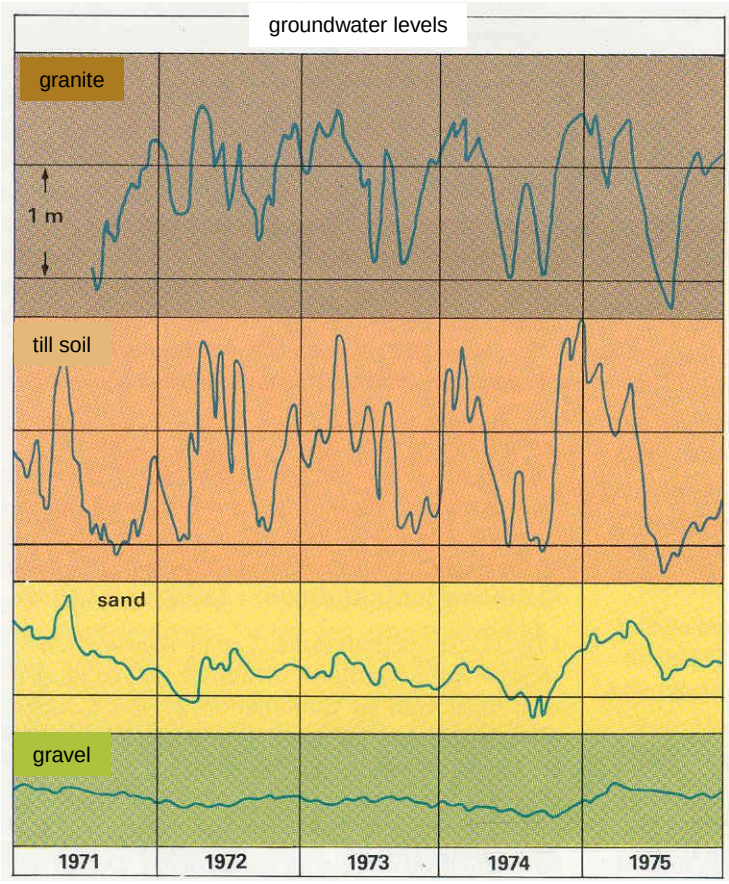


Fig. 32 Observed groundwater levels in the Tärnsjö area, Västmanland, Central Sweden. The fluctuations of the groundwater levels differ characteristically between different geological deposits. In the gravel deposit a lot of water is needed for a given rise in the groundwater level. The water content is low in the unsaturated zone and there is a large empty pore volume that must be filled. In the fine textured till soil the water content is relatively high in the unsaturated zone. Only a small water input is needed to fill the remaining pore volume and increase the groundwater level. The variations are small in the gravel deposit and large in the till.

The figure is based on bi-weekly observations. The fluctuations are probably larger in the till and granite than what is shown. The level fluctuations are rapid here and continuous measurements would be needed to catch the real course.

Large yearly variation in groundwater levels in till soil – small in gravel

The increase in groundwater level at a certain groundwater recharge is, according to page 60, inversely proportional to the storage coefficient of the soil. In Fig. 32 the groundwater levels in different aquifers are compared. In the till aquifer with the lowest storage coefficient the groundwater level varies with 2 m during a year; in the gravel aquifer it varies only with about one dm.

The difference between the two soils is exaggerated by the smoothing of the percolation peaks that takes place in the unsaturated zone. In the till soil with its high water content and its shallow groundwater the groundwater can respond within one or a few hours (cf. Fig. 29). In many gravel- and sand deposits, as for example eskers and other glacio-fluvial deposits, the groundwater level is low leading to a delay of the pulses of groundwater recharge and they will be smoothened before they reach the groundwater. In Fig. 32 a delay of about one month can be seen between the variations in the till and gravel deposits, respectively.

The temporal variation in groundwater level is not only dependent on storage coefficient and depth to the groundwater surface. The change in level in a point is an effect of the groundwater balance in that point, that is, the difference between groundwater recharge and groundwater inflow on one hand and groundwater outflow on the other. Therefore the variation in groundwater levels is different, for example, along a hillslope even if the soil is the same.

The start of a stream is often well defined. It is not small, tiny rivulets that merge and form a stream. Here the stream starts in a permanent discharge area, an alder fen in a hollow. Spring flood in Saltsjöbaden, Södermanland, Southern Sweden. Photo: Allan Rodhe.

5. Water in discharge areas

In *discharge areas* groundwater is tapped. This tapping takes place in two ways. As an unsaturated outflow to the root zone, where the water is taken up by the plants and is transpired back to the atmosphere (in some cases plant water uptake is directly from the groundwater) and as a saturated outflow from the soil to the soil surface or directly to the streams.

Discharge areas can be defined as "areas where the groundwater has a flow component directed out from the groundwater zone". Discharge areas are therefore recognised as areas where the total potential increases with depth, that is, the water level in observational tubes with intake openings at depth is higher than in tubes with shallow intake openings. We will see later that the extent of the discharge areas wary with time and is related to groundwater fluctuations.

The topography gives recharge- and discharge areas

In Fig. 33 calculated total potential and streamlines for the groundwater flow in a homogenous aquifer with undulating groundwater surface are shown. The ground surface is not shown in the picture. We can assume that it equals the groundwater surface in the valleys and is somewhat higher than the groundwater surface on the hills. The figure shows an equilibrium state. A prerequisite for this is that there is a continuous recharge from above. Otherwise the groundwater surface would have declined in the recharge areas. A further prerequisite is that there is a continuous discharge of water from the discharge areas. Otherwise the groundwater surface in those areas would have risen.

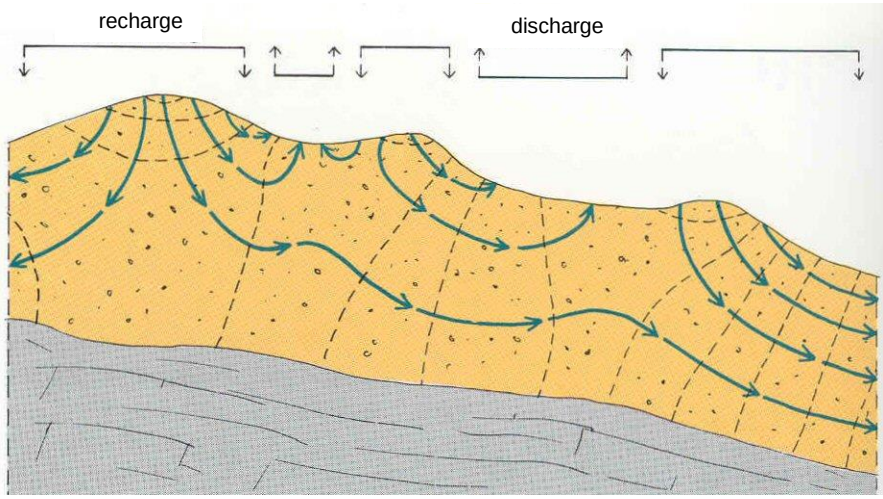


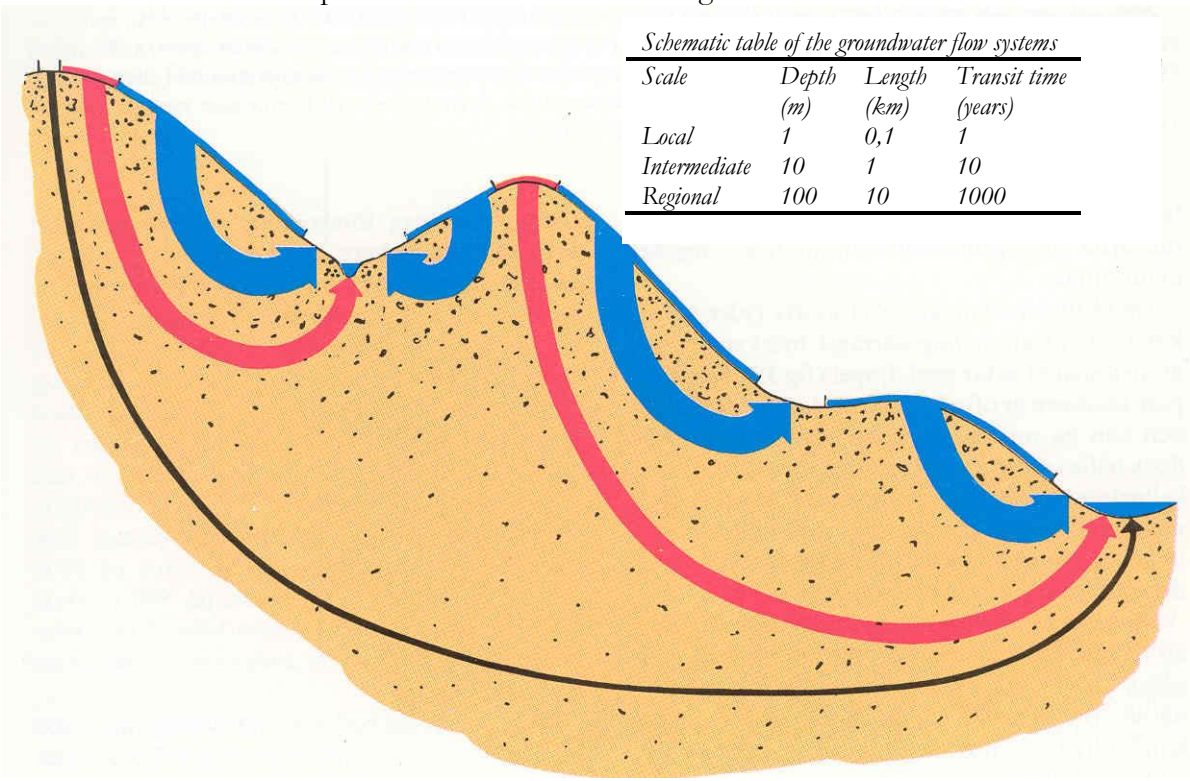
Fig. 33 Calculated total potential (hatched lines) and streamlines for groundwater in a homogenous aquifer with undulating groundwater surface. The figure has equal height- and length scale. The absolute scale is arbitrary, but the depth of the aquifer is put to 100 m giving a reasonable topography of the groundwater surface and also for the soil surface.

The undulating groundwater surface results in groundwater recharge on the hills and groundwater discharge in the valleys.

Dupuit's assumption is not applicable for calculation of the flow pathways of groundwater (page 35). Dupuit assumed that the total potential was constant with depth, giving a horizontal flow at all depths. In Fig. 33 it is just the variation in total potential with depth that is calculated with a more complicated application of Darcy's Law. When the total potential has been calculated streamlines can be drawn. These are always crossing the flow lines under right angle. The water flow between two streamlines is constant along the flow because no flow can take place across streamlines. If the streamlines, as in this case, are drawn in such a way that the total flow between two streamlines are equal between all pairs of streamlines, the streamlines show the flow

pathways as well as the magnitude of the flow. The flow in every point is inversely proportional to the distance between the lines. Dense streamlines thus indicate large groundwater flow.

From Fig. 33 and comparable analyses of the groundwater flow in different topographical situations a number of important conclusions about the groundwater flow have been drawn:



Schematic table of the groundwater flow systems

Scale	*Depth (m)*	*Length (km)*	*Transit time (years)*
Local	*1*	*0,1*	*1*
Intermediate	*10*	*1*	*10*
Regional	*100*	*10*	*1000*

Fig. 34 The groundwater at different depths and in different positions in the landscape has different origins. The deeper below the ground surface and the lower in the landscape, the further away may the groundwater come from and the older it can be.

- If the groundwater surface follows an undulating ground surface recharge takes place on local hills and discharge in local valleys.
- There is groundwater flow even at large depths, but the magnitude of the flow decreases rapidly with depth.
- Different flow systems with recharge and discharge can be recognised, from local to regional scales. The local flow system consists of the nearest slope to a discharge area. The main part of the groundwater that is formed in an undulating landscape discharges in local discharges areas. The further downstream in a catchment one is, the larger is the proportion of the discharging groundwater that could have been formed in recharge areas further away and higher up than the local ones. The proportion of groundwater that has been transported long distances is small, but may be of importance for the chemical composition of the groundwater. With regional scale we mean distances of tens of kilometres. The regional flow is probably very small.

The real recharge area in a local system may consequently be a little smaller than what the local topography shows, because some of the infiltrating water finds its ways further down in the main slope of the landscape without discharging in local discharge areas. This "groundwater leakage" might be seen as an increasing specific discharge when the catchment area increases. The effect is however hard to estimate because precipitation and evaporation also vary.

- The groundwater age increases with soil depth and also with distance from the water divide. The youngest groundwater is therefore found in shallow wells on hilltops and the oldest in deep wells in valley bottoms.

What was said above is valid for the amount of groundwater flow at different points, but not for the particle flow velocities, with the exception of the comment on the groundwater age. The particle velocity is, as stated above, given by the quotient between flow and porosity. If the porosity is constant, as in Fig. 33, the particle velocity is proportional to the flow in every point, that is, inversely proportional to the distance between streamlines. Broadly speaking this is also true for a soil with varying porosity because the variation in porosity is comparatively small.

The situation may be different in bedrock with cracks. The mean porosity may be very small, perhaps only tenth of a percentage, but have large variations. Water is flowing in a few cracks, which relative volume to the total rock volume is very small. The particle velocities of water may therefore be much larger than they normally are in soil deposits, even though the flow per cross-sectional area of the rock is smaller. The distance between the streamlines in a soil and in a rock, respectively, are therefore not directly comparable in terms of water particle velocity and age.

The age of the groundwater at different depths has been estimated, for example, from studies of natural radioactive and stable isotopes in water (for example carbon-14, tritium and oxygen-18). Such investigations have given very varying results for the age of deep groundwater in the bedrock, with estimated ages from thousands of years to months.

The conductivity of till soils vary with soil depth

The hydraulic conductivity of the soil in Fig. 33 was assumed to be constant down to the impermeable underlying rock at 100 m depth. In the till landscape the conditions for groundwater flow are quite different.

In the few investigations that have been made it seems that the conductivity is high close to the soil surface, but that it decreases rapidly with soil depth (Fig. 35). The deeper till soil is often more compacted than the upper soil layers and may have low conductivity. It should be remembered though that the composition of a till soil profile may vary with soil depth. It is for example common with sandy or gravely layers with high conductivity here and there down the profile, and they can have high conductivity and can have a large impact on the groundwater flow.

The till soil depth is seldom thicker than one or a few metres on the hills, but may be some tens metres thick in the valley bottoms. The bedrock below the till soil may have a conductivity that is higher or lower than that of the overlaying till. At the border between till and bedrock a crushed zone with high conductivity is often found. In solid rocks the conductivity is determined by the frequency of cracks, in turn dependent on tectonic conditions and type of rock. The rock is, for example, often crushed near faults. Further, granite is rich in cracks as compared with some gneiss that have low crack frequency. The uppermost 50 metres may be especially rich in cracks and have relatively high conductivity. Examples of conductivity measured in Swedish solid rocks are 10^{-7} m/s at 50 m depth and decreasing to 10^{-10} m/s at 500 m depth (mean conductivity measured in bore holes in 25 m sections of the rock that is including tight rock as well as cracks).

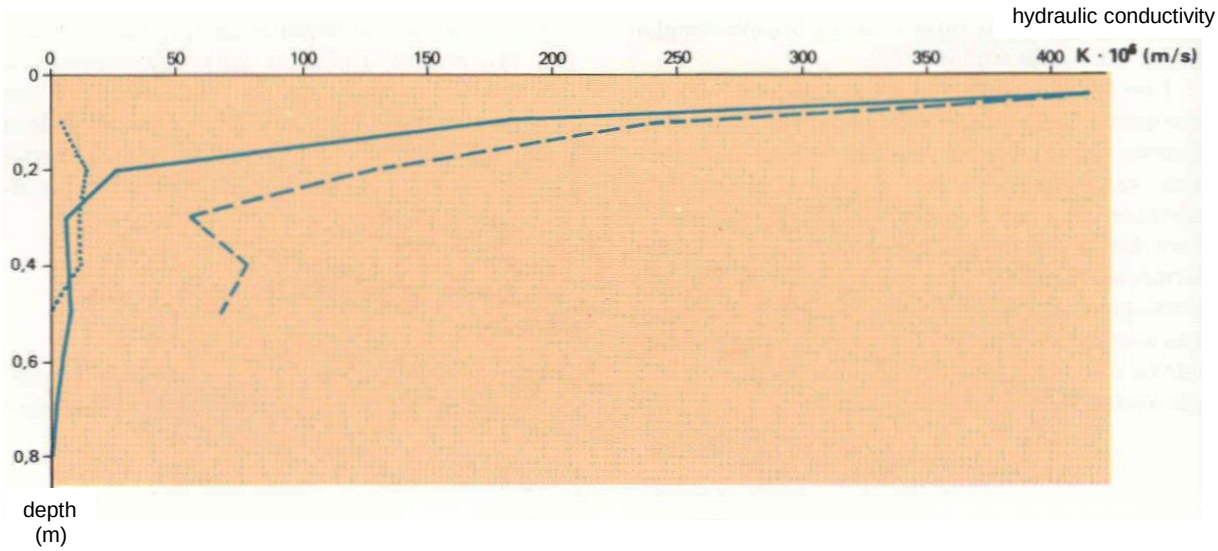


Fig. 35 Saturated hydraulic conductivity, K, as a function of depth in a till soil. Measurements were taken at three different locations in the catchment of Masbybäcken, Kloten, Central Sweden.

A number of factors act in the same way resulting in a maximum hydraulic conductivity near the soil surface. There the till soil is least compacted because it does not have the burden of soil layers above. There the soil is loosened by soil processes as decomposition of dead roots, digging worms and insects, leaching of minerals, downward transport of fine textured materials and effects of soil frost.

The increase in hydraulic conductivity towards the soil surface was obvious in this area, even if one of the curves deviates. There are few measurements on till soils at different locations in Sweden.

The principal picture that is shown in Fig. 33, and the conclusions drawn from it, is valid even when the conductivity decreases with depth. The difference is that the shallow flow is dominating even more. The streamlines close to the surface will therefore be closer to each other than streamlines at greater depth. The figure may be seen as depicting the flow lines of the deeper groundwater. Above this and hardly seen in the depth scale of the figure the most shallow groundwater is added. Except at the top and at the foot of the hillslopes the streamlines are nearly parallel to the groundwater surface.

In solid rock the flow takes place in the cracks. The streamlines can therefore not be expected to be as smooth as those in the figure where the soil was homogenous.

Discharge areas may be saturated or unsaturated

We have defined discharge areas as areas where the groundwater has an upward flow component. It may be appropriate to distinguish two types of discharge areas. In a *saturated discharge area* the groundwater surface is at the soil surface. Here no infiltration can exist and all precipitation will form saturated overland flow together with discharging groundwater. In an *unsaturated discharge area* the groundwater does not reach the soil surface. An unsaturated flow up into the root zone takes place during dry periods. During wet periods the water flow may be directed downwards in the unsaturated zone, as Fig. 36 shows, at the same time as it is directed upwards in the saturated zone. A condition for this is that the hydraulic conductivity is large at the groundwater surface. The groundwater discharges into a coarse layer that is able to laterally transfer the up-welling groundwater as well as the percolating soil water. Such unsaturated discharge areas are probably common in till soils due to the large vertical variation in hydraulic conductivity. The soil surface may therefore be unsaturated also in connection with large rain storms on places where the topography indicates that there should be groundwater discharge.

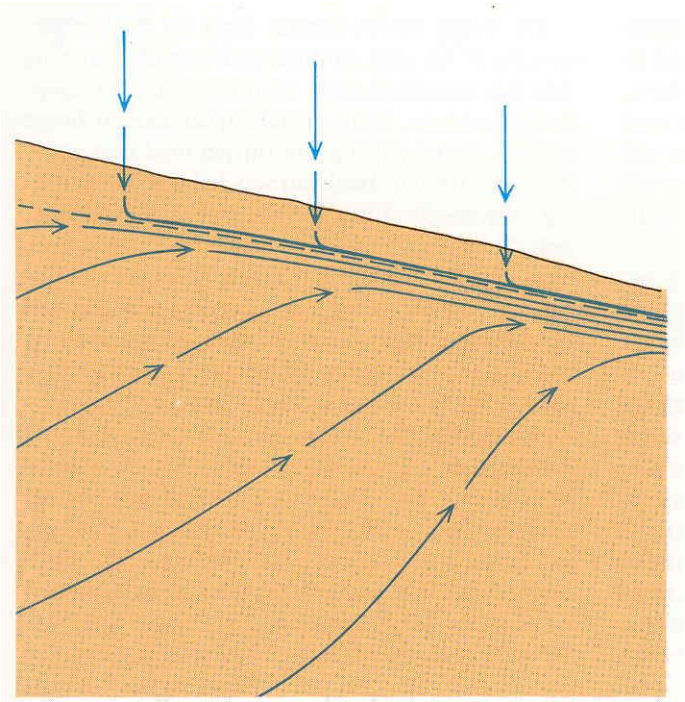


Fig. 36 Unsaturated discharge area. The high hydraulic conductivity in shallow layers result in a lateral diversion of the discharging groundwater and the soil surface will not saturate. During dry periods the up-welling groundwater contributes to plant water supply. If the conductivity is high enough close to the soil surface the surface can remain unsaturated even when infiltration takes place, as shown in the figure.

Discharging groundwater of different age

The chemical composition of groundwater reflects the groundwater flow pathways and residence times up to the sampling point. The electric conductivity of the water (see square) is an easily determined measure of this. Generally the salt content and therefore the electric conductivity increases with groundwater age, see Ch. 7.

The ability of a water sample to transmit electrical current, it's so called *electric conductivity,* is a useful measure of salt content. Since it is the dissolved salts in the water that transmit the current the electric conductivity is largely proportional to the concentration of dissolved substances, i.e. its salt content.
The electric conductivity has many names; often it is named just conductivity. We use electric conductivity to separate it from hydraulic conductivity, which sometimes also is named conductivity. The risk for confusion is normally small since electric conductivity is a quality of the water, while the hydraulic conductivity is a quality of the soil or bedrock.
The unit for electric conductivity is mS/m (millisiemens per meter).
Because of their small dimensions, hydrogen ions transmit electrical current five to ten times more effectively than other common ions in natural water. For the electric conductivity to be a measure of total salt content in a water sample the contribution from the hydrogen ion is normally subtracted. This is especially important in acid waters. When we use the concept of electric conductivity we have always subtracted the contribution from the hydrogen ions.

In Fig. 37 the electric conductivity of the shallow groundwater at the lower part of a hillslope can be related to the flow pathways of the groundwater. At every sampling place the salt content increased with soil depth. The highest salt content was observed at the bottom of the hillslope, where the total potential indicates an upward flow component. Here comparatively old

groundwater, rich in solutes, is flowing. A closer look at Fig. 37 shows that flow pathways and salt content do not coincide in detail. If streamlines are drawn through the lowest sampling points a decreasing electric conductivity is found in the direction of the flow. One reason for this may be errors in the measurement of the electric conductivity as well as of the total potential. A more important fact is that the potential lines give a snapshot view, the situation just when the measurements took place, while the flow pattern is actually varying with the water recharge from above. A perfect agreement between streamlines and salt content cannot be expected.

The different flow pathways up to the stream give, as Fig. 37 indicates, a varying composition of the discharging groundwater. The deeply transported older groundwater has a higher salt content than the shallow groundwater with short residence time in the soil. The electric conductivity of the streamwater in Fig. 38 shows how the relative share of "deep" and "shallow" groundwater varies (in the same stream as in Fig. 37) along the stream and between different time points. At both sampling occasions there was only groundwater flowing to the stream because long time had passed since last rain. In both cases the salt content increased along the stream. A reasonable

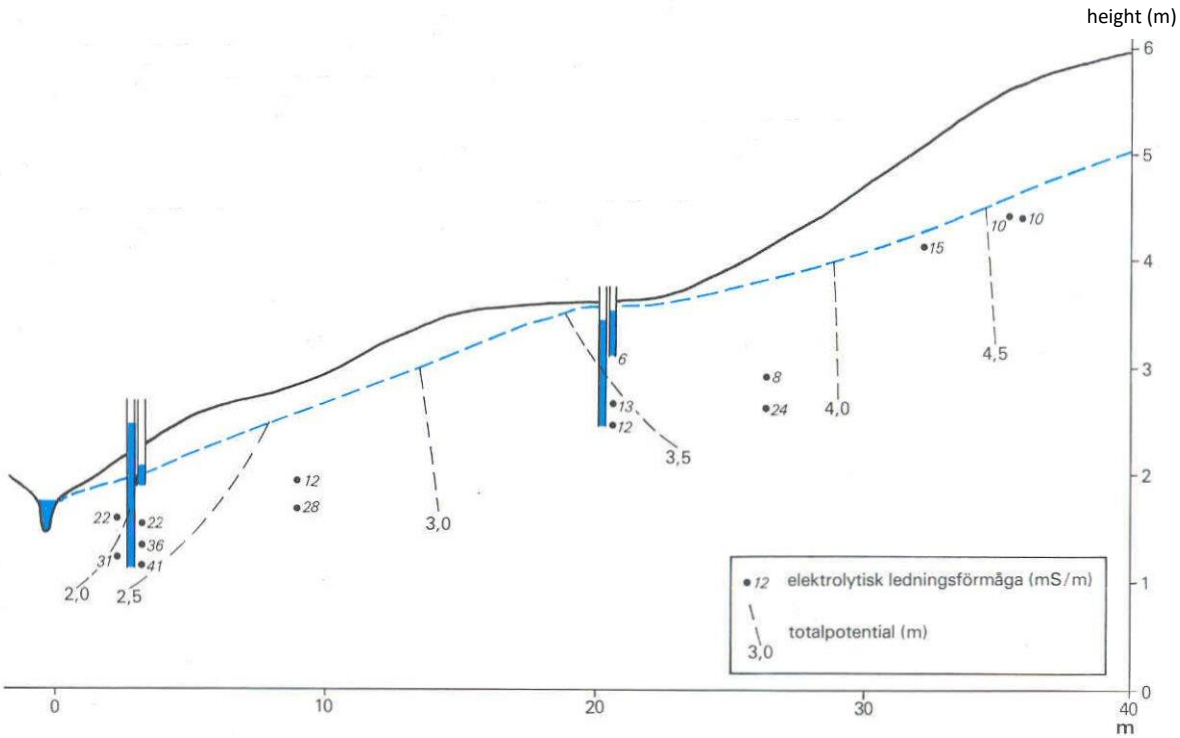


Fig. 37 Total potential and electric conductivity (salt content) of the groundwater in the lower parts of a till hillslope in Kilmyrbäcken, Jämtland, North West Sweden, 1 October, 1981.

The total potential has been determined in different points by water level measurements in tubes with intake openings only at the bottom. The total potential in a point = the vertical distance from a reference level to the point + the vertical distance from the point to the water level in the tube = the vertical distance from the reference level to the water level. At the groundwater surface the total potential is equal to the vertical distance from the reference level, because the pressure potential is equal to zero.

The flow is perpendicular to the total potential lines and in the direction of decreasing potential.

In the groundwater tubes closest to the stream the water level in the deeper tube is higher than in the shallow tube. The total potential decreases upwards in the soil profile and the groundwater has an upward directed component, that is, there is groundwater discharge. The next pair of tubes shows groundwater recharge.

The change in electric conductivity with depth and along the hillslope results from differences in flow pathways and ages of the groundwater.

explanation is that the relative proportion of deep, long transported groundwater increased further down in the catchment when the distance to the water divide increased (cf. Fig. 33). The increase of the proportion of deep groundwater in streamwater should be faster in the very upmost streams that don´t have any tributaries. This stream reaches almost one km before it gets its first tributary, a stream of about the same size as the original one. In this case the increase in salt content is completely stopped after the junction. We think that the reason might be the mire, but we have not been able to find any satisfactory explanation to this phenomenon (see also "Chemical processes in discharge areas", page 101). (The reason for the sudden decrease in electric conductivity after the junction at one of the sampling occasions was that the tributary at that time had lower conductivity than the main stream.)

The difference in salt content between the two sampling occasions is probably explained by higher proportion of shallow groundwater, with short residence time in the soil, at the high flow situation.

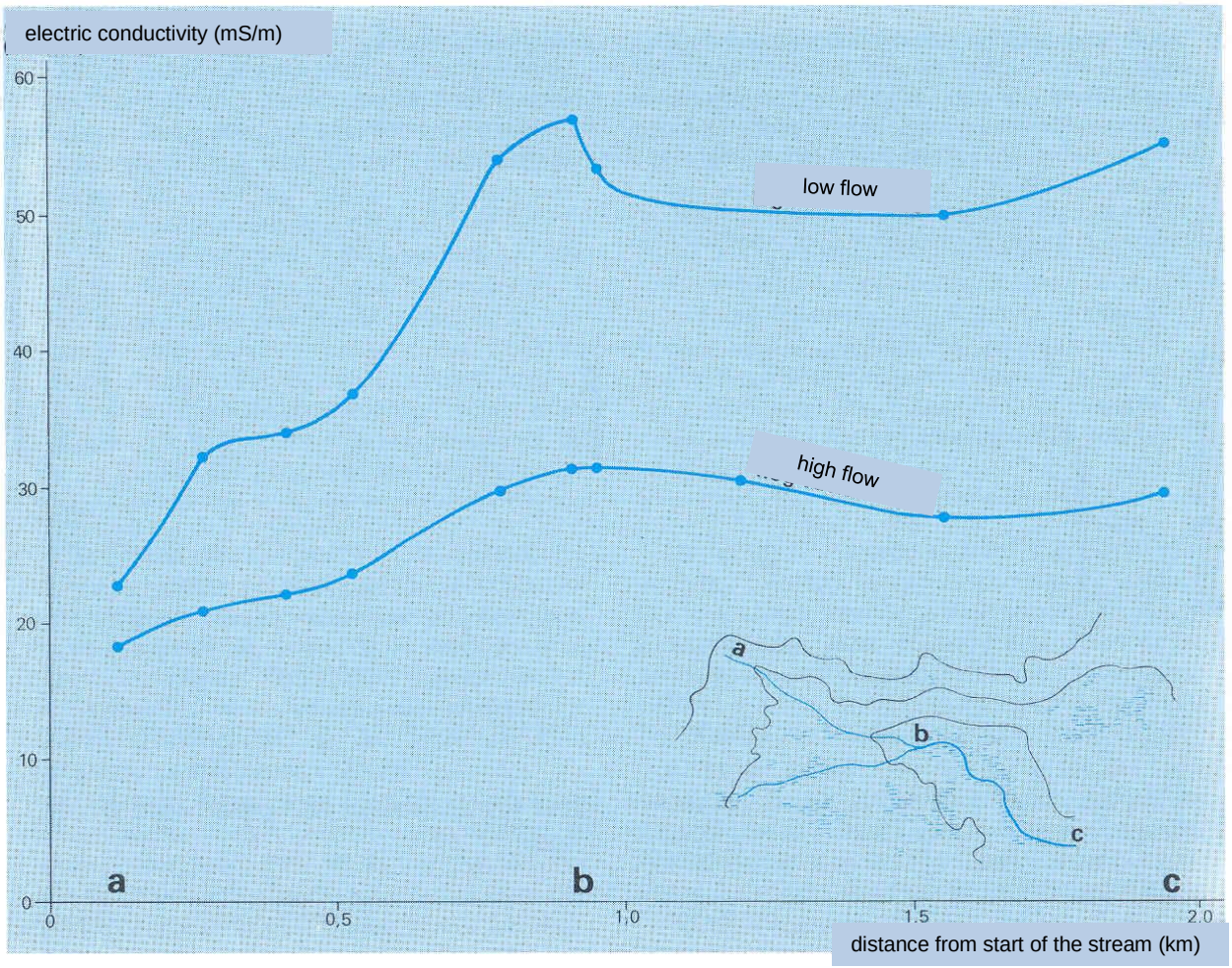


Fig 38 Electric conductivity along the stream Kilmyrbäcken, Jämtland, North West Sweden at two different occasions; 26 August (low flow) and 27 July (high flow), 1979.

The salt content increases along the stream, probably because the proportion of deep, long transported groundwater increases downstream. At high flow the proportion of shallow groundwater with low salt content is high all along the stream.

Large increase in runoff when the groundwater level increases

Fig. 33 and the principal discussion about it started from an equilibrium condition where the groundwater level and the flow pattern did not change with time. Actually the groundwater level has large and often rapid variations. The discharge of deep groundwater, however, changes relatively little when the groundwater level rises in a recharge area. The vertical outflow is proportional to the height difference between the groundwater levels of the recharge- and discharge areas. This difference may be tens of metres before an infiltration event and even if the groundwater level increases one metre in the recharge areas the relative changes are limited. When shallow layers with high hydraulic conductivity are activated and contribute to a lateral flow in parallel with the soil surface, the outflow of shallow groundwater can increase substantially.

Fig. 39 shows the groundwater surface along the lower parts of a hillslope at two occasions, one with normal groundwater level and one in the middle of the spring flood when the groundwater level is at its maximum. The slope of the groundwater surface coincides in both cases well with the soil surface. The water flow is much larger at the higher water level. The cross sectional area that contributes to the flow has increased with the higher groundwater level. But above all the reason for the increase is that the shallow layers with high hydraulic conductivity have started to contribute to the flow.

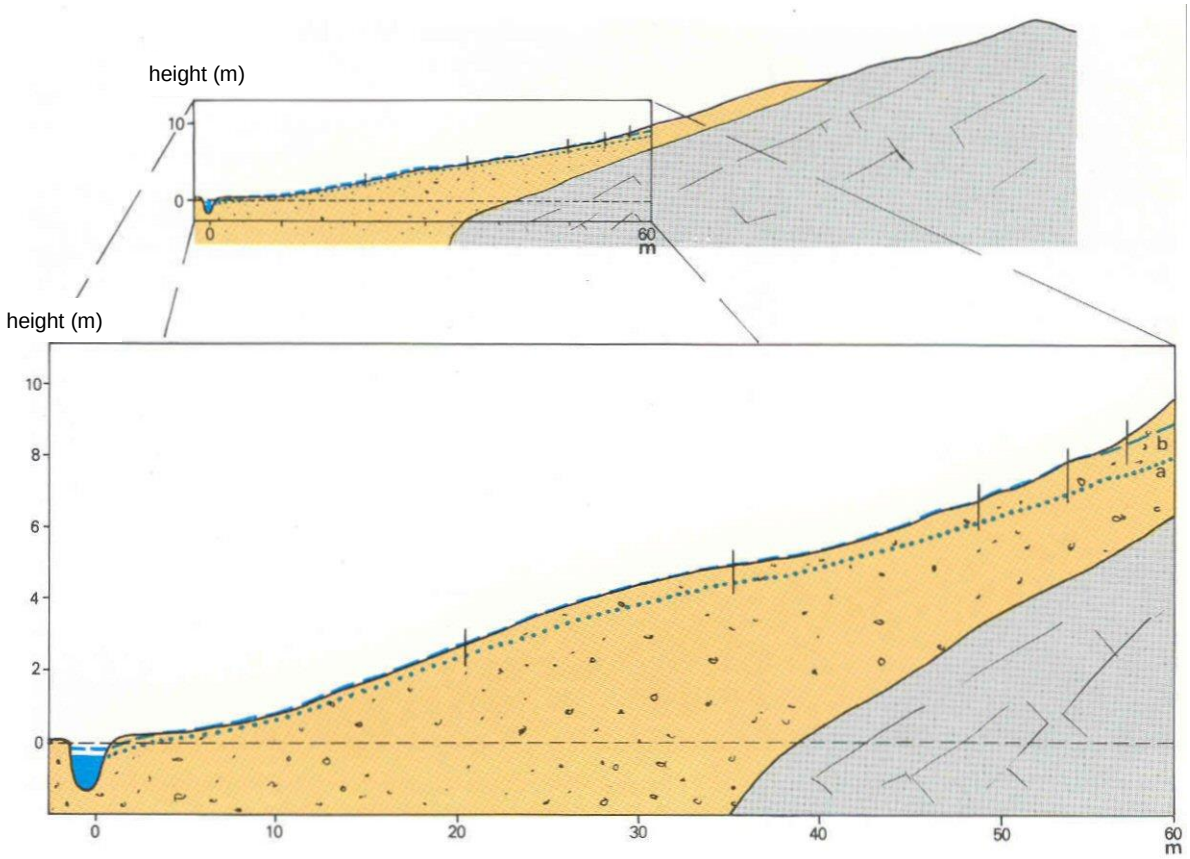
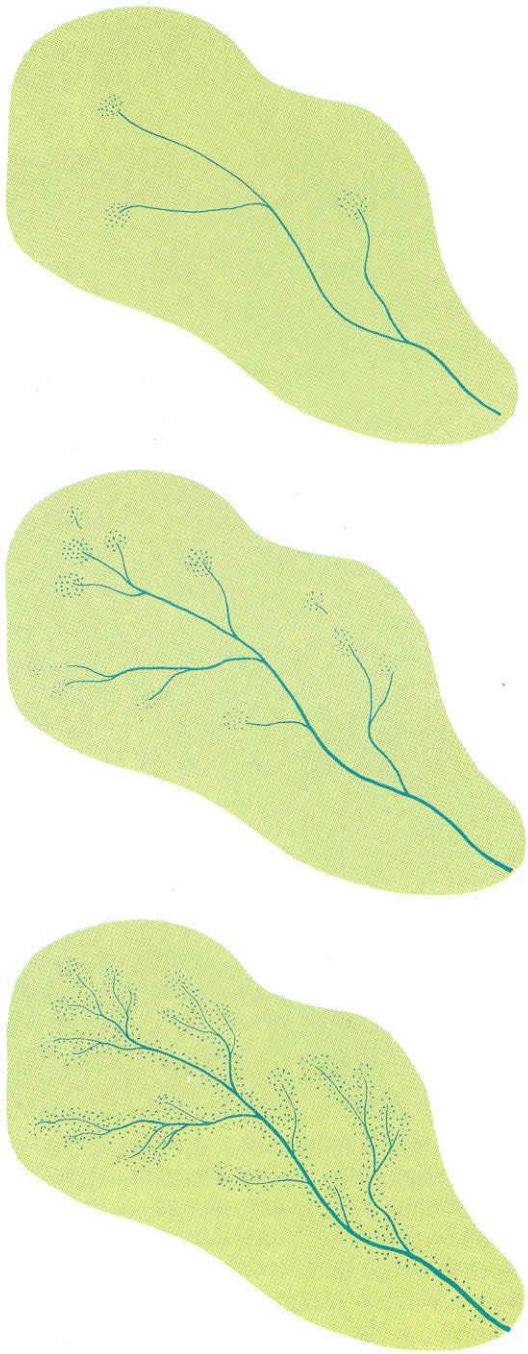


*Fig. 39 Groundwater level along a hillslope at Lagga, Uppland, Central Sweden, at two occasions. Curve **a** is from October 1977, and curve **b** is from March 1978.*

Normally the groundwater surface in this hillslope does not reach the soil surface except in the stream, but with abundant water supply the saturated discharge area can reach up to half of the hillslope during a few days.

The extent of the discharge areas varies over time

The water level observations in Fig. 39 were done in the normal way, that is, in groundwater tubes that were perforated along the whole tube. The water level in the tubes then gives the groundwater level, the slope of which gives groundwater flow in the direction of the slope. The total potential at different depths was not measured and therefore we have no information on the vertical groundwater flow components and cannot determine where the upper limit of the discharge area is. At the lower groundwater level the discharge area, as defined on page 68, may have reached 10 m from the stream and at the higher perhaps 60 m. The saturated discharge area is however easy to identify. During the measurement in the autumn it was only the stream itself and during the snowmelt it reached 55 m up the hillslope. (The stream is actually a dug ditch. Before ditching the soil surface would probably have been saturated some distance from the stream even at the lower groundwater level.) The extent of a saturated discharge area can thus vary considerably over time. At frequent visits to this and other nearby hillslopes rapid changes in the extent of discharge areas were observed. During dry and semi-dry periods the soil surface was saturated only in the streams and in some fens. When the groundwater level rose in connection with snowmelt, intermediate and large autumn rains, or at some large rainstorms during summer, the limit for saturated discharge area moved rapidly up the hillslopes. When the water input ceased the limit moved rather rapidly down the hillslopes again and after some days it was only the soil surface closest to the streams that was saturated.



The sizes of saturated discharge areas were determined during one or two spring floods in six Swedish catchments. The catchments were all dominated by coniferous forested till soils and were between 0.03 and 6.6 km^2 in size. Sites where water splashed around your feet, i.e., areas with water on or close to the soil surface were classified as saturated discharge areas. According to these field investigations the saturated discharge areas covered from 10 to 35% of the catchments. The investigations were done when there was high flow in the streams and should therefore show the proportion of discharge areas close to their maximal extent. During dryer periods the proportion of discharge areas is much smaller and in areas without fens it may be only a few percent or less. (See further the results of the ^{18}O-experiments, page 86.)

Fig. 40 When the groundwater surface rises the areal extent of the discharge areas increase. The risen groundwater level also results in an extended stream network.

The water surfaces of the streams form a part of the groundwater surface and therefore the extent of the drainage network varies with the groundwater level. At high groundwater level the drainage network can be considerably larger than what is shown on a topographical map. After a long lasting dry spell the stream channel may be dry far down in the catchment.

The role of topography

The extent of the discharge areas is of interest for the runoff generation in many ways. First their variation in time at one location is a measure of groundwater discharge. They are also a measure of the catchment's ability to rapidly form runoff after a precipitation input. A large part of the precipitation will fast reach the watercourses as saturated overland flow, that is, without passing the soil profile if the saturated discharge areas are large. When the saturated areas are large there are also large areas around them with shallow groundwater and high soil water content. Infiltration in such areas rapidly results in increased groundwater discharge and increased runoff in the watercourses.

In the discussion about the temporal variation in groundwater levels (page 63) it was noticed that the groundwater level is usually deep in glacio-fluvial eskers. Let us here shortly discuss how the geology influences the depth to the groundwater surface and the extent of the saturated discharge areas. Given a certain groundwater recharge the depth to the groundwater surface is determined by the ability of the soil and bedrock to drain the recharged groundwater. The larger the hydraulic conductivity is, the lower will the slope of the groundwater surface be, and the less thick groundwater zone is needed to drain the recharged groundwater. The groundwater surface therefore has a tendency to be deep and the saturated discharge areas to be small in coarse textured soils. However, for a geologic formation to have a deep groundwater level it is not enough to have a high hydraulic conductivity. In order to have a deep unsaturated zone a thick soil layer is also needed above the threshold level determining the groundwater level in the aquifer. Even this condition is often fulfilled for glacio-fluvial eskers, and the groundwater level is commonly found at depths of tens of metres. In thin sand deposits, as well as in deep sand deposits where the groundwater is dammed by a rock threshold or by a lake, shallow groundwater and extended saturated discharge areas may be found.

We will now give some hints on how information on the topography of the ground surface can be used to estimate possible discharge areas and areas with high soil water content in a till soil landscape.

As a result of the comparatively low hydraulic conductivity of the till soils and the rock below it is, as we have stated earlier, that the groundwater surface largely follows the ground surface. The groundwater is often found at some metres depth even at hills with heights of tens or hundreds of metres. In smaller scales it is found that the topography has a large influence on the depth to the groundwater surface and on the soil water content.

Largest discharge area in a concave hillslope

At equilibrium with a certain water input from above the limit for the saturated discharge area can be estimated by Darcy's law with Dupuit's assumption. Look at the hillslopes in Fig. 41. The groundwater surface is forced to the soil surface where the accumulated water input to the

groundwater above the point starts to exceed the largest possible groundwater flow, that is, when the groundwater surface is at the soil surface. If the hydraulic conductivity and depth of the water transmitting layer are known, the maximum total groundwater flow of the hillslope can be calculated. When the groundwater surface reaches the soil surface the slope of the groundwater surface equals the slope of the ground surface and it is therefore known. For the straight hillslope the calculation is simple because the slope of the soil surface, and therefore the maximum groundwater flow, is constant (see example on page 120). Given a certain slope of the ground surface, the upper limit of the discharge area depends on the groundwater recharge, the conductivity and the total depth of the soil layer. In the cases of the convex and concave hillslopes, the calculations are somewhat more complicated, because the slope of the ground surface changes along the hillslopes. Starting from a chosen position of the discharge area limit in the straight hillslope, that is, a chosen combination of groundwater recharge, hydraulic conductivity and soil depth, the position of the limit for the convex and concave hillslopes have been calculated and marked.

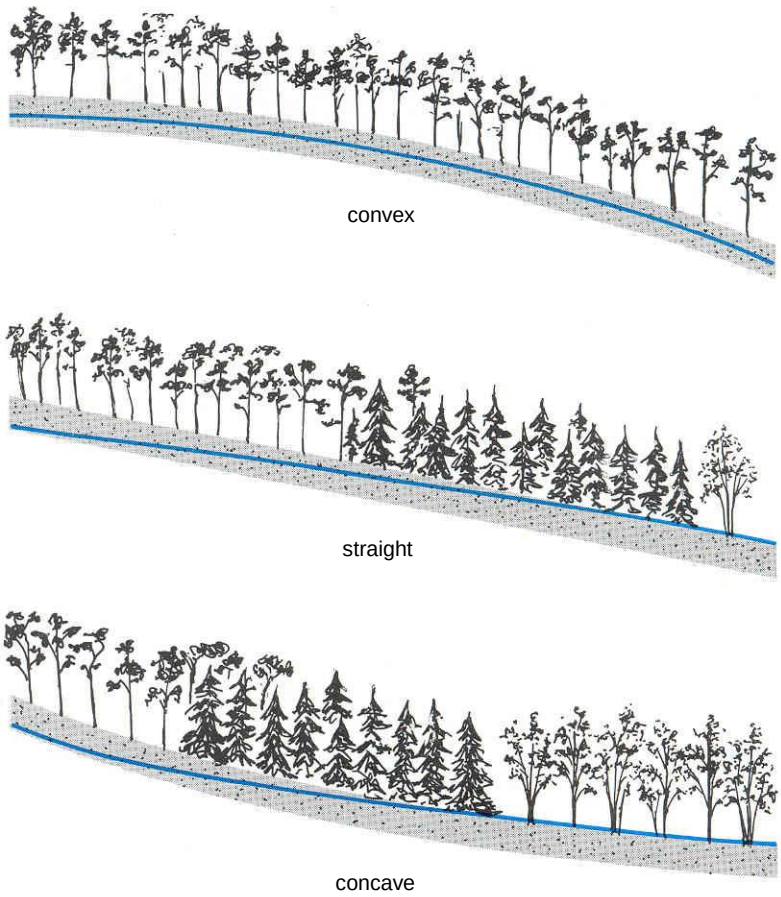


Fig. 41 If the whole soil layer is used for groundwater flow the slope of the groundwater surface cannot be larger than that of the soil surface. In the convex hillslope the slope of the ground surface increases down the hillslope. The soil profile can therefore transmit the gradually increasing groundwater flow from uphill. In the concave hillslope, where the slope decreases down the hillslope, a point is soon reached where the accumulated groundwater recharge uphill is larger than the maximum possible groundwater flow. The groundwater surface is forced to the soil surface and the saturated discharge area will be large.

In the concave hillslope the groundwater flow capacity of the soil decreases downhill because the slope of the soil surface, and thus of the groundwater surface at saturation, decreases. Therefore a large saturated discharge area is formed. In the convex hillslope the condition is the opposite. The slope and therefore the capacity increases downhill. The soil surface in the convex hillslope remains unsaturated.

It should be noted that we have made significant simplifications of the conditions for groundwater flow in these and the following examples to elucidate single effects of the topography. In real hillslopes of different forms, geology and soil conditions vary and the final result may be different.

Hollows wettest – react most rapidly on precipitation

So far we have discussed groundwater flow in two dimensions, the vertical and the horizontal along the hillslopes. Let us now look at the problem in three dimensions and let the form of the slope vary across the hillslope. The effects of topography will then be even more obvious.

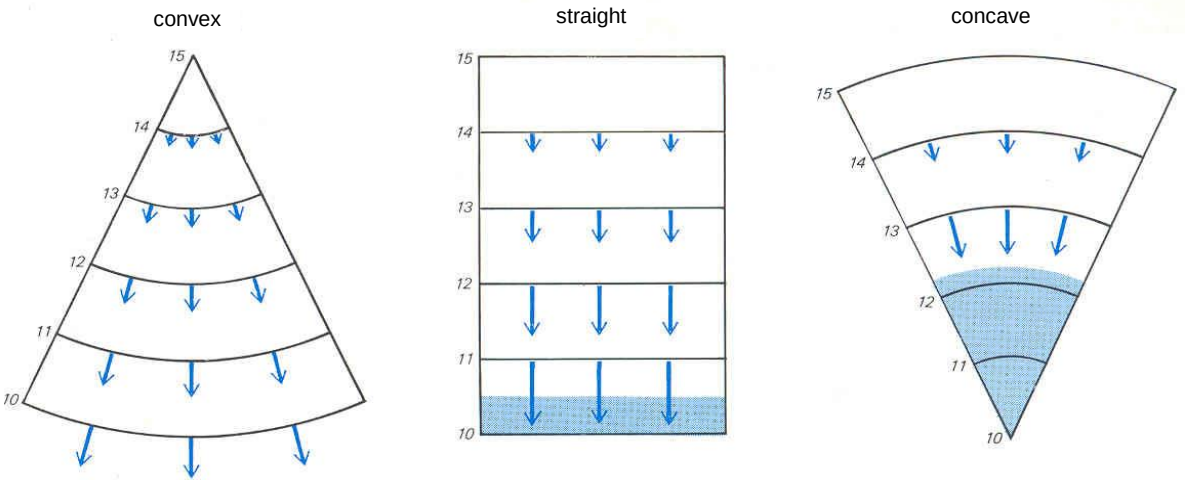


Fig. 42 The hillslopes on these sketched maps have the same constant slope but different forms. In the concave hillslope the water flow is concentrated. Soon the soil profile cannot transmit all water that is added from upslope and the soil surface is saturated (blue). The demand on groundwater flow per unit width will never be this large in the convex hillslope, where an ever increasing width can be utilised. Even far down in such a hillslope the soil profile can transmit all water from above and the soil surface remains unsaturated.

In the same way as in the last example the differences between the hillslopes in Fig. 42 may be shown by simple examples. Let us assume that the hillslopes all have the same constant slope, soil depth, and saturated hydraulic conductivity. Then there exists a maximum possible flow per unit width that the three hillslopes can transmit. At equilibrium conditions the limit of the saturated discharge area will be where the groundwater flow over the whole width of the hillslope is equal to the water recharge above it. It is now only a geometrical problem to find where the limit is.

In the figure we have chosen the position of the limit in the straight hillslope. With the same conditions the limit will be found much higher in the hillslope in the concave hillslope while the convex hillslope will not get saturated at all.

To simplify the calculations in Fig. 42 hillslopes with constant slopes, that is, with the same distance between contour lines, were chosen. The difference between the hillslopes would have been even larger if we would let the concave hillslope be concave also along the hillslope, and the convex hillslope be convex also along the hillslope. This would mean that we compare a hollow with a ridge (cf. Fig. 43). In the hollow a saturated discharge area will soon develop. Water is flowing to this location both as saturated and unsaturated flow. By this water recharge, which continues long after the rain has ceased, the saturated area will remain for a long time after the rain.

In an English investigation it was possible from precise runoff measurements along a stream to show how the flow in a stream during a rain event increased where hollows reached the stream, even though the soil was not saturated to the ground surface. Below ridges the flow in the stream did not change significantly. The hollows and the ridges were weakly developed in this investigation, but they still showed a clear influence on the groundwater flow.

For a considerable flow to take place in the soil it must be saturated or nearly saturated. The hollows may be seen as troughs that concentrate the water. By this concentration the hydraulic conductivity of the soil increases and a large lateral flow can take place.

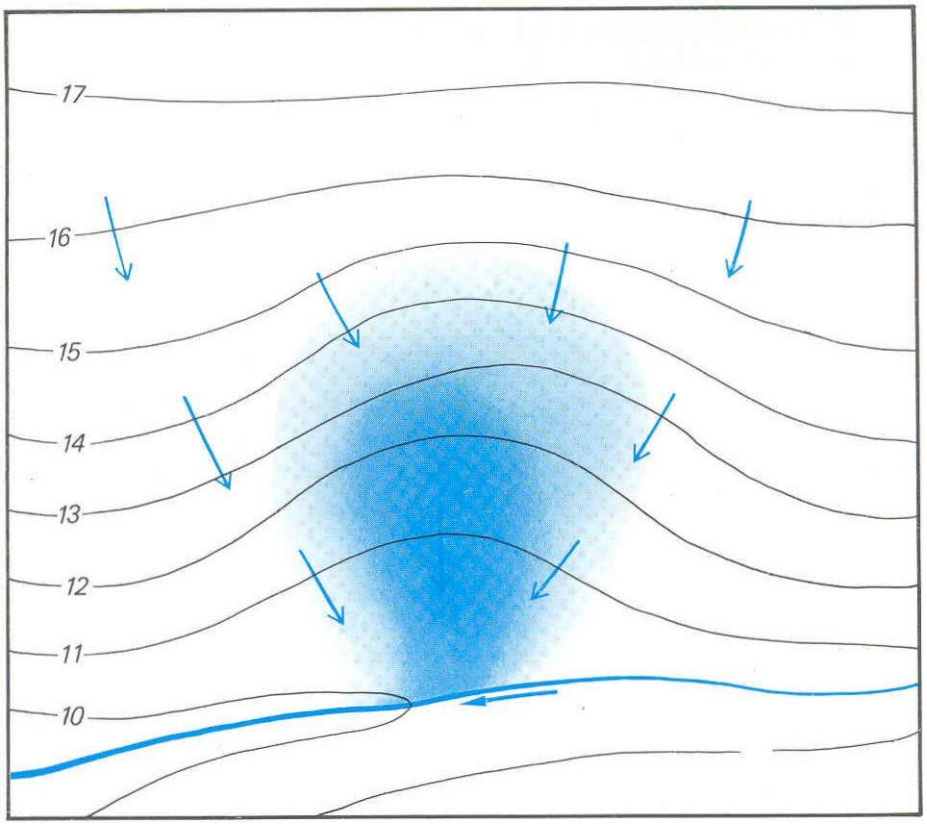


Fig. 43 Discharge areas are abundant in hollows. There the effects that were shown in Figs. 41 and 42 are combined. The slope of the ground surface decreases down the hillslope, and the flow is also concentrated from a large part of the hillslope. The groundwater surface will be shallow and the soil moisture content high.

The transect in Fig. 39 is, at least in its lowest parts, a weakly developed hollow. If we look at the situation about ten metres aside of the transect, the groundwater surface is found at a larger depth below the ground surface and the saturated area does not reach as far up the hillslope as in the transect shown in the figure.

Mires – discharge- or recharge areas?

Mires are wet areas where dead organic material is accumulated and peat is formed. Depending primarily on its position in the landscape and the way in which the water recharge takes place different types of mires form with different characteristic vegetation. *Fens* get, in addition to precipitation, its water from the surrounding mineral soils. *Bogs* get their water from precipitation only.

Many fens have an almost horizontal surface. These fens were formed when residues from vegetation filled lakes, bays and water courses, or when hollows with shallow groundwater become waterlogged. In an undulating landscape sloping fens may be found. In these the soil water and groundwater are less stagnant than in other mire types. The sloping fens are more frequent and occupy steeper slopes the larger the net precipitation is. So called "hanging mires" (literal translation from Swedish) in the mountains are examples of such fens.

Bogs are former fens that have grown in height, and the contact with the groundwater in the surrounding mineral soil has been lost. They are frequently formed on flat land, often on water divides. When they grow the groundwater surface in the bog is also increasing above the surrounding groundwater surface. Especially well separated from the surrounding mineral soil water is the water in the raised bogs. These bogs have a parabolic form and they have a drainage zone along their edges. This drainage zone is a fen because it also receives and drains mineral water from the surroundings. It is common with different mixed forms between bogs and fens. Groundwater from the surrounding mineral soil then only influences some parts of the mire complex.

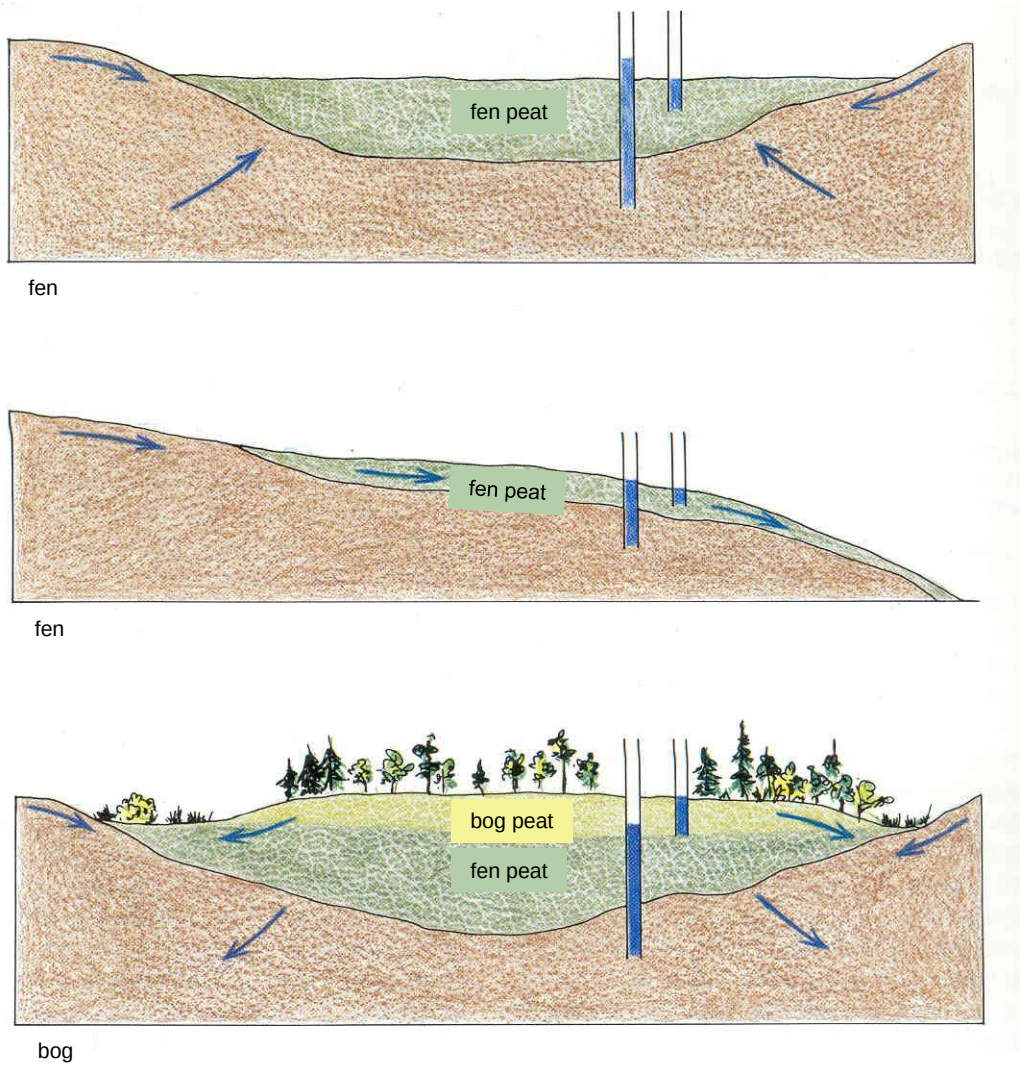


Fig. 44 The fen gets its water from the surrounding mineral soil and also from the precipitation. The bogs, on the other hand, get their water only from the precipitation. Fens are discharge areas and bogs are recharge areas. This can be shown by the pressure differences between the water in the mires and in the mineral soil below them.

In a fen the pressure surfaces of the mire and the underlying aquifer coincides, or is the pressure surface of the underlying aquifer higher than that of the fen. A fen is therefore a discharge area for groundwater. In a bog the circumstances are the opposite. The pressure surface of the bog aquifer is higher than that of the underlying aquifer. A bog is therefore a recharge area (Fig. 44).

The physical properties of the peat in mire can vary considerably, especially vertically. This is because the peat forming vegetation may have varied during the growth of the mire and the

decomposition of the peat, i.e., the humification, usually is more developed in the deeper parts of the mire than in the shallower.

The dry bulk density of peat (page 18) increases with degree of humification. Peat may be classified by its degree of humification. With few exceptions the dry bulk density of peat is between 40 and 200 kg m^{-3}. (For mineral soils it is normally between 1100 and 2000 kg/m^3.)

The porosity of peat is very high. In hardly decomposed peat it may be 97% and in highly decomposed peat 85%. The porosity decreases when the peat is decomposed. At the same time a shift in pore size distribution takes place from large to smaller and smaller pores. A result of this is that the relative proportion of water that is hold with greater tension than the wilting point increases from about 5% to about 20%. The water content at field capacity in hardly decomposed peat is about 30%, but more than 60% in well decomposed peat. The plant available storage is largest and the storage coefficient smallest in the highly decomposed peat.

The hydraulic conductivity of peat decreases rapidly with the degree of humification. In Finnish peat soils values from $1.1 \cdot 10^{-4}$ to $2.0 \cdot 10^{-8}$ m/s were found, which should also be valid for Swedish conditions. Generally the conductivity is lower in *Sphagnum* peat than in other peat forms of the same humification degree.

In peat a lot of water (20 – 30 percent by volume) is held in dead pant cells and in other more or less closed pores. The adsorptively held water may be 5 – 10 percent by volume. In total it is therefore 25 – 40 percent by volume of water that to a very small degree takes part in the water flow in mires. In the same way as in till soils the turn-over-rate for water is long in mires. The residence time for the discharging water is on the contrary rather short.

6. Runoff generation

In the two previous chapters we have treated the conditions for runoff generation in watercourses. The water input to the soil surface, the question of infiltration or overland flow, the groundwater recharge, the groundwater flow pathways and the existence of discharge areas. In this chapter we will discuss the runoff generation starting from the discharge in the streams. We will often leave the stream and look at the surrounding hillslopes to return to the former discussions about water storage and flow.

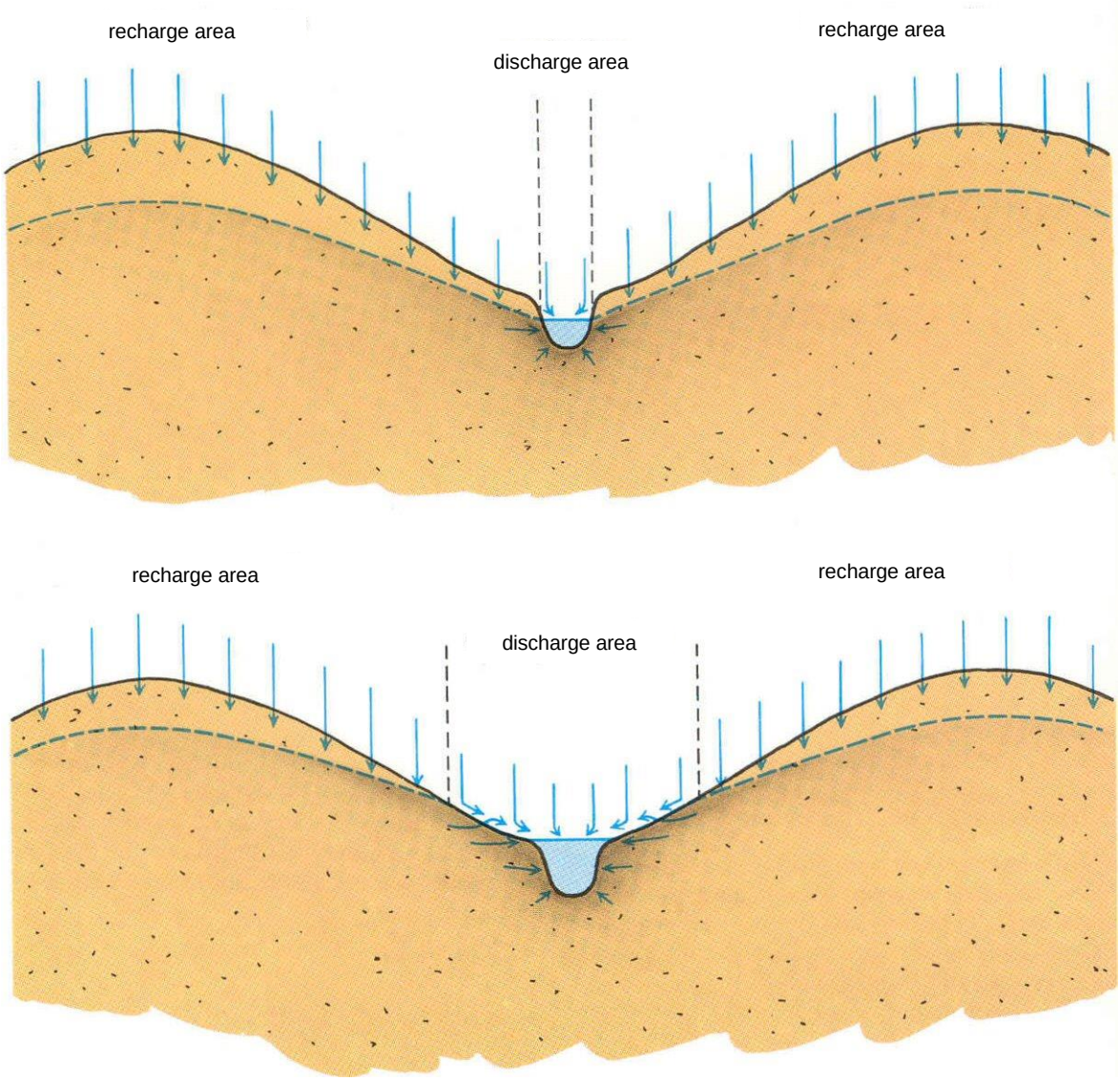


Fig. 45 When the precipitation on recharge areas infiltrates, the groundwater level increases and the discharge in the discharge area increases. The reason for this is that the slope of the groundwater surface increases and layers with high conductivity start to contribute to the groundwater discharge.

In the figure the size of the groundwater flow is marked by the intensity of the grey color. Precipitation on saturated discharge areas cannot infiltrate but forms saturated surface discharge together with the discharging groundwater. Normally the groundwater discharge dominates over the rainwater flow in the stream.

The runoff is described by a hydrograph

A diagram in which stream discharge is plotted against time is called a *hydrograph*. A hydrograph from a single rain event has a characteristic form. It starts with a relatively fast flow increase and ends with a slow decline. We will refer to these parts of the hydrograph as the *rising limb* and *recession phase*, respectively, and between them the *peak value*. The total response of a stream to a rain- or snowmelt event, i.e., a rising limb, peak value, and recession phase constitutes a *flow event*.

Given a certain water input, the rising limb and the peak value of the hydrograph at a point along a stream depend on the position of the groundwater surface and the soil moisture content before the water input. The recession phase describes, on the other hand, almost the same curve after all peak values, with rapid recession at high flow rates and successively slower recession as the flow decreases (cf. the flow events in Fig. 5).

The peak value is normally reached somewhat after the water input has stopped or declined substantially in intensity. This time lag may be a few hours in a small stream. The larger the catchment is, the larger the delay. Expressed per area unit, the peak value decreases with increasing catchment area, other conditions being similar. The flow peaks are more spread in time because the water inputs from the different parts of the catchment reach the observation point with increasing time span. Water storage in for example lakes or flooded areas has similar influence as the catchment size, i.e., it decreases the peak value and makes the flow event more extended in time.

The origin of streamwater can be determined by tracers

The water in a stream could be seen as a mixture of two types of water, groundwater and fresh rain- or meltwater. If the concentration of a suitable tracer in groundwater and rain or meltwater is known, it is possible to calculate the relative contributions from the two sources in the streamwater.

Let c_g, c_p and c_b be the concentrations of the substance in groundwater, rainwater (including meltwater) and streamwater. Further, let the discharge of the stream be Q and let the fraction of groundwater in the stream be X. Mass balance of the substance, i.e., that the input of the substance from groundwater and rainwater is equal to the transport of the substance in the stream, gives

$$X \cdot Q \cdot c_g + (1 - X) \cdot Q \cdot c_p = Q \cdot c_s$$

This gives

$$X = (c_s - c_p) / (c_g - c_p)$$

X = fraction of groundwater
C_s = tracer concentration in streamwater
c_p = tracer concentration in rain- or meltwater
c_g = tracer concentration in groundwater

This equation can be applied if the concentration of the tracer in discharging groundwater is fairly uniform and different from that of the precipitation. The stable oxygen isotope ^{18}O (page 57) has been shown to be a useful tracer.

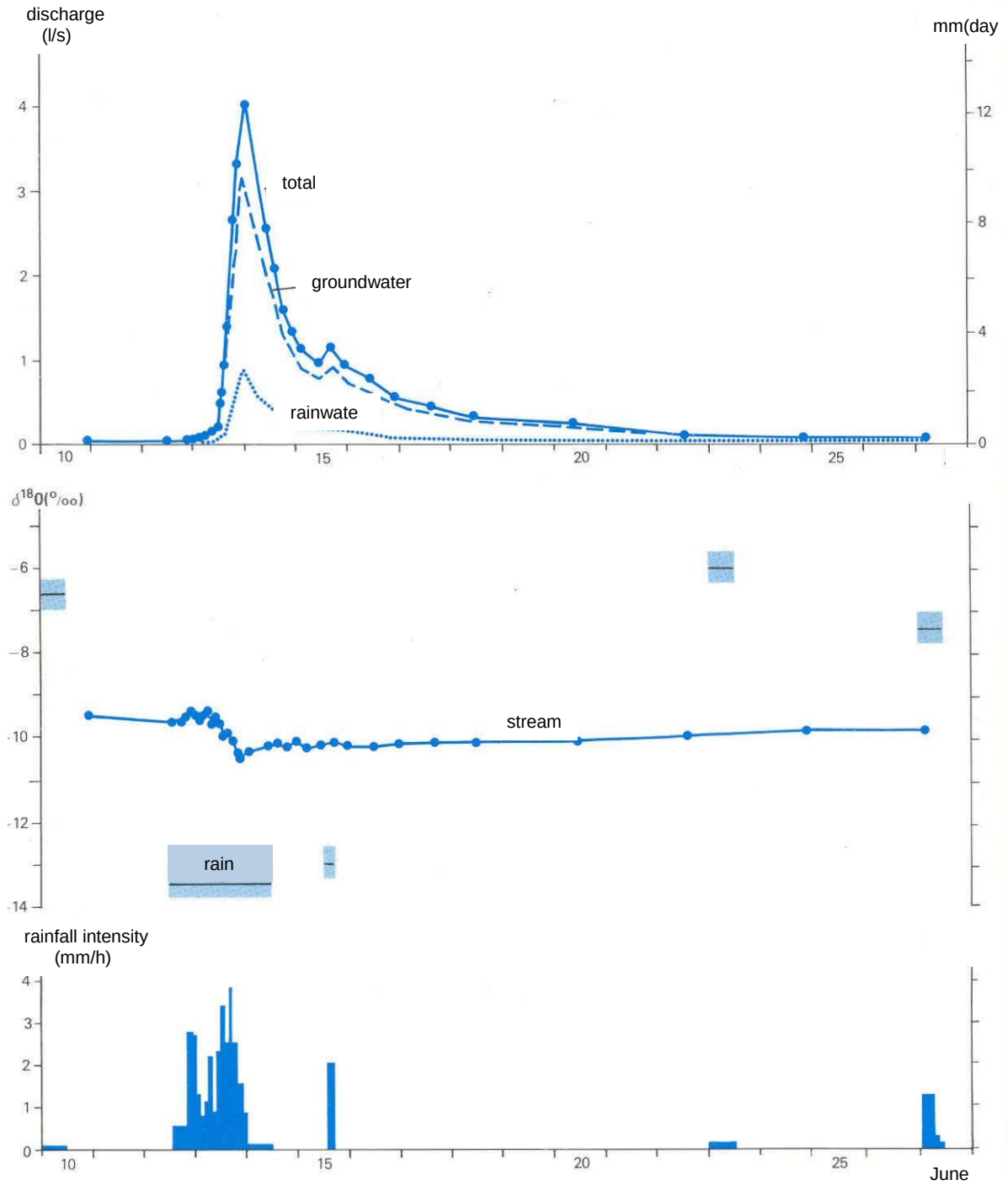


Fig. 46 Hydrograph and calculated contributions from groundwater and rainwater (upper panel), ^{18}O-concentrations in streamwater and rain (middle panel) , and rainfall intensity (bottom panel). Stream F1, Gårdsjö-area, Bohuslän, West Sweden, 1982.

It was assumed that groundwater was the only component in streamwater before rain started. Then it was assumed that streamwater was composed of a mixture of new rainwater and groundwater. From the ^{18}O-concentrations the contribution of respective water to streamwater was calculated. The reason that the ^{18}O-concentration in streamwater changed so little even though the ^{18}O-concentration of the rainwater deviated a lot was that streamwater at the flow event mainly consisted of groundwater.

We have earlier given examples where the seasonal variation in ^{18}O-concentration in precipitation could be found in some soil profiles (Fig. 30). It was concluded that the water particles at one soil level were of about the same age. The groundwater that reaches a well or is discharging in a discharge area is on the contrary composed of water with very different ages because of different transit times in the unsaturated zone and different transit times in the groundwater zone.

Therefore the ^{18}O-signal in rainwater is strongly damped, and the seasonal variation in the ^{18}O-concentration of groundwater is rather small. Thanks to this damping of the seasonal variation, the ^{18}O-concentration of groundwater and precipitation will differ at some precipitation events and the equation can be applied. At other precipitation events the difference is too small for the equation to be applied.

In Fig. 46 calculated contributions from groundwater and rainwater during a flow event in a small stream at the Gårdsjö area, Bohuslän, West Sweden are shown. The area is covered with coniferous forest and has a thin till soil on solid rock. In the calculations, ^{18}O-concentration of the groundwater has been assumed to be equal to that of streamwater before the start of the rain. This assumption is reasonable because a long time had passed since last rain event and surface water storages were probably emptied, so that streamwater was only groundwater. A long lasting frontal rain, with low intensity of about 3 mm per hour, gave in total 56 mm rain. This rain gave rise to a large flow event in the stream. The ^{18}O-concentration of the rain was quite different from that of groundwater and it was possible to apply the equation. Obviously the rainwater had only a small influence on the ^{18}O-concentration of streamwater and it can be concluded that the flow event was dominated by groundwater.

In the calculations a mean value of the ^{18}O-concentration of the rain was used, and it is this mean value that is given in the figure. In reality the ^{18}O-concentration of the rain varied a lot during the rain event. Because of this variation the appearance of the rainwater- and groundwater hydrographs are uncertain. The total volumes of rainwater and groundwater in the stream are still secure in this case.

The total runoff during the flow event was 17 mm. Out of these 82%, i.e., 14 mm was groundwater. The error in the calculated groundwater contribution was estimated to ± 10 percentage units.

From the discharged volume of rainwater the size of the saturated discharge areas can be estimated. We then assume that all rainwater that appears in the stream originates from rain that had fallen on saturated discharge areas (inclusive the stream itself), and that no other rainwater had reached the stream (cf. Fig. 45). The total rainwater volume that was discharged during the flow event is then equal to the amount of rain that fell on saturated discharge areas and we get

$$V_p = P \cdot A_u$$

That is

$$A_u = V_p / P$$

A_u = saturated discharge area (m^2)
V_p = discharged volume new rainwater (m^3)
P = rainfall or snowmelt amount (given in m)

With this equation the saturated discharge area at the flow event in Fig. 46 could be estimated to about 4% of the catchment area.

After a few weeks without larger rainfall events a rainstorm passed. The rainfall intensity was higher, up to 15 mm per hour, but the total rainfall was smaller, only 17 mm. The flow event was smaller and in the same way as the former event it was dominated by groundwater, now constituting 96% of the discharged volume. The saturated discharge area was calculated to only 0.4%, which is about the area of the stream.

If the flow events in streams are mainly generated by groundwater discharge, there should be a good relation between groundwater level and runoff. Such relations for two streams in the Gårdsjö area are shown in Fig. 47. The data are from one summer and autumn period with a number of flow events in the streams. Both observational wells for the groundwater level were situated close to discharge areas. Far up in the recharge areas the relation is less good. The bending curves could be interpreted as the result of increasing hydraulic conductivity of the soil profile closer to the soil surface. A given increase in groundwater level at high groundwater level, results in a larger increase in runoff than the same increase in groundwater level at low groundwater levels. The relation does not show that the flow events are dominated by groundwater, but it is a prerequisite.

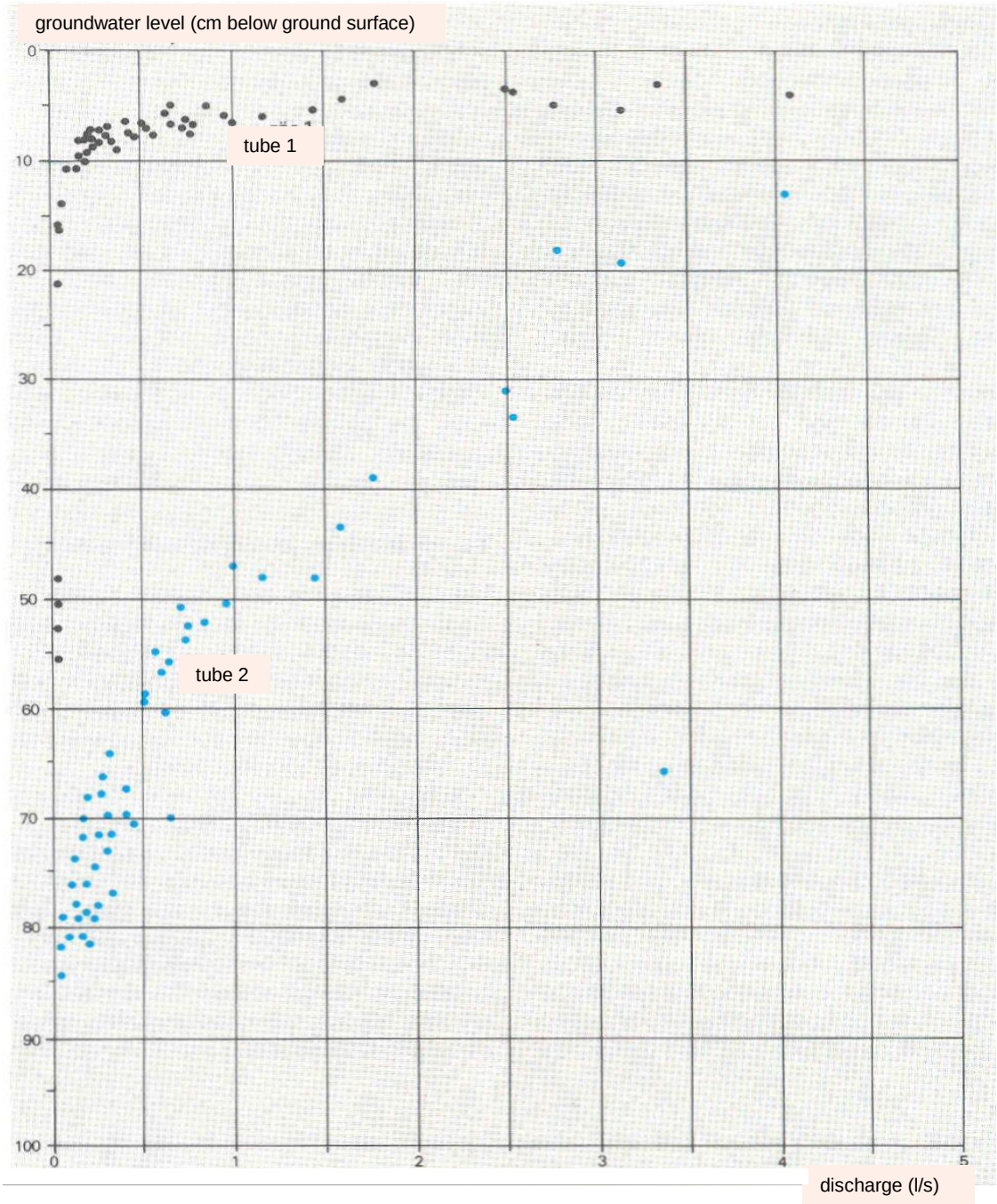


Fig. 47 The groundwater level given as a function of stream discharge. Stream F1, Gårdsjö area, Bohuslän, Sweden.

Many Swedish catchments have rock outcrops in their upper parts. The Gårdsjö stream discharges a typical area in Bohuslän, with outcrops on the hills and soil in the valleys. The high proportion of groundwater in streamwater may be amazing considering that no infiltration takes place on the outcrops. The explanation is that the stream is flowing on the soil deposit and has no contact with the outcrops. The Hortonian overland flow that is formed on the outcrops infiltrates in the cracks and in the soil below the outcrops and contributes to the groundwater runoff in the stream. In this way the outcrops concentrate the precipitation to the lower, soil covered parts of the area. This favours groundwater recharge and groundwater discharge there. The ^{18}O-investigations show that the transmissivity of the soil surrounding the stream is large enough for the groundwater discharge to dominate the flow events. They also show that the size of the soil- and groundwater storages is so large that the water only to a small degree is exchanged by the water that infiltrates during a single flow event.

Groundwater dominates in many streams

The investigation of the flow events in the Gårdsjö area is a part of a larger investigation in which the origin of streamwater at rainfall and snowmelt events is studied by ^{18}O in one or a couple of small catchments at six locations in Sweden. All areas are dominated by coniferous forested till soils on solid rock.

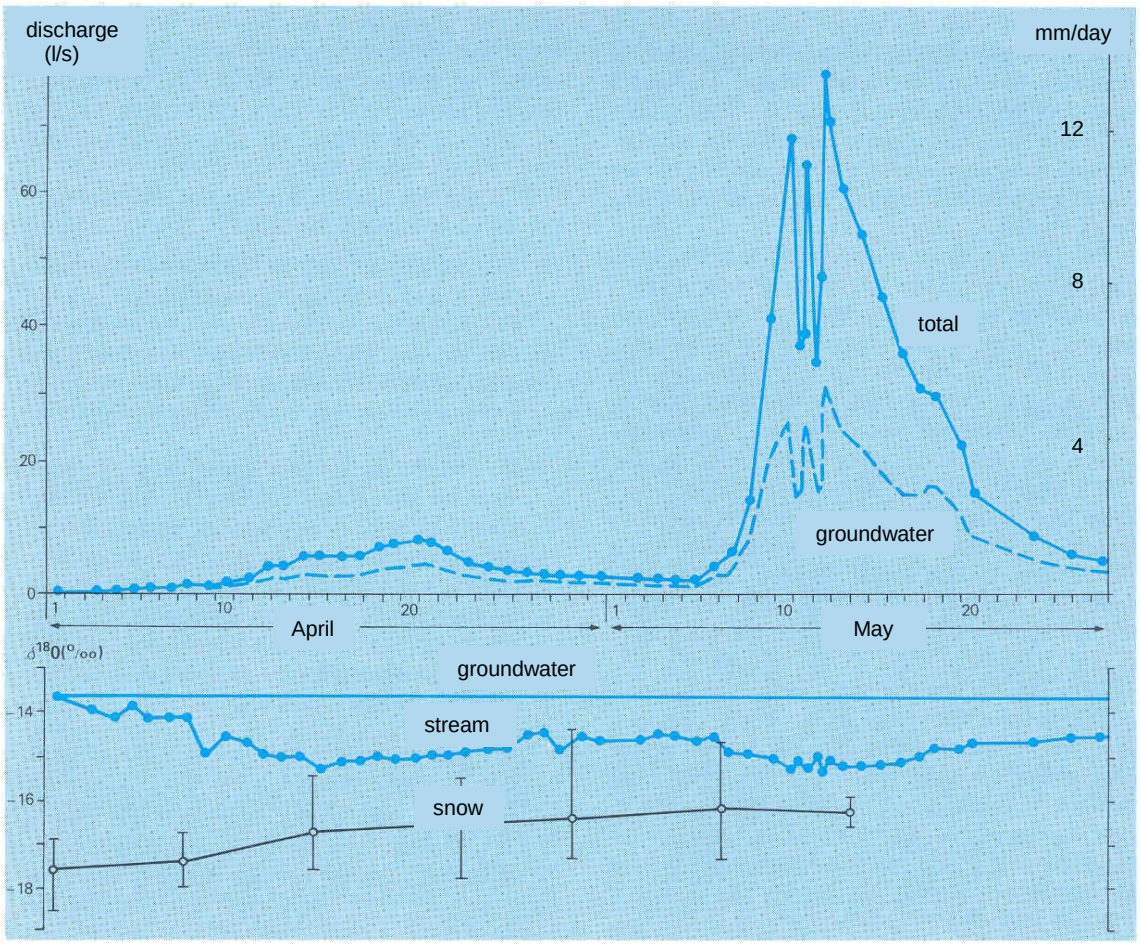


Fig. 48 Hydrograph and calculated contribution from groundwater, and ^{18}O-concentrations in streamwater and snow pack. Svartberget, Vindeln, Västerbotten, Northern Sweden, 1981.

The ^{18}O-concentration of the snow pack was measured once per week at ten locations in the catchment. The mean value, maximum and, minimum ^{18}O-concentrations from these measurements are given in the diagram. About half of the spring flood was contribution from groundwater.

The result from a flow event after rain at Stormyra (Södermanland, Southern Sweden) is remarkable (not shown here). A thunderstorm in June 1980 gave 47 mm rain during a little more than one hour. The streamflow increased rapidly to the level of a large spring flood. (The rainstorm resulted in flooding in many places in the area.) Water sampling at the peak value of the flood showed that the proportion of groundwater then was 60%. The proportion of groundwater in the total flow event was about 70%.

A spring flood at Svartberget, Vindeln (see map of the catchment in Fig.1) is shown in Fig. 48. The snow water storage was 200 mm when the snowmelt started and the precipitation during the snow-melt period was 11 mm. According to the mass budget calculations for ^{18}O, 47% the total runoff during the two peaks during this spring was groundwater. The proportion of saturated discharge area during the two peaks was calculated to 14% and 28%, respectively, of the catchment area. Field taxation soon after the spring flood culmination gave about 25% saturated discharge area. The relatively large meltwater runoff was explained by snowmelt on the rather large saturated discharge areas.

Simultaneous investigations in the western tributary (see map in Fig. 1) resulted in a larger groundwater fraction of 64% of the total flow. The difference was probably due to the influence of the mire. The high groundwater level in the mire soon constitutes a large saturated discharge area contributing to a large meltwater runoff in the eastern stream and also to the total flow from the catchment.

The sampling for the experiment shown in Fig. 48 was normally done once per day. The discharge that is shown in the figure is based on momentary flow values at the time of water sampling. During a couple of days when runoff peaked, samples were taken a couple of times per day. The continuous hydrograph would have shown a large daily variation, which is common during snowmelt periods in small streams. The largest meltwater recharge to the area coincides with the highest snowmelt intensity in the afternoon. The time lag in the snow pack, soil and watercourse leads to a peak flow in the stream in the evening, night or in the following morning, with increasing time lag the larger the catchment area. An interesting observation is that the ^{18}O-concentration has only a small daily variation. This fact indicates that the meltwater and groundwater contribute to the streamflow in about the same proportions during the day, even when the daily variation in runoff is large. When the meltwater input increases during the day, the meltwater runoff from saturated discharge areas and the groundwater discharge increase in the same proportions.

The interpretation of the ^{18}O-concentration in streamwater that was shown in Fig. 46 and 48 requires that the ^{18}O-concentration of the groundwater is constant during the whole flow event. Actually, the ^{18}O-concentration of the discharging groundwater changes a little because of the input of new rain or meltwater that generate the groundwater discharge. The calculated groundwater proportions are therefore underestimates. They show the proportion of "old" water in streamflow, i.e. , water that was groundwater or soil water before the flow event started. Some of the water that was named rain or meltwater in the streamflow is groundwater that was formed by water particles that infiltrated during the flow event. The error is largest at large water input. The error was estimated to about 6% of the total discharge during the spring flood at Svartberget. Correcting for this error would give a groundwater fraction of 53 instead of 47%.

The following conclusions were drawn from the investigations on the origin of water in streams:

- All flow events generated by rain, and most of those generated by snowmelt, were dominated by groundwater.
- Field taxations of saturated discharge areas were done at seven spring floods. The results from these coincided, with one exception, surprisingly well with the fractions calculated from ^{18}O-concentrations.
- When all flow events in connection with snowmelt were compared, the groundwater fraction of the total discharged volume was least at the largest snowmelt events.
- No connection was found between rainfall intensity and fraction of groundwater in streamflow. If the flow events were generated by rain intensity exceeding the infiltration capacity of the soil the fraction of groundwater should have decreased with increased intensity.

The origin of streamwater can also be studied using the chemical composition of water. The electric conductivity, i.e., the salt content, is much easier to measure than ^{18}O. As we have already shown, the electric conductivity in groundwater is higher than in precipitation. A common observation is that the electric conductivity of streamwater decreases when the flow increases, which indicates increased contribution from rainwater to the stream. The decrease is however hard to interpret with the simple model that is explained with the equation on page 81, because the electric conductivity of the discharging groundwater has large variations due to the different flow pathways and transit times of this water (cf. Fig. 37). The contribution from the shallow groundwater with relatively low salt content increases during the flow event, and therefore it seems as if streamwater contains more rainwater than it actually does.

Table 3. Summary results from investigations of the origin of streamwater with ^{18}O. (After Rodhe, 1984.)

		Groundwater fraction (%)			Fraction of saturated discharge area by ^{18}O (%)		
	n	Max	Min	Median	Max	Min	Median
Rain events	15	100	68	85	17	0	2
Snowmelt events	24	91	32	64	63[b)]	2	23
Whole spring floods[a)]	16	86	41	59	49[b)]	9	26

a) The spring floods were often composed of a number of peaks (cf. Fig. 46). These are given separately under "Snowmelt events".
b) This value, from Aspåsen, Hälsingland, is considerably larger than the fraction observed in the field.

A mathematical model

The Swedish investigations with ^{18}O show that the flow events are often dominated by groundwater. This has also been shown by investigations with ^{18}O and other natural isotopes in water, in for example Germany and Canada. The results support the view with recharge- and discharge areas for runoff generation. In earlier chapters we have in some detail discussed prerequisites for this view. It still remains to discuss the mechanism of groundwater discharge. How is it possible that the groundwater discharge can respond so quickly to a water input to a catchment? Where in the catchment does the infiltration take place that generates the

groundwater discharge? One way to elucidate these questions is to calculate the flow during some selected situations with a mathematical model.

Using Darcy's Law it is possible to estimate how the position of the groundwater table and the groundwater discharge for a given slope react on infiltration from different rainfall events. Because it is not a matter of equilibrium conditions (as in for example Fig. 41), but rather of the variation of flow with time, the calculations are rather complicated. A mathematical model, i.e., a series of equations that are solved stepwise, may be constructed where the pressure and flow of soil- and groundwater are calculated in a large number of points in the hillslope with small time intervals. The calculations require that the saturated hydraulic conductivity and soil water retention curve are known for the different parts of the hillslope.

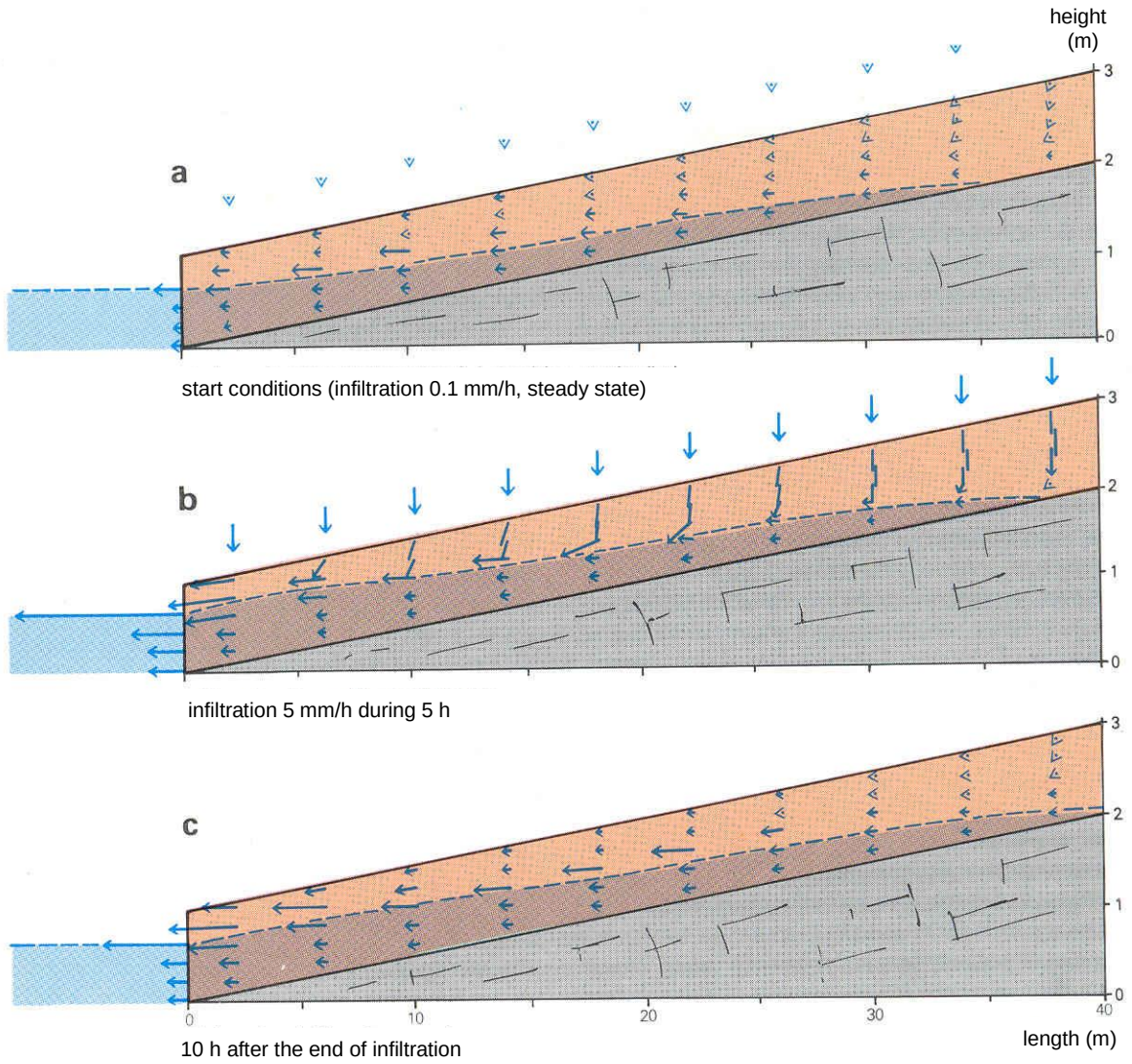


Fig. 49 Soil- and groundwater flow in a layered hillslope calculated by means of a mathematical model. To give the figure a reasonable size the vertical scale is exaggerated (as in many other figures in this book). The arrows that indicate the size and direction of flow are drawn in the same vertical and horizontal scale. The longest arrow denotes a flow of $8.7 \cdot 10^{-6}$ $m^3/(s\ m^2)$ = 32 litres / (m^2 hour). At small flows the direction is given as an arrow around a point. In places where the arrows overlays each other the arrowheads are omitted (the vertical flow in fig. b).

In Fig. 49 soil- and groundwater flow before, during and after an infiltration event has been calculated. The water flow is assumed to take place in a one metre thick layer on an impermeable

bed (the deep flow in Fig. 33 is disregarded). Based on measured values of the saturated hydraulic conductivity in till soils, the conductivity was chosen to decrease from $2.4 \cdot 10^{-4}$ m/s in the surface soil to $1.3 \cdot 10^{-5}$ m/s at the bottom of the profile.

The response of the water flow to precipitation depends not only on static properties as the form of the hillslope and hydraulic properties, but also on dynamic properties as the soil moisture distribution and the position of the groundwater table before the rain starts. During these calculations the soil moisture and groundwater conditions before the rain were assumed to be in equilibrium, with an infiltration intensity of 0.07 mm per hour. The soil water content is therefore close to field capacity when the rain starts, and the groundwater surface has a realistic shape. The position of the groundwater surface at the bottom of the hillslope was held constant, mimicking outflow to a stream.

Direction of soil water and groundwater flow

In the saturated zone the water flow is directed downslope. In the unsaturated zone the flow direction varies. When there is no or little infiltration the flow is mainly downslope. When infiltration is large the flow is mainly vertically downward. The difference may be explained from differences in total potential conditions of the soil water. In a case when there is no or only little infiltration, the soil water potential is almost equal to the height above the groundwater table (cf. page 23). Because the groundwater surface is sloping, two points at the same level, but at different distances from the stream, will have different total potential with the lowest total potential closest to the stream. The gradient in total potential is largely parallel with the groundwater surface. When the water input through the soil surface is large, the soil water potential at a given level will be less than the distance to the groundwater level (cf. curve c in Fig. 14) and a substantial gradient in total potential will be directed downwards. The lateral gradient still exists but the vertical gradient will dominate and give a practically vertical unsaturated flow.

In the zone just above the groundwater surface the flow is similar to the groundwater flow, both regarding size and direction. In this zone the hydraulic conductivity is as large as in the saturated zone. This is because all pores are actually water filled at the small suction prevailing just above the groundwater surface. The pressure is larger (less negative) than the pressure that causes the first bend on the water retention curve, which is called the air entry pressure (cf. page 32). From the point of view of water flow, the saturated zone is somewhat thicker than indicated by the position of the groundwater surface. When the groundwater zone is thick this has little influence, but when it is thin as in this hillslope the flow just above the groundwater surface may constitute a considerable part of the total flow. This is especially is the case when there is a large increase in the hydraulic conductivity towards the soil surface.

The response of the groundwater surface and groundwater flow to infiltration

The groundwater level in the model changes in different ways in different parts of the hillslope. At the discharge point, as mentioned, the groundwater level is held constant in the model. A few metres from the stream the groundwater level starts to increase already within one hour after the infiltration started. In the middle of the hillslope the increase of the groundwater level starts after a few hours. In the beginning there is no groundwater at the top of the hillslope. After a few hours the groundwater zone reaches the top of the hillslope and the groundwater level rises also here. When the rain stops the groundwater level immediately sinks close to the stream, but continues to rise during some hours at the top of the hillslope.

Despite the different rates of rising mentioned above, the slope of the groundwater surface remains almost constant, except closest to the stream. The water flow in the hillslope increases because the cross sectional area (height) increases and especially because the added saturated layers have higher conductivity than those below. Closest to the stream the slope of the groundwater surface increases and the flow increases at all levels. In a real stream the water level of course increases when the runoff increases, but in many situations the rise of the stream surface is small compared to rise of the groundwater level. The increase in the slope of the groundwater surface close to the stream, and the increase in groundwater discharge this gives, is therefore realistic.

When the rain has stopped the flow in the soil water zone declines rapidly, because the water content, and therefore the hydraulic conductivity, decreases. The groundwater flow from the upper part of the hillslope continues for a long time and the recession of discharge is slow.

In this example the groundwater surface never reached the soil surface and we never got saturated overland flow from the hillslope. When the same calculations were performed for a concave hillslope a saturated discharge area soon appeared. In that example the rainwater outflow dominated the total runoff from the hillslope. In the discussion about the role of the topography the importance of hollows for generating saturated discharge areas was stressed (cf. Fig. 43). With the type of model that was used in previous example, such hollows cannot be imitated. A model that treats flow in three dimensions is needed. It is therefore not possible for us to calculate the growth of saturated discharge areas in a catchment, even if we would combine hillslopes of different form to create a "catchment". At present we have to leave the appearance of saturated overland flow and only study groundwater discharge.

The groundwater discharge from the ^{18}O-studies can be simulated

The calculated groundwater discharge and its time variation can be compared with the groundwater flow that was measured in streams with ^{18}O-techniques. Suppose that a small catchment consists of a stream surrounded on both sides of the modelled type of hillslope. The groundwater hydrograph from the model with the flow expressed in mm per day may be compared with the groundwater hydrograph from the ^{18}O-studies. (We assume that the catchments are so small that water storage in the drainage net does not change the picture in the time scale we chose). In the presented model example, the groundwater discharge increased from the initial value of 1.7 mm per day to a top value of 11 mm per day within about 6 hours. Despite the higher initial value the increase, the maximum value and the following recession phase is nearly equal to the groundwater hydrograph from the Gårdsjö stream in Fig. 46. No attempts were made to imitate exactly this catchment, or the moisture conditions and the rain that determined the shape of the hydrograph at Gårdsjön. The coincidence is therefore by chance, but the comparison shows that with reasonable hydraulic conductivity, the groundwater discharge can react on infiltration in the same way as the ^{18}O-study showed.

The groundwater surface responds rapidly to infiltration close to the discharge area

In a hillslope with homogenous soil the groundwater surface should react most rapidly at the bottom of the hillslope. Because of the shallow groundwater, the soil water content is highest there and therefore the pressure from the infiltrating water is transmitted with high velocity. In addition the distance to the groundwater surface is short here.

The pressure propagation of course implies a water transport that is needed to fill the larger pores with water. This is the water transport that constitutes the percolation. The water content distribution in the unsaturated zone far down and far up in the hillslope may look as the right and the left curve respectively in the figure with mats (Fig. 12). When the groundwater level rises the curves are displaced upwards. At the deep groundwater surface upslope the soil water content has to increase in a large part of the unsaturated zone until a new equilibrium state is reached. (The increase in water content is equal to the product of storage coefficient and the rise of the groundwater surface.) Downslope, on the other hand, the unsaturated zone is so thin that no layers with low water content exist. The groundwater surface can rise without any substantial increase in water content in the soil water zone, because almost the whole pore volume is already water filled. The storage coefficient is very small in this case.

The rapid pressure propagation and the tendency of the groundwater surface to rise a lot when little water is added act together to make the infiltration generate the fastest and largest rises in groundwater levels close to the discharge areas. Closest to the stream the groundwater level rises only a little because the discharge to the stream keeps the groundwater level down (cf. the model example). This requires that the soil is homogenous, i.e., for example that it has the same water holding properties at all levels. In reality the proportion of large, easily drained pores increases towards the soil surface. This counteracts, but does not take away, the difference described here.

The rapid rise of the groundwater surface close to the discharge area may lead to the formation of a temporary "ridge" in the groundwater surface. From this ridge the water flows towards the discharge area and contributes to the increased discharge in the stream, but also up the hillslope. Such ridges have been observed in soils where the groundwater surface has small slope before the infiltration event. A tendency of a ridge can be seen in the model example (Fig. 49, diagram b), even if the groundwater surface never came to slope "backwards". In other model examples fully developed ridges have appeared.

Infiltration close to the discharge area is decisive for the flow event

In a hillslope of the same type as the one in the model, with relatively thin water conducting layer on impermeable bedrock, the groundwater discharge is determined by the height and slope of the groundwater surface closest to the discharge area. If the water-conducting layer is deeper, the discharge is also determined by the height of the groundwater surface far up in the recharge area (cf. Fig. 33).

As stated on page 72 it is not probable that the deeper groundwater contributes substantially to the increase in flow even if the hydraulic conductivity may be considerable also at great depths.

A conclusion that can be drawn from the model exercises is that the increase in groundwater discharge at infiltration events is related to the infiltration close to the discharge area. It is that infiltration that results in the rise of the groundwater level there, which in turn results in increased groundwater discharge. The infiltration in the upper parts of the hillslope is important to keep the high discharge and for the recession phase.

Similar conclusions can be drawn from other mathematical model experiments where infiltration was admitted along the whole hillslope, as well as in parts of the hillslope, and the the resulting hydrographs were compared. Field experiments have shown that also infiltration in the upper

parts of hillslopes can contribute to the flow increase in the stream. It was possible to irrigate an eight-hectare catchment in USA, either the whole catchment or some chosen parts of it. No Hortonian overland flow and very little saturated overland flow were seen during the experiments that were done with an irrigation intensity of 6 mm per hour during 8 hours. When the soil was "dry" before irrigation, irrigation of the lower third part of the hillslopes gave the same rising limb and peak value in the stream hydrograph as did irrigation of the whole catchment. The streamflow recession that started soon after the end of the irrigation was much faster after the partial irrigation than after irrigation of the whole catchment. When the soil water content on the other hand was high before the irrigation, the partial irrigation of the lower part of the hillslopes led only to the initial streamflow increase as compared with irrigation of the whole catchment. At the later experiment the increase in streamflow continued for two hours after the end of the irrigation, while the partial irrigation immediately was followed by a recession phase. During moist conditions thus also the infiltration in the upper parts of the hillslopes contributed to the rising limb and to the peak value.

From where is the flow event coming?

We will try to answer the above question from two perspectives. First we will discuss how different parts of the catchment contribute to the flow event. Then we discuss which water particles that are discharging, i.e., from where the water in the flow event emanates. When it is stated that a certain area contributes we mean that water input to the area gives impulses to the flow event in the stream, i.e., the water input influences the hydrograph. It should be noted that it is not necessary that the added water particles are discharged during the flow event.

Where in the catchment are the impulses to flow events given?

Generally the impulses from a water input to a flow event are weaker the more distant from a stream they are given.

The saturated discharge areas (including the stream itself) can be said to contribute to 100%. Disregarding the evaporation during the flow event, the flow impulse from these areas is equal to the precipitation on them. The impulse is here propagated by saturated overland flow.

In the recharge area the precipitation infiltrates and generates groundwater discharge in the discharge areas. Because of storage and evaporation the discharging volume is less than the total infiltration on recharge areas. The recharge areas therefore contribute by less than 100%. Earlier we have stressed that the areas close to the discharge areas, where the groundwater is shallow and the soil water content is high, play a central role in runoff generation. Infiltration here will rapidly lead to increased groundwater discharge. In these areas the contribution can be close to 100% and it is gradually decreasing when the distance to the discharge area increases.

In the discussion about the role of topography it was emphasised that hollows favour the formation of saturated discharge areas and areas with high soil water content. An important effect of this fact is that hollows to a great extent give impulses to flow events from saturated overland flow as well as from infiltration and groundwater discharge. A hillslope with somewhat fluctuating soil surface, in which hollows and ridges have been formed, may therefore give higher flow events than similar straight hillslopes with the same mean slope.

As stated on page 83 the sizes of saturated discharge areas (A_u) may be estimated from the volume of discharged fresh rainwater (V_p) in relation to the precipitation (P),

$A_u = V_p/P$

The size of the areas in which the water input gives impulses to groundwater discharge during a flow event can, however, not be estimated in the same way from the discharged groundwater volume, because these areas does not contribute by 100%. We know that they are larger than the area that would be required if the contribution was 100%, i.e., V_g/P, where V_g is discharged groundwater volume. Because the flow events normally are dominated by groundwater ($V_g > V_p$) the impulses must come from an area that is considerable larger than the saturated discharge area.

An estimate of the size of the area that in some way contributes to the flow event is achieved when the total discharged volume, V_t is compared with the precipitation. The impulses must come from an area that is larger than V_t/P, which is the area that would have been required if the contribution was 100%. Given as a fraction of the catchment, this area is equal to the fraction of the water input that is discharged during the flow event, $V_t/(P \cdot A)$, where A is the catchment area. This fraction is easy to determine. It depends among other things on the soil- and groundwater conditions before the water input. On a dry area considerable amounts of rain may be required (tens of mm) to get any noticeable increase in streamflow. The discharged water volume may in such cases correspond to a few percentages of the water input. We don't know where the infiltration takes place that results in groundwater discharge under such circumstances. A reasonable guess is that it is only areas with shallow groundwater, and therefore high soil water content that contribute. When it is raining on a wet area, as well as at large snowmelt, the discharged water volume could reach 50 - 75% of the water input. In such cases it is probable that the whole catchment gives impulses to the flow event.

The fact that the impulses to the flow events often emanate from only a part of the recharge area does not imply that the other parts of the recharge area are unimportant for the flow events. The groundwater flow from these other parts contributes to maintain a shallow groundwater surface and high soil water content in the lower parts of the catchment between flow events. In this way the influence the response of the catchment to a later precipitation event.

What is the origin of water in the flow event?

We have now discussed how the water input to the different parts of a catchment contributes to a flow event. Because this influence to a large extent is indirect, by pressure propagation, a partial area's contribution to a flow event does not automatically mean that the added water particles are discharged during the flow event. The result from the ^{18}O-study was that the major part of the water discharged during flow events was "old" water.

On the saturated discharge areas the precipitation acts directly, i.e., the rainwater that falls on it reaches the stream during the flow event (disregarding eventual storage in pools). The velocity of the water particles in the saturated overland flow depends, among other things on the water input and the slope of the soil surface and its unevenness. Sheet flow on grass may have a velocity of some cm per second. A saturated discharge area of this type and reaching 10 m from the stream would transmit rainwater to the stream within a few minutes.

Outside the saturated discharge area the velocity of the water particles is considerable lower. The vertical flow in the unsaturated zone seldom has higher particle velocity than one centimetre per hour; a velocity that only remains during and short after the rainfall (cf. the mean velocity of 0.1 – 0.2 m per month, page 56). It is only in areas that have shallow groundwater that the infiltrating rainwater reaches the groundwater during the flow event it helped to create.

The large variation in groundwater flow with soil depth in till soils is reflected in a corresponding depth variation in the flow velocity of the water particles. During experiments with added tracers at Masbybäcken (Kloten area, Västmanland, Central Sweden) particle velocities of up to 25 m per day were measured in the top 10 cm of the soil profile. Below that the particle velocity of the groundwater declined rapidly to about 1, 0.1, and 0.05 m per day at the depths 0.2, 1.0 and1.3 m respectively. Other tracer experiment in till soils with somewhat deeper groundwater level have shown similar particle velocities and also emphasized the importance of sand- and gravel lenses with high hydraulic conductivity. In such lenses at some metres depth particle velocities of about 1 m per day have been measured.

The recession of a flow event induced by a single rain event may take a week. Where in the catchment could the water particles have started that are discharged during this time? Knowing the high particle velocities in shallow groundwater it is natural to consider the groundwater level as decisive for which water particles that can reach the stream during the flow recession. The groundwater particles that reach the stream via the soil surface or directly from the uppermost soil layer should come from the area where the depth to the groundwater surface is less than a few decimetres, i.e., from the saturated or perhaps from the unsaturated discharge area. Before the flow event these water particles were probably situated at a somewhat deeper level within the discharge area, in layers with lower conductivity and particle velocity. By means of the upward directed flow in the discharge area these particles reach shallower layers where the particle velocity is high and they can therefore reach the stream rapidly. Other water particles were probably stored in the unsaturated zone, just above the groundwater surface in the vicinity of the discharge area. This soil water was transformed to groundwater after small water input by infiltration (the groundwater level rises). The soil water particles, that earlier were relatively immobile may in this way dominate the shallow, large groundwater discharge. In the major part of the recharge area the groundwater level is so deep that the groundwater particles only move a few metres during the streamflow recession. The groundwater that discharges closer to the bottom of the stream emanates from the nearest metre.

Large flow and high velocity in macropores

The tracer experiments at Masbybäcken resulted in quite different particle velocities at the same depth, one “high” and one “low”. An interpretation of this observation is that some water particles follow some more penetrable routes than others, through connected systems of larger pores. (In such case the flow is not the piston flow type.) Such structural pores, with diameter larger than a millimetre, are called macropores as distinguished from the “normal” micro pores of the soil. Because such macropores may be of great importance for the water residence times in the soil, and by that also for chemical equilibrium reactions, we will shortly treat them, in spite of almost lacking information on the occurrence and importance of macropores in Swedish conditions.

Macropores are mainly formed in the unsaturated zone by the activities of digging worms and insects, decomposition of roots, soil shrinkage when drying, and volume changes in connection with soil frost and by the dissolution of minerals from the soil. All of these activities are largest closest to the soil surface, contributing to the large saturated hydraulic conductivity there. (Darcy’s Law is valid for flow in macropores under saturated conditions as long as the flow is not turbulent. Field determinations of the saturated hydraulic conductivity include possible effects of macropore flow.)

As mentioned on page 41 the importance of the larger pore systems in connection with infiltration and percolation is debated. For a pore with 2 mm in diameter to be saturated the pressure potential of the water must, according to the equation on page 20, be above –1.5 cm, i.e., there must be almost saturated conditions. In a soil having a pore size distribution (disregarding macropores) that gives a small saturated conductivity, a root channel or crack and their closest vicinity may be saturated even if the rest of the soil is unsaturated. The macropores will then conduct water even if the main part of the soil is unsaturated. This can be the case at large water input to a clay soil. If water, on the other hand, is added to a macropore in a coarse unsaturated soil, the water will be sucked out into the unsaturated vicinities of the pore. The macropores will then not contribute to the percolation. This is probably the case in our till soils at normal low rain intensities and reasonably large depth to the groundwater surface.

The conditions are different during saturated conditions. Then all pores are water filled and if there are continuous larger pores the flow in them will dominate the flow. Because such pores preferably are formed in the root zone, their importance for groundwater flow should be greatest in areas where the groundwater level occasionally rises close to the soil surface. In the referred overland flow experiment close to Uppsala (page 53) it was observed that groundwater from one of the hillslopes flow out of 1 – 5 mm large pores in the pit wall, when the groundwater level after large rain storms or snowmelt reached a few cm from the soil surface. In this case the soil was silt soil on top of clay soil. The large pores existed only close to the soil surface, and the flow seemed to take place through the whole cross sectional area at lower groundwater levels. In this hillslope flow in macropores contributed to the groundwater flow at high groundwater levels. In some international reports on macropores, lateral flow in such pores were reported to take place in limited saturated or nearly saturated layers some decimetres below the soil surface. Contrary to the experiment in Uppsala there was an unsaturated zone below the conducting layer. It seems unlikely that such situations would appear in Swedish till soils where the groundwater is often shallow.

The role of evapotranspiration

In the discussion on runoff generation we have more or less disregarded the role of evapotranspiration. This simplification does not lead to any principal errors. We can imagine that we discussed rains that were followed by periods with small evapotranspiration (late autumn with calm, cold and moist weather) or melt periods during the winter. Let us anyhow look at some effects of evapotranspiration.

Primarily the evapotranspiration is decisive for the soil- and groundwater conditions before the rain, and thereby for the catchment's ability to generate runoff. The water input to the ground surface also decreases due to interception. By evapotranspiration water is withdrawn from both the recharge- and the discharge areas during the flow event. This effect can be observed as a seasonal variation of the streamflow recession. During summer, when the evapotranspiration is largest, the runoff declines faster than during winter. In some watercourses a daily variation in the recession may be observed (Fig. 50). This variation has been interpreted as an effect of the daily variation in evapotranspiration. Probably it is an effect of evapotranspiration from discharge areas (saturated and unsaturated) and their vicinities. Depending on time of the day larger or smaller parts of the water flow from up-slope disappear into the atmosphere instead of to the stream. This is by evaporation from free water surfaces and by transpiration from water taken up in the unsaturated and saturated zones. Model simulations have shown that daily variations in water loss to the atmosphere from the unsaturated zone can lead to daily variation in discharging groundwater.

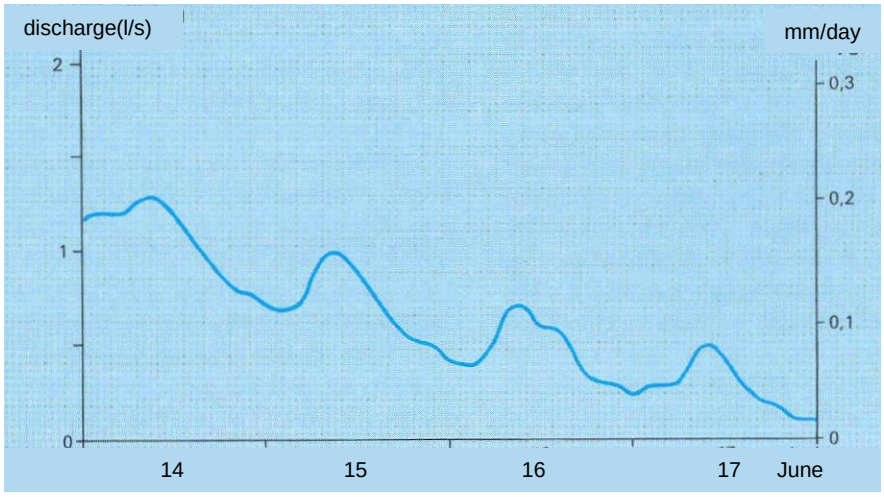


Fig. 50 Runoff during some rain free summer days in the Kilmyrbäcken, Jämtland, Central Sweden, in June 1980.
The daily variation in runoff is probably caused by evaporation from discharge areas and their vicinities. Some part of the flow from the hillslopes is evaporated to the atmosphere and does not influence the streamflow. The runoff in the stream was small during these days leading to a large relative influence of the daily variation in evaporation.

The hydrograph can be calculated if precipitation is known

We will end this chapter with a short presentation of a mathematical model by which the runoff in watercourses can be calculated if the precipitation is known. The model is developed at the Swedish Meteorological and Hydrological Institute (SMHI) and is today used on a routine basis to make forecasts on floods and recharge to hydropower dams, and also to estimate runoff in streams and rivers for periods where discharge data are lacking. It is based on the basic principles given in this book.

In the mathematical model that was presented on page 87 a real hillslope was imitated and the flow was calculated from basic physical relationships; Darcy's Law and continuity conditions. In the present model another way is chosen. The catchment could be looked upon as a number of flowerpots, having holes in their sides at different levels. From these holes the groundwater is drained to a delaying storage and then to the watercourse. Let us see how such a pot works. The calculations are performed on a basis daily, separately for each pot.

First, the water input to the ground surface is determined from measured precipitation and air temperature. At freezing temperatures the precipitation is stored as snow and at temperatures above 0°C the precipitation falls as rain. If there is a snow pack, snow melts in proportion to the number of positive degrees (cf. page 50).

All water added to the soil surface infiltrates. The percolation to the groundwater is estimated from water balance calculations for the root zone, see method 1 on page 61. These estimates use data on potential evaporation and plant available soil water storage. The fraction of a certain water input that percolates increases gradually when the calculated soil water storage approaches field capacity.

All added water percolates when the field capacity is reached. The limited percolation that takes place before field capacity in the model is reached may imitate total percolation on a limited part

of the catchment. In this way variability in the water holding capacity within the catchment is imitated. The occurrence of saturated discharge areas is not directly considered, nor the influence of topography on soil moisture conditions, but their role for runoff generation is partly satisfied by the variable water holding capacity.

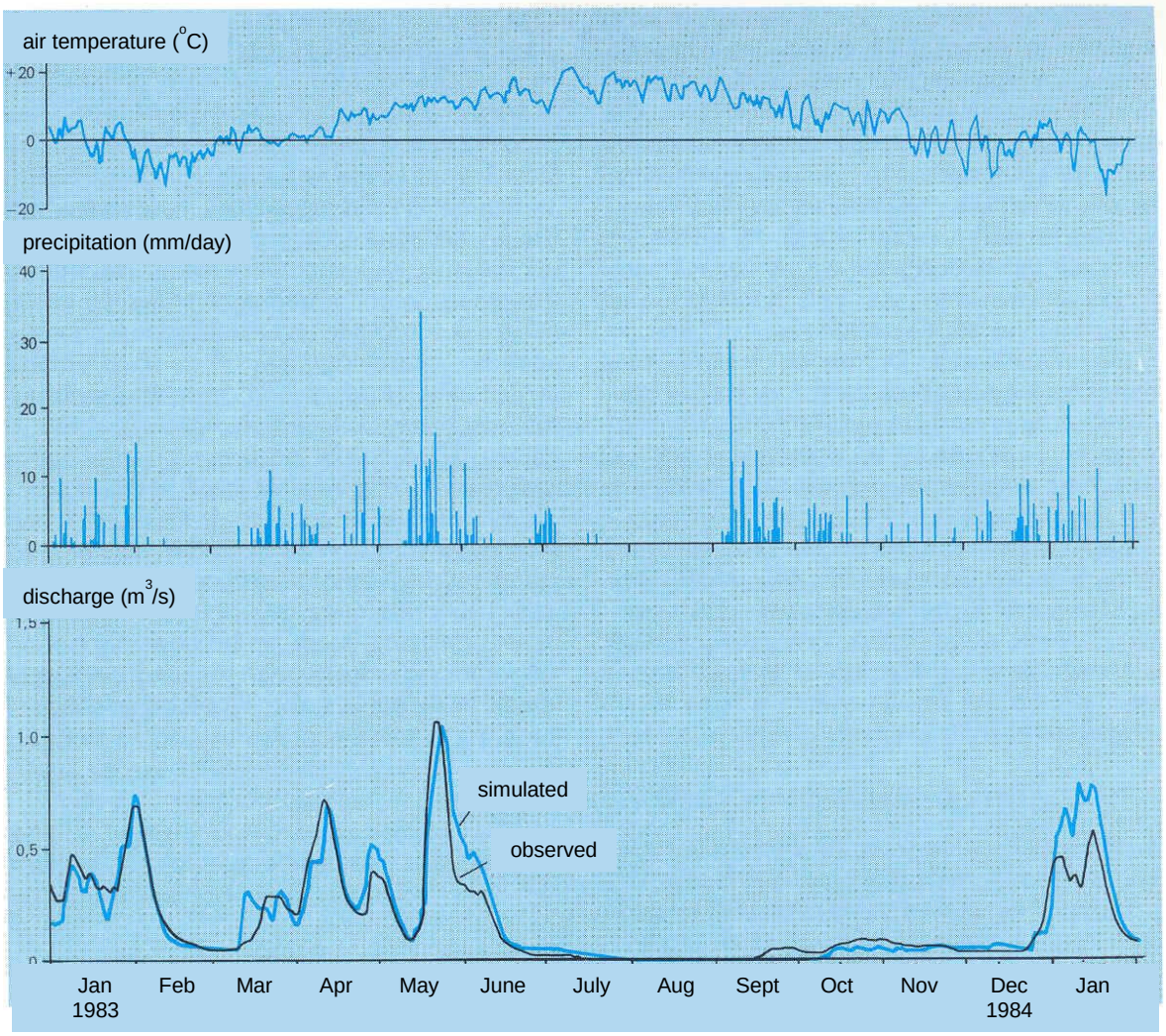


Fig. 51 Streamflow calculated from daily air temperature (determines snow storage and snowmelt), precipitation, and potential evaporation by means of a mathematical model. The diagram shows (from the top) air temperature, precipitation and observed and calculated runoff at Nolsjön, Velen-area, Västergötland, South Sweden.

The demand on the degree of agreement between calculated and measured hydrograph varies. For some applications the agreement in total runoff over a long period is stressed, i.e., that the areas below the curves are equal. In other situations it may be more important to accurately determine the maximum discharge or the date for the flow increase. .

The discharge from the groundwater zone is determined in different ways and it depends on the calculated groundwater level. The discharge is proportional to the groundwater level, but the proportional coefficient is increasing with increasing groundwater level. (The openings in the pot increase from the bottom to the top of the pot). A given rise in water level at high water levels thus leads to a larger increase in discharge than the same rise in water level at low water levels (cf. Fig. 47). In this way the model pay attention to the depth variation of the hydraulic conductivity and to the increasing slope of the groundwater surface when it approaches the soil surface.

The delay in flow response in a catchment is handled by the delay storage that the discharge has to pass. One part of the groundwater discharge from one day is discharged the same day, another the next day and so on.

Both precipitation and air temperature vary in a real catchment, for example with distance above sea level (the precipitation is largest and the temperature is lowest in the most elevated areas). In addition the snowmelt differs considerably between forest and open areas (cf. page 50). These variations in a catchment are taken into account in the model when the catchment is represented by a number of pots, having different climate and different fractions of forest and open land.

The model includes a large number of coefficients that must be determined for each new catchment where the model is to be applied. Examples of such coefficients are the maximum size of plant available water and its distribution, proportionality constants between groundwater level and discharge, and delay factors for runoff. The coefficients are determined from trial and error runs with the model and the combination of coefficients, which gives the optimal similarity between observed and model calculated runoff are chosen. When the coefficients are determined the model can calculate runoff from precipitation, air temperature and potential evapotranspiration. Five-day-forecasts of the weather can be used to make short time forecasts of runoff. To make forecasts of for example probable maximum floods for longer periods, statistics from historical flows calculated with the model by means of historical climate data from the actual or a nearby catchment are used.

Shallow groundwater during spring flood in Kassjöån area, Medelpad, Central Sweden. In the road cut at the foot of a long till hillslope the discharging groundwater reflects the meeting light. Water in the unsaturated zone does not reflect the light. There the water layers around the soil particles are too thin and the small water surfaces between the particles are concave and give no reflection.

7. Chemical processes along the water flow path

In the preceding chapters we have given examples where the chemical composition of water was used for defining the water flow paths through a watershed. We used the dissolved salts in the water or the isotopic composition of the water as tracers to infer its origin and flow paths. In this chapter we take the opposite view, and treat the chemical changes of water along its flow paths. We focus on chemical processes that change the content of dissolved substances in the water and thereby also change the environment in which the water flows.

Some basic chemical concepts that are used in this chapter are defined on pages 140 – 141. On these pages there is also a list of chemical notations used in the text.

The chemical processes in overview

Infiltrating water first passes the topsoil layer consisting mainly of more or less decomposed organic material. This humus layer contains large amounts of exchangeable hydrogen and metal cations, resulting in a damping of temporal variations in the chemical composition of the precipitation.

In the root zone soil minerals are weathered. More metal cations are then added to the soil solution. The plant roots take up nutrients and water and release carbon dioxide to the root zone when they respire. When the carbon dioxide is dissolved in soil water carbonic acid is formed, which contributes to further weathering during the water percolation to the groundwater.

Along the groundwater flow paths further weathering and mineral transformations take place. Therefore the concentration of many ions increases successively. When the groundwater finally leaves the groundwater zone in a discharge area, the partial pressures of oxygen and carbon dioxide changes and iron and calcium may precipitate.

Supply of substances to the soil surface

The soil surface in a catchment is supplied with chemical substances from the atmosphere, i.e. from outside the catchment, and from the vegetation, i.e. from inside the catchment. The external supply is a part of each element's biogeochemical cycle, including both very slow geological processes and fast chemical reactions.

The internal supply is a part of what is often called the small cycle. It consists of uptake of elements by roots, incorporation into living tissues, decay and litter fall, decomposition, release and again uptake by roots.

External supply

The inflow of chemical substances from the atmosphere to the soil surface takes place in dissolved form in precipitation, so called wet deposition or as deposition of gases and particles, so called dry deposition. Dust, salt crystals and other particles that have entered the atmosphere in different ways act in many cases as starting points for raindrop formation; they constitute the kernels on which water vapour condensates. They follow the falling raindrops to the ground and

are a part of the wet deposition. During their flight to the ground the raindrops may capture further particles from the air.

Larger particles may sediment from the air on the vegetation or directly on the ground. Smaller particles cannot sediment directly, but the trees filter the passing air and a large part of the particles get stuck on leaves and needles. These are in turn washed by precipitation. The deposition in forests is therefore larger than on open ground.

Gases, as sulphur- and nitrogen oxides, are efficiently taken up by trees through the stomata of leaves and needles. These gases may later be further oxidised to sulphur- and nitric acids and eventually reach the ground.

Sulphur (S) is supplied to the catchment as sulphate ions (SO_4^{2-}) or sulphuric acid (H_2SO_4) dissolved in precipitation, as sulphates (salt) in particles or as sulphur-dioxide gas (SO_2). The most important natural sources for the sulphur in the atmosphere is splashes when sea waves and bubbles break, volcanic activity and release of hydrogen sulphide from swamps. In the industrial countries the largest part of the emission into the atmosphere comes from combustion of fossil fuels.

In 1987 the part of the wet deposition of sulphur that did not originate from the sea, was in South Western Sweden about 10 kg per ha per year, and in Northern Sweden about 3 kg per ha per year. The total deposition is about double that of the wet deposition.

Nitrogen (N) is supplied to the catchment mainly as ammonium ions (NH_4^+), as nitrate ions (NO_3^-) dissolved in precipitation, or as nitrogen oxides (NO and NO_2) that are taken up by the vegetation. Biological nitrogen fixation mediated by certain microorganisms in the catchment adds nitrogen directly from the atmosphere.

Nitrogen gas constitutes the main part of the atmosphere. In spite of this, the supply of nitrogen as a nutrient is often limiting plant growth. To be biological useful the nitrogen must be bound to hydrogen or oxygen ions. It can be bound to hydrogen ions by biological nitrogen fixation. Nitrogen oxides are naturally formed in the atmosphere by electrical discharges. The amount of nitrogen oxides formed by combustion at high temperature is globally of the same order of magnitude.

Ammoniac (NH_3) is the end product after complete decomposition of organic material. In an acid environment, ammoniac takes up one hydrogen ion and forms an ammonium ion. On calcareous ground, limed agricultural land, in manure deposits and in other basic environments, ammoniac gas may be emitted to the atmosphere. One of the largest sources for ammoniac is today the fertiliser industry.

The wet deposition of ammonium and nitrate ions was in 1987 about 4 kg per ha per year in the South Western Sweden and 1 kg per ha per year in Northern Sweden. As for sulphur the total deposition of nitrogen is about twice that of wet deposition.

The positive ions calcium (Ca^{2+}), magnesium (Mg^{2+}), sodium (Na^+) and potassium (K^+) and the negative ion chloride (Cl^-) are supplied to the catchment dissolved in precipitation, but also as dry deposition of salt particles. These ions have entered the atmosphere from breaking waves, wind erosion from land and from combustion.

The supply of aluminium (Al), phosphorous (P), silicon (Si), iron (Fe), manganese (Mn) and other elements to the catchment is small as compared with the supply of other elements mentioned above.

Natural rainwater in equilibrium with the carbon dioxide in the atmosphere contains hydrogen carbonate and has a pH value around 5.5. In acid precipitation the negative hydrogen carbonate ion (HCO_3^-) from carbonic acid, which is a weak acid, has been replaced by the negative ions of the strong sulphuric and nitric acids, i.e. sulphate (SO_4^{2-}) and nitrate (NO_3^-) ions. The pH-value of precipitation was in the mid-1980's about 4.3 in Central Sweden.

Internal supply

Dead plant material falling to the ground as litter adds considerable amounts of nutrients to the soil surface. These have earlier been taken up from the root zone. In addition to the supply by dead plant material some elements as, e.g., potassium is leaching from the canopies and thus contribute to the internal supply. Other elements, like chloride, are not plant nutrients and do not take part in the small cycle to any considerable extent.

Chemical processes in recharge areas

The upper part of the ground, i.e. the part that is influenced by climate, living organisms and processes caused by them, is called the *soil*. When water percolates through a soil, and thereby takes part in the soil forming processes, it is itself at the same time dramatically transformed chemically – precipitation water is transformed to soil water and groundwater.

The natural dead organic debris that is not yet decomposed by the soil forming processes is called *litter*. Litter is added to the soil surface from the canopies of the vegetation (dead needles, leaves, twigs, branches etc.), from the bottom layer (dead animals and mosses) and from within the soil (dead roots, soil fauna, fungi, bacteria and other organisms). *Humus* is the organic material in the soil excluding litter and living organisms, i.e. organic material the origin of which origin is not identifiable. The humus layer consists of both humus and decomposing litter.

In recharge areas the vertical downward water movement dominates. Here *podzols* are formed; a soil characterised by more or less well developed layers. These are from above the humus layer, the bleached eluvial horizon, and the enriched illuvial horizon. Below these layers, the soil is less affected by the surface processes but may contain material in different states of weathering. The humus layer of podzols contains minimal amounts of mineral particles and is called *mor* (cf. Fig. 52).

If the soil contains a lot of digging animals, especially earth worms, the characteristic horizons of the podzols are not formed. The soil organisms mix effectively the dead organic material with mineral particles. Instead of mor and an eluvial horizon a *mull* horizon is formed. This soil type is called *brown earth*. It is formed in broad leaf forests and is especially abundant in South Western Sweden and around Lake Mälaren. In the following text we will limit the discussion to the podzols, which cover the largest area in Sweden and most clearly show the chemical processes.

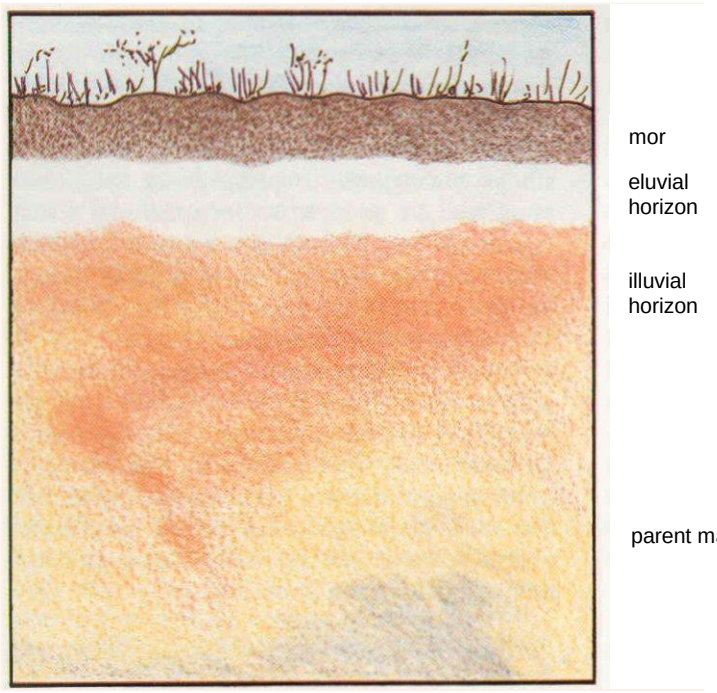


Fig. 52 Iron podzol. This soil dominates in a large moisture interval on sloping recharge areas. Iron and aluminium that went into solution in the eluvial horizon are deposited in the illuvial horizon. Deposited iron hydroxides give the illuvial horizon its rusty colour.

The litter fall from the trees and the bottom layer continuously supplies the mor layer with dead organic material (a Scots pine loses about one third of its needles every year). Bacteria, fungi, and soil fauna use the dead material as energy source and thereby successively decomposes it. The readily decomposable parts turn over relatively fast, while the material most hard to decompose may rest in the mor layer for several hundred years. In the mor layer there are also plant roots, which can take up mineralised nutrients.

The mor layer gives its signature to passing water

The decomposition of the organic material is an oxidation, ultimately forming carbon dioxide gas (CO_2), but also a large number of hydroxyl- (*-OH*) and carboxyl- (*-COOH*) groups are formed. The hydrogen ions in these are weakly bound and can easily be removed into the soil solution. They are therefore acids. The organic acids formed have molecules of very different sizes, from humic acids with many thousand carbon atoms to citric acid, malic acid and butyric acid with four to six carbon atoms. While small organic acids are soluble in water and can follow the infiltrating water down through the soil profile, the larger humic acids have lower solubility and remain to a larger extent in the mor layer.

When the humus hydroxyl groups release hydrogen ions these are exchanged with other positive ions, cations, from the soil water. The humus therefore acts as an *ion exchanger* for positive ions. The mor layer in a forest represents a very large cation exchanger and can damp short term fluctuations in the composition of the precipitation. Regardless which positive ions there are in the precipitation of one storm, the percolating water gets a rather uniform cation composition when it passes the ion exchanger in the mor layer.

The relative amount of all negative sites on the ion exchanger that is occupied by the so called basic cations Ca^{2+}, Mg^{2+}, Na^{+}, and K^{+} is called *base saturation*. To the rest of the sites mainly hydrogen and aluminium ions are bound. The basic cations, most of which are essential mineral nutrients, have got their name because their oxides have a basic reaction. This should not be mixed up with the pH-reaction of the soil. The larger proportion of the cation exchange sites in the mor layer that is occupied by hydrogen ions, the larger is also the hydrogen ion concentration

in the soil water, i.e., pH is low. A higher pH-value in the soil water can either be the result of a higher proportion of basic cations on the exchange sites of the solid phase, or a higher proportion of aluminium ions. In the latter case, aluminium displays basic properties, while it may have acid properties in other cases.

The mor layer also represents a large reservoir of basic cations. In a forest soil in Central Sweden this reservoir is as large as about 50 years' dry and wet deposition from the atmosphere. In coniferous forests the basic cation addition from litter fall is of the same order of magnitude. The turnover rate (cf. page 9)) for basic cations in the mor layer is therefore about 25 years.

When water evaporates from the soil surface, the concentration of dissolved substances increases, because it is only the pure water that evaporates. The percolating water from the mor layer therefore has a higher concentration of dissolved substances than the precipitation. Soluble organic acids have also been added to the water and the pH-value corresponds to the humus base saturation.

The crystal structure of the different minerals determines their tendency to weather

The mineral soil in till consists of a mechanical mixture of larger and smaller fragments from different rock species and of different transformation products. Which species of rock a certain till contains depends on where the glacial ice scraped off the soil. The species of rock consist of different minerals, each of which has its individual chemical composition. Oxygen is the most common element, followed by silica, aluminium and iron. Because the oxygen atom is also relatively large, oxygen occupies about 92% of the total volume of the solid rock.

Besides oxygen, the most common minerals have silica as their dominating atoms. These minerals are called silicates. Before we treat the decomposition or *weathering* of silicates in the soil we will briefly treat their structure.

The basic structure of silicate minerals is symmetrically ordered oxygen atoms. These are commonly ordered as corners of a tetrahedron. (A tetrahedron is a body, with sides consisting of four triangles with equal sides). The negative charge of the oxygen ion (two negative charges per oxygen ion) has to be compensated by positive ions. For the structure to be stable, the positive ions must be in contact with all surrounding oxygen ions (for tetrahedron order with four oxygen ions). The better the positive ions fit into the spaces between the oxygen ions, the more stable is the structure. The ratio between the radius of the positive ion and that of the oxygen ion is therefore a measure of the stability, i.e., the ability of a mineral to resist weathering.

If the ratio between the radius of the positive ions and the oxygen ions is 0.15, the oxygen ions will form corners in equilateral triangles. If the ratio is 0.22 they will form corners in tetrahedrons, if it is 0.41 corners in octahedrons and if it is larger than 0.73 corners in cubes. (An octahedron is a body which sides consist of eight equilateral triangles, like two pyramids with their bottoms together.) The oxygen ions can be ordered like a tetrahedron for all ion radius ratios from and including 0.22 to 0.41. If the ratio gets smaller than 0.22 the structure collapses as the positive ions cannot have contact with all for oxygen ions. If, on the other hand, the ratio gets larger than 0.22 the stability decreases, but can still be kept together up to an ion ratio of 0.41.

The silica ion is the ion that best fits into the spaces between oxygen ions ordered like a tetrahedron. It has an ion ratio to the oxygen ion of 0.30 and in addition the charge of +4. Quartz (SiO_2) has this structure and is therefore the most stable silicate mineral. Aluminium ions,

having an ion-ratio to oxygen ions of 0.43 fits second best, but as the ratio is larger than 0.41 the oxygen ions will be ordered like an octahedron around the aluminium ions. Single aluminium ions can substitute silica ions while the structure like a tetrahedron remains. But the aluminium ions have only the charge +3, so some other cation has to be included in the structure to keep the charge balance between positive and negative charges.

Potassium feldspar ($K\,Al\,Si_3\,O_8$) has the same composition as quartz, but every fourth silica ion has been exchanged by an aluminium ion. To make the charges in balance a potassium ion (K^+) is included for every aluminium ion. Potassium feldspar has a less optimal structure than quartz and decomposes more easily. In this way the different minerals can be ordered into a hierarchy after falling stability (table 4).

Weathering processes		
Physical	Temperature variations	Alternating temperatures may break a rock that consists of minerals with different thermal expansion coefficients. The tension that arises between the surface of a rock and its deeper parts due to daily temperature variations also results in breakage of the rock.
	Frost	Water that freezes in cracks exerts a large breaking effect.
	Vegetation	Roots in cracks exert a breaking effect.
Chemical	Hydration	Water is absorbed in the mineral.
	Hydrolysis	The formation of an acid and a base when water reacts with a salt.
	Dissolution	A mineral is dissolved in water and carbonic acid.
	Oxidation/Reduction	Mainly iron that may be oxidized in a mineral leading to its destruction.

Primary minerals weather primarily through hydrolysis

Physical weathering is a mechanical disintegration of a species of rock, while *chemical weathering* is a decomposition of the crystal structure of a mineral. The larger the relative surface area of a mineral grain is, i.e., the smaller the grain is, the more efficient is the chemical attack. The physical weathering therefore enhances the chemical weathering. The products from the chemical weathering may be free ions or new solid phases, called secondary minerals in contrast to the original or primary minerals. The secondary minerals are relatively richer in aluminium and iron than the primary minerals from which they stem.

Some minerals can take up water, so called *hydration*. Among the primary minerals this is the case almost only for biotite, which consists of thin plates. When water is absorbed between the plates, the biotite swells and finally breaks. Under influence of water salt minerals can *dissolve*. Calcite is also weathered through dissolution, but the reaction will only be of importance if carbon dioxide is dissolved in the water. Minerals containing iron (II) (Fe^{2+}) can weather in the presence of oxygen gas. The iron is *oxidised* to iron (III) (Fe^{3+}). Then the mineral will be unstable and disintegrates.

Table 4. *Weathering ability of small mineral particles (1 = easy, 15 = difficult), abundance of the minerals in Sweden (1 most abundant) and their chemical formula.*

Weathering group	Abundance	Mineral (p=primary)	Chemical formula
1		Gypsum	$CaSO_4\ 2H_2O$
2	7	Calcite	$CaCO_3$
3	5	Hornblende (p)	$3CaO{\cdot}Na_2O{\cdot}8(Fe^{2+}, Mg)O{\cdot}2(Al,Fe^{3+})_2O_3{\cdot}\ 14SiO_2{\cdot}2H_2O$
4	4	Biotite (p)	$K_2O{\cdot}6(Fe^{2+}, Mg)O{\cdot}Al_2O_3{\cdot}6SiO_2{\cdot}2H_2O$
5		Albite (p)	$Na_2O{\cdot}Al_2O_3{\cdot}6SiO_2$
	2	Anorthite (p)	$CaO{\cdot}Al_2O_3{\cdot}2SiO_2$
	3	Potassium feldspar (p)	$K_2O{\cdot}Al_2O_3{\cdot}6SiO_2$
6	1	Quartz (p)	SiO_2
7	6	Muscovite (p)	$K_2O{\cdot}3Al_2O_3{\cdot}6SiO_2{\cdot}2H_2O$
8		Vermiculite	$Mg_{0,55}(Al_{2,3}Si_{5,7})(Al_{0,5}Fe^{3+}{}_{0,7}Mg_{4,8})O_{20}(OH)_4{\cdot}nH_2O$
9		Montmorillonite	$Ca_{0,4}(Al_{0,3}Si_{7,7})Al_{2,6}(Fe^{3+}{}_{0,9}Mg_{0,5})O_{20}(OH)_4{\cdot}nH_2O$
10	8	Chlorite	$AlMg_5(OH)_{12}(Al_2Si_6)AlMg_5O_{20}(OH)_4$
11		Allophane	$Al_2O_3{\cdot}mSiO_2{\cdot}nH_2O$
12		Caolinite	$Al_4(Si_4O_{10})(OH)_8$
13		Gibbsite	$Al(OH)_3$
14		Goethite	$FeO{\cdot}OH$
	10	Magnethite (p)	Fe_3O_4
15		Rutile	TiO_2

The process that is of largest quantitative importance in disintegration of primary mineral is *hydrolysis*. At hydrolysis the primary minerals react with water and an acid (silica acid) and a base (aluminium hydroxide) are formed. The decomposition of the crystal gitter is facilitated when hydrogen ions are absorbed into it at the same time as basic cations are released. Silica acid may react with aluminium hydroxide to form secondary silicates, i.e., clay minerals. The clay mineral that is formed depends on the concentrations of the different components in the soil solution.

As an example of hydrolysis we give potassium feldspar ($K_2O \cdot Al_2O_3 \cdot 6SiO_2$) disintegration into silica acid (H_4SiO_4), aluminium hydroxide, or gibbsite ($Al \cdot (OH)_3$) and potassium ions (K^+).

$$2\ K_2O \cdot Al_2O_3 \cdot 6SiO_2 + 28\ H_2O + 4\ H^+ \leftrightarrow 12\ H_4SiO_4 + 4\ Al \cdot (OH)_3 + 4\ K^+$$

During the first level of silicate weathering, hydrogen ions are exchanged with base cations in the surface layer of the mineral. Successively hydrogen ions diffuse into the mineral crystal, while basic cations diffuse outwards. Then also aluminium and silica disappear from the surface layer. In a very acid environment (pH-values lower than 3.5) aluminium has a greater tendency to leave the mineral crystal than silica, while the opposite is true in very basic environment (pH-values larger than 9.5). In the first case the weathered surface layer contains, apart from oxygen, only silica. In the second case it contains only oxygen and aluminium.

At more normal pH-value (between 4.5 and 9) a new solid phase is formed on the surface of the weathering mineral. This so called *weathering skin* mainly consists of aluminium hydroxide, but can also contain clay minerals. These secondary silicates are also called layered minerals and are formed from aluminium hydroxide and silica acid if the silica acid concentration is high enough.

Clay minerals mainly consist of *Si-O*-tetrahedrons and *Al-O-OH*-octahedrons. All clay minerals have originated from aluminium hydroxide and silica acid that have been combined. Our clay soils contain, apart from clay minerals, also fragments of primary minerals. The relative concentrations of primary and secondary minerals vary with the textural composition of the soils. The fine clay, for example, with particles smaller than 0.0002 mm in diameter, almost exclusively

consists of secondary minerals, while fine silt (Swedish: mjäla), with particles between 0.02 and 0.002 mm in diameter, almost exclusively consists of primary minerals.

It has been shown that the weathering rate increases at very low or very high pH-values, and that the rate increases more the more extreme the pH-value is. When the pH-value is between 4.5 and 9, i.e., the pH-value of most natural waters, the weathering rate is rather independent of the pH-value. This might be the effect of the weathering skin. It may act as a resistance for the outflow of weathering products and the inflow of water. The weathering skin is continuously formed at the surface of the weathering mineral and is disintegrated at the boundary to the free soil water, especially from the action of organic complex formers.

From the mor layer, organic decomposition products, especially organic acids are leached. These effectively bind iron (III) and aluminium ions in so called *chelate* complexes (Greek: chelé, pincers). The iron and aluminium ions that otherwise are bound to solid organic material (humus), or to oxygen ions in hydroxides, can be transported as complexes with soil water down in the soil profile. The organic acids therefore contribute actively to enhance the weathering rate.

Weathering and enrichment give the soil profile a characteristic colour

Due to the continuous flux of organic acids from the mor layer, the uppermost mineral soil layer has been subjected to heavy chemical weathering. As it has lost a lot of iron and manganese its colour has become pale grey and the horizon is called the *bleached* or *eluvial* horizon. During the weathering process the easily decomposed minerals are first decomposed. Therefore the proportion of quartz, that hardly weather at all as long as feldspar is present, increases with time in the eluvial horizon.

When the percolating water has passed the eluvial horizon, where hydrogen ions are consumed in the weathering process, it has become less acid and contains more metal cations than it did when it left the humic layer. The addition of sodium, potassium, calcium and magnesium from weathering may be of the same order as the atmospheric deposition of these elements.

In the enrichment or *illuvial* horizon that is formed below the eluvial horizon, the iron from the eluvial horizon is normally deposited as hydroxide. This gives the illuvial horizon its rusty colour. Humic acids which have percolated to this horizon are also deposited and give the horizon a darker brown colour. The boundary between the eluvial and the illuvial horizons is normally very sharp, while the transition from the illuvial horizon to the unaffected parent material often is diffuse.

Roots are found in the eluvial as well as in the illuvial horizon. The amount of roots decreases rapidly with soil depth, and below 0.6 m the amount of roots is small. Plant roots take up most of the mineralised nitrogen and a large proportion of the released mineral nutrients. When parts of the plants die, the nutrients will again be transferred to the soil as litter. In this way a large amount of plant nutrients circulates.

Root respiration gives carbonic acid to soil water

The roots and the micro-organisms in the soil consume oxygen and release the same amount of carbon dioxide in their respiration. The soil atmosphere is therefore to some extent depleted in oxygen, but has a surplus of carbon dioxide as compared with the free atmosphere. The carbon dioxide concentration in the root zone is often 30 to 100 times higher than in the atmosphere, but can under some circumstances be much higher. In the same manner as heat is transferred

through a solid body from the warm to the cold side, a gas moves through the air or water in the soil from areas with high to areas with low concentrations of that gas. The reason for these fluxes is the chaotic movements of the molecules and the process is called molecular *diffusion*. Because the carbon dioxide concentration in the soil gas is many times that in the free atmosphere carbon dioxide diffuses from the soil into the atmosphere. In the same way oxygen diffuses from the free atmosphere down into the soil profile where the oxygen concentration is lower. In water the gas diffusion is more than thousand times slower than in free air. The soil water content plays an important role for the gas exchange between the soil and the atmosphere. If the root zone is water saturated, the oxygen concentration very soon falls to near zero and reducing conditions arise. Then sulphate in the water may be reduced to sulphide, for example in the form of hydrogen sulphide (H_2S) or pyrite (FeS_2). Nitrate may be reduced to free nitrogen gas or nitrous oxide gas. All these reactions are mediated by micro-organisms, which use sulphate-sulphur or nitrate-nitrogen as electron acceptors, while organic carbon is electron donors. Deposited Fe (III) may be reduced to water soluble Fe (II). The carbon dioxide concentration will at the same time be very high.

When carbon dioxide dissolves in water (reaction 1 below) carbonic acid is formed (reaction 2). The carbonic acid dissociates into a hydrogen ion and a hydrogen carbonate ion (reaction 3). The latter is further dissociated into a hydrogen ion and a carbonate ion (reaction 4).

$$CO_2(g) \leftrightarrow CO_2(aq) \quad (1)$$
$$CO_2(aq) + H_2O \leftrightarrow H_2CO_3 \quad (2)$$
$$H_2CO_3 \leftrightarrow H^+ + HCO_3^- \quad (3)$$
$$HCO_3^- \leftrightarrow H^+ + CO_3^{2-} \quad (4)$$

The notations in parenthesis mean g = gas and aq = dissolved in water.

The last equation can be neglected at pH-values lower than 8, i.e., for almost all natural waters. The hydration of the carbon dioxide (reaction 2) is not of interest at equilibrium situations. Therefore reaction (2) and (3) are combined to

$$CO_2(aq) + H_2O \leftrightarrow H^+ + HCO_3^-$$

The reactions are *equilibrium reactions*, which means that they can go to the right as well as to the left in the formulas above. In which direction they go is determined, apart from pressure and temperature, by the concentrations of the different species that take part in the reactions. If the concentration of a species to the left in an equilibrium reaction is increased, the reaction will go to the right and vice versa. The reactions are an example of a *buffer system*. If hydrogen ions are consumed in any other reaction the reactions in the carbonic acid system will go to the right and produce more hydrogen ions. If hydrogen ions are added the reactions will go to the left. In this way a buffer system counteracts changes in hydrogen ion concentration of a system. In the root zone hydrogen ions are produced when the carbon dioxide from the respiratory processes dissolves in soil water. These hydrogen ions are consumed in the weathering process. Then the reactions go to the right and more carbon dioxide can dissolve. Since one hydrogen carbonate ion is formed for every hydrogen ion that is consumed, the hydrogen carbonate concentration is a measure of the weathering. Since the addition of carbon dioxide in the root zone is large the carbon dioxide concentrations are not influenced by this consumption, and the root zone is said to be an open system with respect to carbon dioxide.

If the velocity of the water particles through the root zone is small a lot of weathering can take place during its passage, and the hydrogen carbonate concentration in the water that leaves the zone will be high as well as the concentration of cations released by weathering. If, on the other

hand, the particle velocity is high, the hydrogen carbonate and the cation concentration will be low.

If there is calcite, i.e., pure calcium carbonate ($CaCO_3$) present in the root zone it weathers as

$$CaCO_3(c) + H_2O + CO_2(aq) \leftrightarrow Ca^{2+} + 2HCO_3^-$$

The notation (c) means that calcite is in crystalline form.

From the reaction formula we can see that two hydrogen carbonate ions are formed for each calcium ion that is released. Since calcite is an easily weathered mineral the groundwater in calcite rich soils have high concentration of hydrogen carbonate. These ions act as buffers if acids are added, which makes carbonate rich waters resistant against acidification.

Below the root zone the weathering rate is decreasing

In the intermediate zone, between the root zone and the groundwater zone, the partial pressure of oxygen and carbon dioxide is about the same as in the root zone. Therefore the weathering can proceed, but it will be much less intensive since the organic acids and the complex formers were deposited in the illuvial horizon or have been mineralised. The concentration of basic cations and other weathering products in the soil water also increases down in the unsaturated zone, which hamper further weathering.

As a consequence of the weathering and a decreased concentration of organic acids, pH and the concentration of silica acid (H_4SiO_4) increase down in the soil profile. This leads to a succession of weathering reactions, leaving quartz at the top, goethite ($FeOOH$) and diaspora ($AlOOH$) in the illuvial horizon, caolinit ($Al_4Si_4O_{10}(OH)_8$) further down and finally montmorillonite deep down in the groundwater zone. This order can be seen as a final stage. Our young soils have a long time left to this stage.

Because gas diffusion in water is very slow, the oxygen and the carbon dioxide that were dissolved in the percolating water from the intermediate zone cannot be renewed in the groundwater zone. This means that the groundwater zone is a closed system regarding carbon dioxide, which restricts the possible weathering of calcite there. If the groundwater zone contains organic material, all dissolved oxygen will be consumed and the groundwater will be oxygen free. Iron(III) (Fe^{3+}) may then be reduced to Fe^{2+} and dissolve. Sulphate (SO_4^{2-}) may be reduced to sulphide (e.g. FeS_2) and precipitate.

In cases where weathering is large in the unsaturated zone it will be small in the saturated zone. If the groundwater table is close to the soil surface, then only little weathering can take place in the unsaturated zone and some weathering can take place at depth. But this weathering is limited by the amount of dissolved oxygen and carbon dioxide. In the groundwater zone the weathering rate decreases at the same time as the concentration of weathering products increases. Hydrogen ions are consumed during the weathering and therefore the pH-value increases with increasing groundwater age. Normally the concentration of silica acid and basic cations increases. Old groundwater therefore has higher salt concentration than young groundwater

Chemical processes in discharge areas

In discharge areas on till soils the soil profile normally is characterised as some kind of *humic podzol*. The eluvial and to some extent also the illuvial horizon is impregnated with humus, which gives the profile a dark brown colour (Fig. 53). A prerequisite for the humic podzols to be formed is that the discharge area is unsaturated during the main part of the year, since podzols can only be formed if there is a downward flux of water through the soil profile.

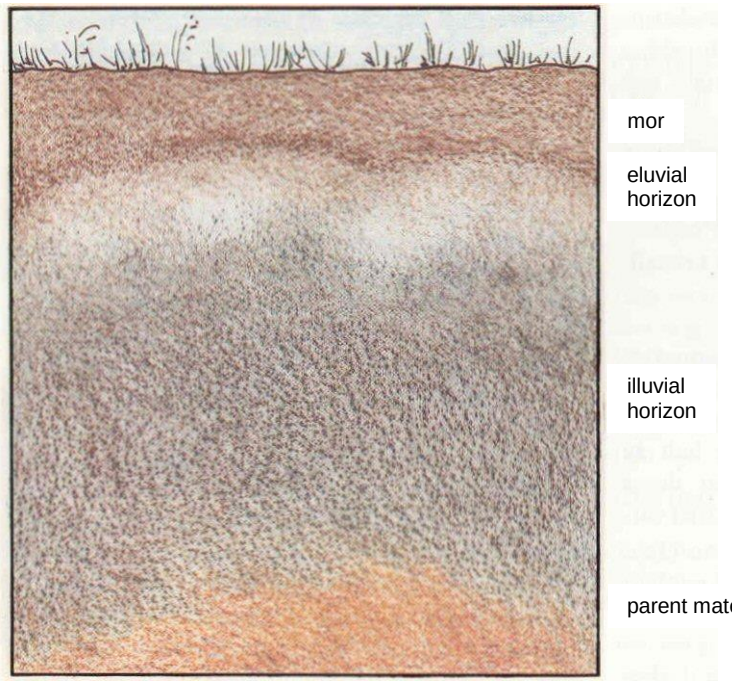


Fig. 53 Humic podzol. This soil is formed in permanent discharge areas with occasional very shallow groundwater. Humic acids have penetrated and deposited in both the eluvial and illuvial horizons.

In the same way as in the recharge areas, discharge areas receive deposition from the atmosphere. There is also a biological cycle with nutrient uptake from the root zone and a litter fall to the ground surface. The great difference is that the soil profile in a discharge area receives discharging groundwater from below and from the upslope. This groundwater has a higher pH-value and is richer in basic cations than infiltrating precipitation. It also often contains aluminium that has formed complexes with organic acids.

The water that percolates from the root zone in a recharge area will sooner or later come out as groundwater in a discharge area. Because the discharge area normally is much smaller than the recharge area the efflux per unit area will be much larger than the corresponding percolation in the recharge area.

When the discharging groundwater reaches the root zone the ion exchanging complexes in this zone strive to reach equilibrium with the composition of the discharging groundwater. In this way the storage of positive ions in the soil will to a large extent be determined by the composition of the groundwater.

The high concentration of basic cations in groundwater, as well as the large water flow, both contribute to the large cation storage in a discharge profile. It may be of the order of three times as large as in a recharge profile. Because of the large flow the turnover time (storage/flux) for the basic cations would still be shorter than in the soil profile in the recharge area. If the lateral flow component is large the aluminium concentration can also be very high.

In the root zone discharging groundwater dissolves some organic material and also brings away some humic particles in suspension. The dissolution will be larger when the hydrogen ion saturation of the humic complexes is large. Streamwater in coniferous forested areas is brown

When groundwater is discharging from the soil, the carbon dioxide goes into equilibrium with the atmosphere. In this carbon rich area this causes a drastic decrease of the carbon dioxide concentration of the water. Parts of the calcium that was dissolved from the soil and bedrock can no longer stay in solution. Calcium carbonate is precipitated and forms turf around the spring. Svindalen, Jämtland, Central Sweden. Photo: Allan Rodhe.

coloured from dissolved humic substances, while streamwater in deciduous forested areas often is almost colour less, even though the two waters can contain equal amounts of dissolved organic material. This is because the latter contains other humic fractions which are colour less.

If the groundwater is oxygen free it will again be oxygenated in the discharge area, and dissolved iron- and manganese compounds will be oxidised and precipitate as hydroxides. At springs one can often see rust iron deposits. Also some forms of iron pans, i.e., soil which has been glued together by iron and aluminium hydroxides and humic substances, may have been formed in this way.

In discharge areas the dissolved carbon dioxide equilibrates with that of the atmosphere, which results in a drastic decrease of the carbon dioxide concentration of the water. This means that the pH-value of moderately acid groundwater increases. In lime rich soils the calcite dissolution can reach equilibrium already in the unsaturated zone. The partial pressure of carbon dioxide is high in such groundwater and it has high concentration of dissolved calcium. When the groundwater is discharging, the carbon dioxide pressure decreases and calcium carbonate may precipitate and form turf.

The characteristics of discharging groundwater vary in time. During periods with shallow groundwater and large flow the discharging groundwater is comparatively young. It has low pH-value and low concentration of dissolved basic cations. When the groundwater level is low the discharging water is older and the pH-value, as well as the concentration of basic cations, is high.

Another possibility is that the characteristics of the groundwater change when it is discharged through occasional discharge areas. Hydrogen ion saturated carboxylic groups, formed during the mineralisation of soil organic substances, can take up metallic cations from the soil solution. The pH-value of the soil water and its concentration of metallic cations will then decrease.

In the discharge areas precipitation is added to the discharging groundwater. Its chemical composition is only weekly influenced in a saturated discharge area. In an unsaturated discharge area on the other hand, the precipitation water will equilibrate with the ion exchange system of the humus layer. Then it takes part in the weathering processes in the eluvial horizon and it may also deposit some weathering products in the illuvial horizon before it mixes with discharging groundwater.

When *Sphagnum*-mosses grow uronic acids are formed in them. These are excellent cation exchangers. With these the mosses can catch basic cations from very dilute solutions (for example from the precipitation), but in exchange they deliver hydrogen ions to the soil water. This is a reason why mire waters are rather acid.

Soil and mosses indicate water conditions

The soil reflects, as we have shown, the water movement through the soil profile with a characteristic difference between recharge- and discharge areas. By digging a soil pit and study the soil profile it is possible to get a view of the water conditions at the site. Also the vegetation can, as is well known, indicate the water conditions. Especially good information is gained from the occurrence of different moss- and lichen species. Let us follow the changes of the soil profile along a hillslope.

At the top of a recharge area the water flow in the unsaturated zone is vertical and downward. The distance from the soil surface to the groundwater surface is large enough for the upper soil horizons to be rather dry and well ventilated. The good aeration of the humic layer favours mineralisation processes. The humic layer is therefore thin. The soil profile is leached from above by the downward moving water, and a podzol profile, with its characteristic eluvial horizon below the mor-layer and then an illuvial horizon, is formed. On the driest soils *lichen podzols* are formed. Star-tipped reindeer lichen (*Cladonia stellaris)*, Reindeer lichen (*Cladonia rangiferina*), and Island Lichen (*Cetraria islandica* L.) are characteristic for these soils. The mor-layer is only 1 – 2 cm thick, the grey eluvial horizon is 2 – 3 cm and the rusty illuvial horizon is normally 10 – 20 cm thick.

On sloping recharge areas *iron podzols* are found. These soils are found within a large moisture interval. Here Red-stemmed Feather-moss (*Pleurozium schreberi*), Broom fork-moss (*Dicranum Scoparium*) and Stair-step Moss (*Hylocomium splendens*) are found. While the lichen podzols were formed below a thin mor-layer, the iron podzols were formed below a rather thick mor-layer. It may be 4 – 10 cm thick. The grey-white eluvial horizon may be 3 – 6 cm thick and is well distinguished from the illuvial horizon, which may be thinner than 10 cm on fine grained soils and thicker on coarser soils. The darker the colour of the illuvial horizon, the moister is the soil on average. Normally this is the reason why the colour is darker further down in the recharge areas.

In depressions, pits and close to boulders the eluvial horizon is often thicker than in the surrounding soil. The reason for this is probably that the soil is comparatively moist in the depression. The reason that water is gathered in the surface layer of a pit is not necessarily Hortonian overland flow. The reason may be that the unsaturated hydraulic conductivity of the humic layer is larger than in the mineral soil below it. In such a case water will flow along the lower parts of the humic layer towards the bottom of the pit, where a vertical break through occasionally takes place (cf. A coarse layer diverts unsaturated flow, page 33). In this on average moister environment, moisture loving mosses like Common Haircap Moss (*Polytrichum commune*) grow. A reason why the eluvial horizon is thicker below such mosses may be the composition of the organic acids that are formed and on the life length of these.

Further down in the hillslope, where the recharge area occasionally is converted to a discharge area, the soil moisture content is higher and patches of *Sphagnum*-mosses occur. The mor-layer will be more peaty and *iron humic podzols* are formed. The eluvial horizon looks like that in the iron podzols, but is normally thicker. If the soil is flat, or only weakly sloping, the lateral water movement is limited. At the top of the illuvial horizon a thin humus rich black-brown layer is then formed. Below that layer a more normal rust-coloured layer is formed. Since the groundwater level often is high, oxygen deficiency is frequent below the illuvial horizon. The soil will here get a blue-grey colour from iron (II) that is transported upward by the groundwater. If there are coarser pore systems, the soil profile is oxygenated along these when the groundwater level falls. Then the iron (II) is oxidised to iron (III), which is deposited and can be observed as rusty streaks or patches in the blue-grey background. These formations are called *gley*.

In coarse soils, or in steep hillslopes having considerable lateral water movement close to the soil surface, the illuvial layer will be more diffuse. The upper humus-rich layer may be split into a number of darker patches. Also the rusty coloured layer can contain such dark patches and it turns diffusely into the gley-layer.

In permanent discharge areas occasionally having very shallow groundwater *humic podzols* are formed. Here Common Haircap Moss or *Sphagnum* spp. dominates the bottom layer. The humic

layer is thin peat or peaty mor. The eluvial horizon is more or less impregnated by humic substances which gives it a dirty grey colour. If the illuvial horizon is thick it will be almost black from deposited humic substances. Otherwise it is dirty brown and less rich in deposited colloids. The latter type is formed when the groundwater surface is very shallow during most of the year and where the flow component that is directed upwards is comparatively large.

In the mineral soil below mires, where the water flow is often directed upwards and where there is oxygen deficiency, the blue-grey marsh soils are found. As mentioned earlier, the blue-grey colour is given by iron (II).

The chemical processes contribute to a change in the hydraulic conductivity of the soil profile. The weathering and leaching from the eluvial horizon mean that small particles slowly disappear. The hydraulic conductivity will then increase, while the water holding capacity of the horizon will decrease. The opposite holds true for the illuvial horizon. When weathering products from the eluvial horizon and humic substances from the mor layer are deposited, the mean pore diameter decreases, the hydraulic conductivity decreases, and the water holding capacity increases. If the deposition has been extremely large and iron pans have formed, the hydraulic conductivity may be so small that the layer is practically impermeable for water.

Streamwater reflects the water flow pathways through the catchment

The chemical composition of streamwater at a given place along a stream is determined, not only by the composition of precipitation, i.e., the starting value, but also by the chemical changes the precipitated water has undergone along its different flow paths to the place of observation. Measurements over time will show temporal variations of the streamwater composition. The reason is above all a changes in the relative contribution of water with different flow paths and transit times, variation in the composition of precipitation and dry deposition, but also changes in biological activity, for example due to temperature variations.

If we instead observe the streamwater composition along a stream at a certain time we will also find a variation. Reasons for that are for example that the flow pathways for groundwater to its point of discharge increases downstream, that the character of the discharge areas varies along the stream, and also biological activity in the stream.

One site – variation over time

We have earlier stressed that the large ion exchange capacity of the humic layer strongly damps occasional variations in the composition of infiltrating water. In this way percolating water gets a fairly constant composition. On saturated discharge areas the precipitation water can be diverted to the watercourse without drastic chemical changes. When the runoff is small and the groundwater level is low the discharge areas are small and the groundwater discharge to the stream takes place at the bottom or walls of the stream and the groundwater has not had much contact with the organic material in the discharge areas. The pH-value of the streams will then be high and corresponding to the pH-value of deep groundwater. When the groundwater level increases, the proportion of shallow groundwater increases as well as the proportion of precipitation running off without any contact with the mineral soil. This is the reason why the pH-value of the streamwater decreases when the discharge increases (fig. 54, see also page 127).

In a similar way, elements such as silicon, calcium, magnesium and sodium, the main source of which is weathering minerals, have high concentration in streamwater in connection with low

groundwater level and low streamflow. In the Kloten area it was found that the concentration of the negatively charged sulphate ion correlated with the calcium- and magnesium concentration, but that the negatively charged chloride ion correlated with sodium, calcium and silicon.

In addition to the hydrogen ion concentration, the concentration of potassium increases with increasing streamflow. Potassium is released by weathering, but also by mineralisation of organic material. It may be fixed in clay mineral formation in the deeper horizons of discharge areas. Nitrate- and ammonium nitrogen are also released by mineralisation. In groundwater the potassium concentration varies in parallel with the other weathering products, but this coupling is not seen in streamwater. The biological role of potassium may dominate in discharge areas.

The electric conductivity of streamwater along a stream at two different discharge situations was shown in Fig. 38. The electric conductivity was reduced for the influence of hydrogen ions and therefore showed how the metal cations increased downstream. At a certain place in the stream, the electric conductivity was lower when the discharge was large than when it was small. The reason for this is that the stream gets a larger proportion of "young" groundwater at high discharge.

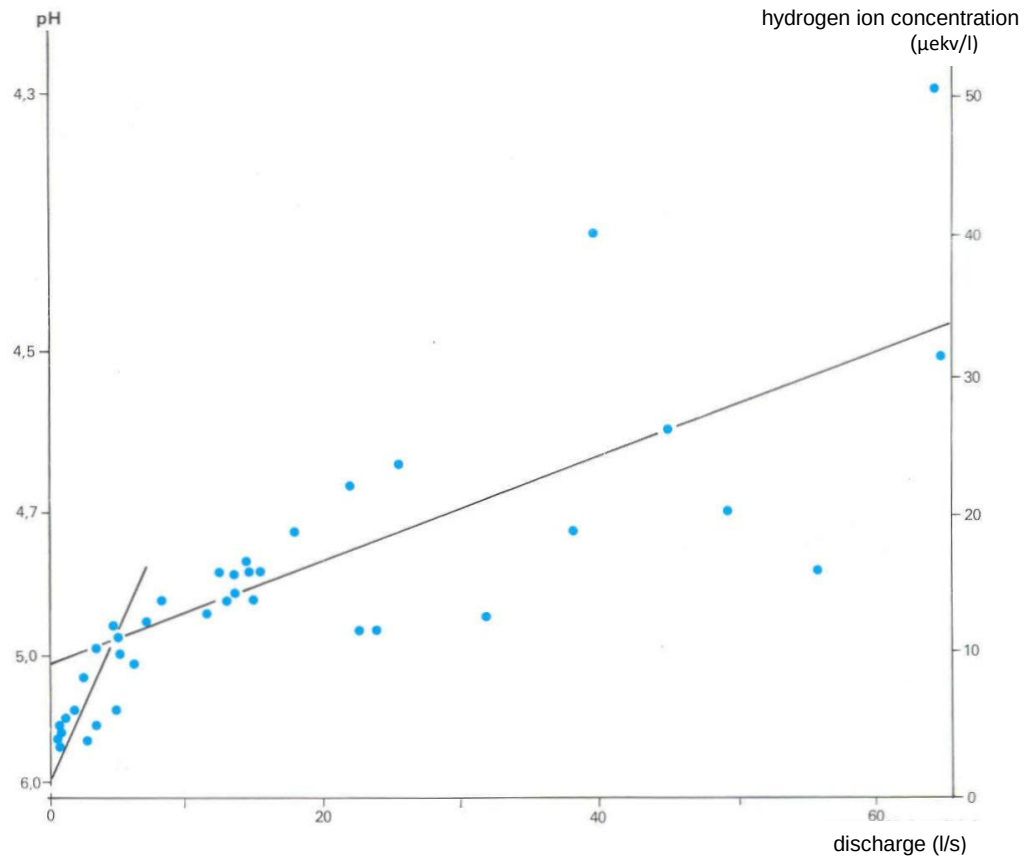


Fig. 54 The relation between runoff and hydrogen ion concentration in the Buskbäcken stream, Kloten area, Central Sweden, from 24 May to 23 June 1970. When the runoff is small the streamwater consists of groundwater that is discharging from considerable depth through the bottom of the stream. This water has a high pH-value. When the runoff is large shallow groundwater with low pH-value is dominating.

The concentration of organic nitrogen in a stream is highest during the summer. At the same time the biological demand for inorganic nitrogen is large and the concentration of nitrate- and ammonium nitrogen is lowest during this time.

One point in time – variation along the stream

In the uppermost part of a stream the water has the highest concentration of hydrogen ions and free aluminium ions. As we go downstream we will find that the concentration of these acid cations decreases, while the concentration of basic cations (calcium, magnesium, sodium and potassium) and dissolved silicon increases. Increasing residence time for the discharging groundwater further down means longer time for the groundwater to take part in slow mineral transformations. It has been shown that more basic cations are released than acid cations are consumed. This shows that there must be still other sources of acid, probably organic acids produced by biological reactions.

In Fig. 38 it was demonstrated how the electric conductivity increased along the stream. The rate of pH increase decreases down the stream, but it is a common observation that pH increases down along a watercourse until it reaches the sea. The increase far down in a drainage system probably depends on changed geological environment, i.e., for example clay deposits and on anthropogenic activities as fertilisation and sewage efflux.

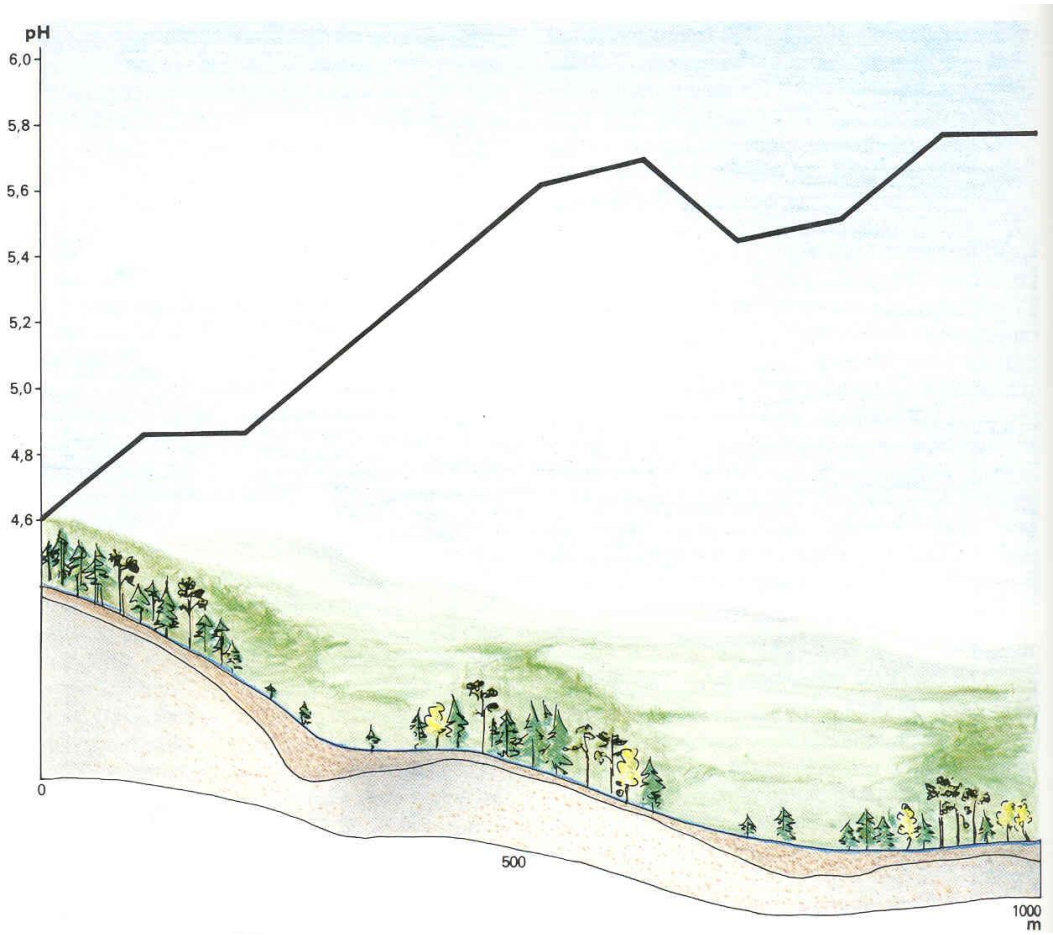


Fig. 55 The pH-value measured along the first 1000m of a stream (Kullarna nedre) in Hälsningland, Central Sweden 23 June 1980. The increasing trend of the pH-values along the stream indicates the increasing flow path lengths of the discharging groundwater and is only broken when the stream passes a mire area.

In Fig. 55 we can see how the pH-value increases downstream in a watercourse, i.e., how the hydrogen ion concentration decreases. The hydrogen ion concentration decreases fastest at the top of the stream and after less than one kilometre the pH-value is fairly constant.

On its way down through the landscape a stream experiences a variety of discharge areas. In convex hillslopes the discharge areas are small and seldom have a thick organic soil layer. At the

base of concave hillslopes, on the other hand, a thick layer of organic material may have developed. Here the discharge areas are large and precipitation that falls on these areas will not be neutralised by passing a mineral soil layer before it enters the stream. This might explain why the pH-value decreases in the right part of Fig. 55.

The chemical composition of streamwater is also altered by biological activity along the stream. At forest fertilisation with Urea in the Kloten-area it was observed that more than 80% of the nitrate that leached into a stream in the fertilised catchment had disappeared 500 m further down along the stream. About half of this was found as dissolved organically bound nitrogen, but the remainder was not recovered. Probably this had been taken up by shore- and water vegetation.

Obviously, the composition of the streamwater in different ways reflects the water's pathway through the catchment. Knowledge of the pathways of water from precipitation to stream can give a better understanding of how water quality arises. Knowledge of the chemical composition of water and its temporal and spatial variation gives us, in a similar way, an understanding of the flowpaths and residence times of water in a catchment.

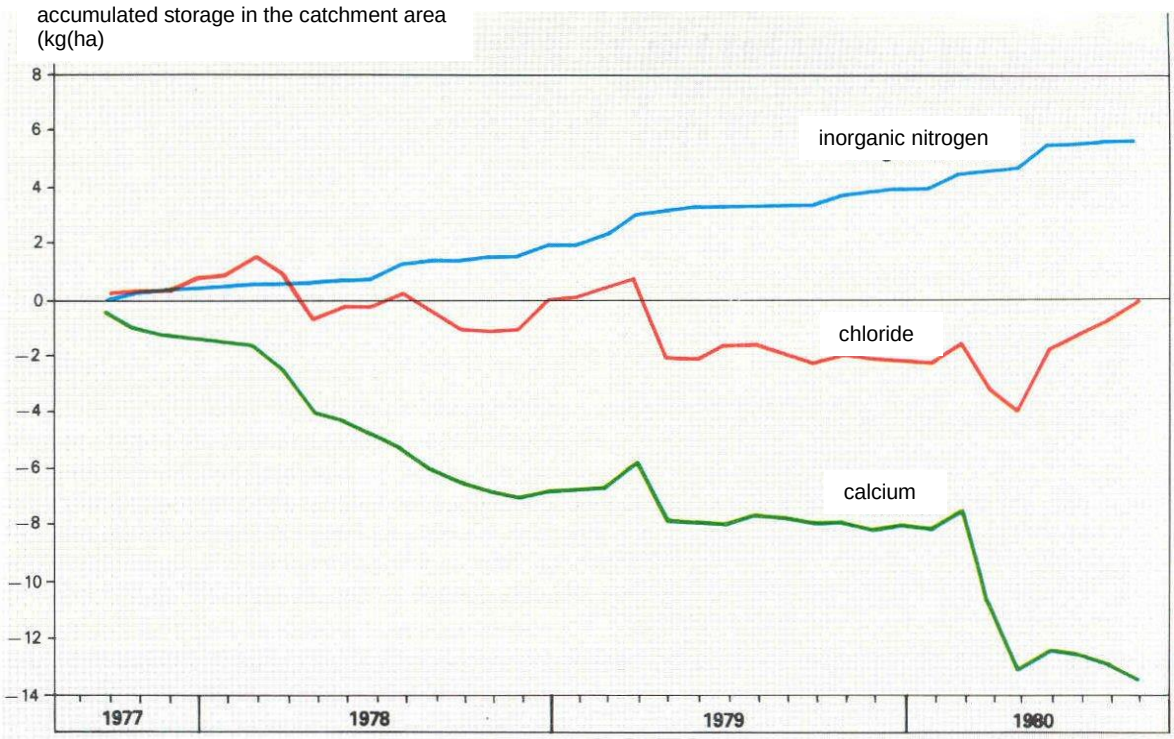


Fig. 56 Storage of different substances in a catchment as calculated from input and output observations. Nitrogen accumulates in the growing biomass of the area, while calcium is released by weathering and is leached. Chloride neither accumulates nor leachs. Data from catchment studies in Hälsingland, Central Sweden.

The storage of many substances in a catchment changes over time

The input of water to a catchment equals the output if a fairly long period of time is considered, say a decade or more. The water storage in the catchment varies during a year and also between years, but if calculated for many years the change in storage is small compared with the volume that has been added or removed. A similar balance between input and output is only valid for some of the chemical compounds that take part in the turnover of water. The compounds that in dissolved form take part in the water flow through a catchment can be divided into three groups according to their appearance. The first group consists of elements having their main source in deposition from the atmosphere (dry and wet deposition) and that do not to any larger extent take part in reactions within the catchment (for example chloride). The second group consists of elements having their main source in deposition from the atmosphere but are actively fixed in the catchment (for example ammonium- and nitrate nitrogen that are taken up by vegetation). The third group consists of elements which to some degree are supplied by deposition but mainly come from weathering in the catchment (for example sodium, calcium, magnesium, phosphorous and aluminium). The deposition of elements belonging to the first group equals the leaching of them if seen over long enough time (Fig. 56). The storage in the catchment may be considerable during some years. The input as deposition of elements in the second group is, if seen over a longer time period, larger than the output, while the input of elements in the third group is smaller than the output.

Besides elements in these three groups there are some elements with an intermediate position. Many investigations during the 1970's and 1980's have shown that sulphate is fixed in the catchment. Later investigations during the 1990's when the sulphate deposition has decreased have seen a net loss of sulphur from catchments. This indicates that the soil was moving from one steady state situation to another due to the increased atmospheric deposition and then again is moving backwards to an earlier steady state when the deposition has decreased. The fixation can take place in discharge areas with thick mor layer and where anoxic conditions are often present.

In the luxuriant vegetation in the foreground intense groundwater discharge takes place. A well is situated on the farmland in front of the cow-house. Its pressure surface is situated a few decimeters above the soil surface. The location of the well below the cow-house is inappropriate, but any contamination of its water has not been found. The recharge area upstream from the well consists of a narrow till ridge where the dwelling-house is found. Probably the well also gets groundwater from more distant recharge areas through cracks in the bedrock. Koxbacken, Västmanland, Central Sweden.
Photo: Harald Grip.

8. Water – a solvent passing through the landscape

We have followed the pathways of water through an area down to the draining stream and sometimes a distance along the stream. There the subject for this book is finished. The processes that govern for example the propagation of floods and their growth along watercourses and the water quality changes during this continued travel should be treated somewhere else. We stop before the stream grows to a river and before lakes are introduced.

In this final chapter we will apply the physical and chemical base from earlier chapters on some practical questions. Our intention is not to treat the problems comprehensively, but to elucidate them outgoing from the ideas we tried to give you about the pathways for water to the stream. We hope that it can give ideas of how other problems about the existence, flow, and chemical composition of water could be solved. Even if data are incomplete it is possible to come up with good estimates if the water problem is put into a general picture of the water flow through the catchment.

Forest and water

In a forest there is a clear interaction between water and production. On one hand the growth is dependent on soil water availability. On the other, forests have a large influence on this water and by that also on the streamwater. Let us look at this interaction in the light of water pathways to the streams.

Water and nutrients in the root zone govern forest production

For a tree to grow the roots need water, nutrients, and oxygen. The water is mainly used for transpiration (cf. page 43). It also acts as transporters of the nutrients the trees use in its growth. The oxygen is needed for root respiration.

In a hillslope the best growth is normally found close to the boundary between recharge and discharge areas. There the roots have the most favourable access to water, nutrient and oxygen. Let us see how these factors vary along a hillslope and how this variation affects forest production.

In the upper recharge area the trees have to rely on water stored in the root zone at infiltration events. In the lower parts of recharge areas and in discharge areas the trees can also use groundwater that is supplied from upslope. If the soil layer is not too thin the storage in the root zone is normally sufficient to support some transpiration, i.e., the trees can live on local precipitation without wilting. But even if the soil is thick the local water resource is often limiting growth, because the trees will close their stomata long before they wilt. Then the transpiration decreases, but also the carbon dioxide uptake and by that the growth. In large parts of Sweden the soil is thin on hilltops and the vegetation will have difficulties to survive long drought periods, even if the water holding capacity of the soil as such is sufficient. Thus, it is the topography in combination with the soil depth distribution over an area that gives drought in elevated areas during some summers.

Water also has a considerable significance for nutrient availability. It is the water that transports the nutrients to the roots. This transport occurs in different ways. Nutrients are transported by the water flow that the root water uptake creates. Nutrients are also transported by molecular diffusion in water. When a root takes up a certain ion, the concentration of that ion decreases in the vicinity of the root. The concentration gradient starts a diffusion flux of that ion along the

gradient. But the diffusion is slow and delivers such ions only from short distances from the root. In recharge areas without lateral water flow the roots enhance their nutrient uptake by actively searching for the nutrients. The roots are continuously growing and penetrate new soil volumes where unexploited nutrient resources may be present. In lower parts of the recharge areas and in discharge areas with lateral or upward water flow in the root zone, the roots are in a better situation. There they can take up the nutrients when they are passing with the water flow. In this way the nutrient uptake is very efficient. In the laboratory it has been shown that tree plants can reach maximum relative growth rate in practically deionised water if the water flux gives enough supply of essential elements. It is not the concentration of nutrients in soil water that determines nutrient availability, but the number of ions that reach the roots per time unit. For forest trees mycorrhiza fungi hyphens are important because they extend the root system and add accessible soil volumes from where nutrients can be taken up. It has also been shown that the hyphens can penetrate mineral grains and take up basic cations and phosphorous from within the minerals and transport them to the tree roots without exposing them to the free soil solution.

Even if the nutrient availability often is good in discharge areas, the tree production may be limited by oxygen deficiency in the root zone. The gas exchange with the atmosphere is hampered by the high water content of the soil. In this case water is a negative factor for the forest. It is assumed that the air content in the soil (soil air content = porosity – water content) should be at least ten percent for the roots to maintain their normal respiration. Close to the boundary of the discharge area the upper groundwater may be rich in oxygen because the water has only been a short time in the groundwater zone. There the trees have the best access to water, nutrients and oxygen

The groundwater level increases after clear-cutting

The transpiring surface decreases after clear-cutting. Also the interception storage decreases. Therefore both transpiration and evaporation decreases when an area is clear-cut. In our climate this leads to increased infiltration, increased soil water content and increased groundwater recharge. The groundwater surface rises, the discharge areas get larger, and the mean runoff increases. In two Swedish investigations it was found that the runoff increased from 540 to 720 mm per year and from 260 to 560 mm per year, respectively, i.e., increases with 180 and 300 mm per year, respectively.

After clear-cutting, the shallow groundwater levels and the large discharge areas lead to faster runoff response to precipitation and higher runoff peaks in the streams. The large runoff peaks is not due to Hortonian overland flow, but rather to the higher groundwater level causing less damping in the unsaturated zone. The clear-cut area has a higher ability to rapidly generate runoff than a forested area. Also the low flow increases after clear-cutting due to the reduced evapotranspiration.

After forest re-growth on the clear-cut hillslope the transpiration increases, the groundwater level becomes lower and the runoff returns to its initial mode. The increased groundwater level on a clear-cutting may obstruct forest regeneration on lower parts of the hillslopes. The roots of the young plants cannot respire due to too high water content and low air content in the root zone.

Clear-cutting is an example of how an action in the recharge area influences on the conditions in the discharge area. When the groundwater recharge increases in a clear-cut recharge area, the groundwater level in the discharge area increases and oxygen deficiency may develop in the root zone reducing forest production there. Even if only the discharge area is clear-cut it may lead to forest regeneration problems. As long as a mature forest was growing there was a balance

between the water flow from upslope and the evapotranspiration, and the root zone was kept aerated. Clear-cutting the unsaturated discharge area may transform it to a saturated discharge area. New plants are not able to grow to the size that they can cope with the water flow from upslope. The problem is especially severe on fine textured soils, where the groundwater flow needs such a large cross sectional area that the soil surface often gets saturated.

The discharge area around the stream is seen as a green band in the clear cut area in late summer. Kilmyrbäcken, Jämtland, Central Sweden (the same stream as in Fig. 37 and 38). Photo: Ulla Maria Calles.

Discharge areas grow after clear-cutting

With Darcy's Law and Dupuit's assumption we can analyse how the boundary between recharge- and discharge areas is displaced upslope in a clear-cut hillslope.

Consider a straight forest hillslope. The till overburden on a watertight bedrock is two meter deep. The groundwater discharge begins at a distance from the water divide where the groundwater recharge above is equal to the largest possible groundwater runoff, i.e., the groundwater runoff when the groundwater surface coincides with the soil surface (cf. the discussion on the role of topography at page 75).

$$R \cdot x \cdot b = - K \cdot m \cdot b \cdot dh/dx$$

R = groundwater recharge
x = distance from the water divide to the boundary of the discharge area
b = considered width of the hillslope
K = saturated hydraulic conductivity of the soil profile, 10^{-5} m/s
m = thickness of the soil profile, 2 m
dh/dx = maximum slope of the groundwater surface, that is, the slope of the ground surface, -0.1 m/m

We assume that the groundwater recharge in the area is 350 mm per year, i.e., R = 350 litre/(m^2 · year). The equation is solved for x and we find that the boundary for the discharge area is 180 m from the water divide. If the area is clear-cut and the evapotranspiration decreases by 200 mm per year, the groundwater recharge increases to 550 mm per year (350 + 200). The boundary for the discharge area will then be at 110 m from the water divide. In this example the result of the clear-cutting was that 70 m of the hillslope was transformed to saturated discharge area.

Nutrient leaching after clear-cutting

The continuous accumulation of nutrients in a growing forest stand depletes the soil from exchangeable cations. The vegetation also keeps a lot of nutrients cycling. They are taken up by the roots of the vegetation and are incorporated mainly in needles and leaves, which ultimately falls to the ground as litter. In the mor layer they will be released by microbiological processes and will again be available for plant roots.

After clear-cutting large amounts of litter as twigs and needles are added to the mor layer. Often the *mineralisation* increases due to the sudden increase in available litter. This decomposition of organic material is a microbiological process where organic material is decomposed into carbon dioxide, water and mineral nutrients. It is favoured by increased soil moisture as long as the oxygen is not limiting. During summer the soil temperature is higher on a clearcut, where the soil is not shadowed by trees, than in the forest. This also promotes mineralisation. Before grasses and other vegetation have established on the clear-cutting there are few living roots that can take up the released nutrients. Some of these will therefore be leached from the soil profile and transported away with the discharging groundwater. Sometimes water in pools and ditches becomes almost black from dissolved organic carbon.

In connection with the mineralisation of organic bound nitrogen ammonium ions are formed. In the cation exchange complex of the soil these positive ammonium ions may be exchanged for hydrogen ions or basic cations and they can be leached from the soil profile. If the ammonium nitrogen is oxidised, so called nitrification, the negatively charged, and therefore easily moving, nitrate ions may be leached. The nitrification, which is also a microbiological process, is favoured by good oxygen supply and high pH-value. When the nitrate ion is leaching it brings with it the same amount of cations. The maximal effect of nitrate leaching may be observed in streams and springs already after one or a couple of years, but it may also be delayed up to ten years.

It is discharge areas that need to be drained

If low-lying parts of forest land are paludificated, i.e., bogs expand due to rising water tables as a consequence of peat growth after clear-cutting, it may be necessary to lower the groundwater level by ditching to create an unsaturated soil horizon where new plants can develop their root systems. What is needed is a temporal help. When the new stand is established and the new canopy starts to close, the trees will be able to keep the root zone aerated. The increased

evapotranspiration will keep the groundwater level down. In some cases it may even be worthwhile to ditch in growing stands.

A ditching the groundwater level is lowered so that it close to the ditch coincides with the water surface in the ditch. The distance around a ditch that is influenced depends on the hydraulic conductivity of the soil. The larger it is the larger is the distance from the ditch that the groundwater level is lowered. In till soils, with its relatively low conductivity, the ditching will only influence the groundwater surface close to the ditch.
The effect of ditching depends on the orientation of the ditches compared to the slope direction. A ditch dug along a till hillslope, going from the top and down, will only lower the groundwater surface close to the ditch. A ditch perpendicular to the slope direction will give better result. On the upslope side of such a ditch, the groundwater surface is only influenced in the area just upslope the ditch. At the downhill side, on the other hand, the groundwater surface is lowered also at longer distances from the ditch, since the ditch diverts a large part of the groundwater flow from upslope. The groundwater surface again reaches close to the ground surface when enough water has accumulated below the ditch, by local groundwater recharge and discharging groundwater from upslope.

Even if the groundwater surface has been lowered to the bottom of the ditch it is not for sure that the water content in the upper soil horizons decreases enough to give good conditions for a young plant to establish. The decrease in water content depends on the soil profile's water content at drainage equilibrium, i.e., on its water retention curve.

The saturated discharge area decreases by ditching. A larger proportion of the infiltrating water from precipitation will then pass through mineral soil horizons before it discharges into the stream. An effect of ditching mineral soils is often that the pH-value in streamwater increases (fig. 57).

Peat soil ditching

The reason for ditching peat land, as for mineral soils, is to lower the groundwater level, most often to increase forest production, but also to enable peat mining. There are two ways to do the ditching. One way is to decrease the inflow to the mire by diverting inflowing water. Another way is to drain water from the mire by a dense ditching network on the mire itself. In the case of a fen (cf. Fig. 44), it may be enough to cut off streams and shallow groundwater inflow, but if inflow of deeper groundwater directly to the fen is substantial, water must be drained by drainage ditches. In a bog, the groundwater level can only be lowered by drainage ditches.

The hydraulic conductivity of peat is highly dependent on the degree of humification (cf. page 79). In the upper living horizon of a mire the conductivity is large, but deeper down in the profile it can be very low. Therefore it is easy to reach a first lowering of the groundwater level, but for a further lowering a dense drainage network is needed. A distance between ditches of 10 – 20 m may be needed. In practise this distance is estimated from the mire vegetation, its degree of humification and the local climate.

When the groundwater level is lowered in mire, the evapotranspiration decreases since the surface gets drier and the total runoff from the mire is increased. Then, if the mire is afforested both evaporation and transpiration increases and the runoff decreases. The unsaturated zone that is formed above the groundwater zone after ditching has a damping effect on runoff because infiltrated water temporarily may be stored there. During rather dry conditions, the ability to generate runoff will thus decrease. If the precipitation is so large that the unsaturated storage is

filled, the runoff peaks will, on the other hand, be larger than before the ditching since the mean distance to drainage channels has been decreased by the ditching.

Fertiliser on discharge areas goes directly to the stream

When forestland is fertilised by nitrogen fertiliser, the part of the fertilizer that falls on drainage channels will run off directly. Also fertiliser that has fallen on discharge areas may soon be washed off if it is not fixed into the organic material. After a forest fertilisation in the Kloten area (Central Sweden) the nitrate fertiliser that was discharged during the first 20 days was equal to the amount that had fallen in and within two metres from the stream.

Even though there are large amounts of nitrogen bound in a forest soil, there is often nitrogen deficit and the vegetation takes up the added fertiliser nitrogen. In a fertiliser experiment with 150 kg nitrogen per hectare as ammonium nitrate still only 21% of the nitrogen was found in the soil and 19% in the needle biomass. It was further estimated that 20% – 30% was taken up by bushes and shrubs. In another experiment with the same amount of fertiliser 4% of the fertiliser was lost to the stream and only 1% was found as nitrate in soil- and groundwater. If the results from these experiments are combined the fate of about 30% of added nitrogen still remains to explain. One explanation may be that the nitrate nitrogen was lost to the atmosphere as nitrogen gas or nitrogen gas oxides.

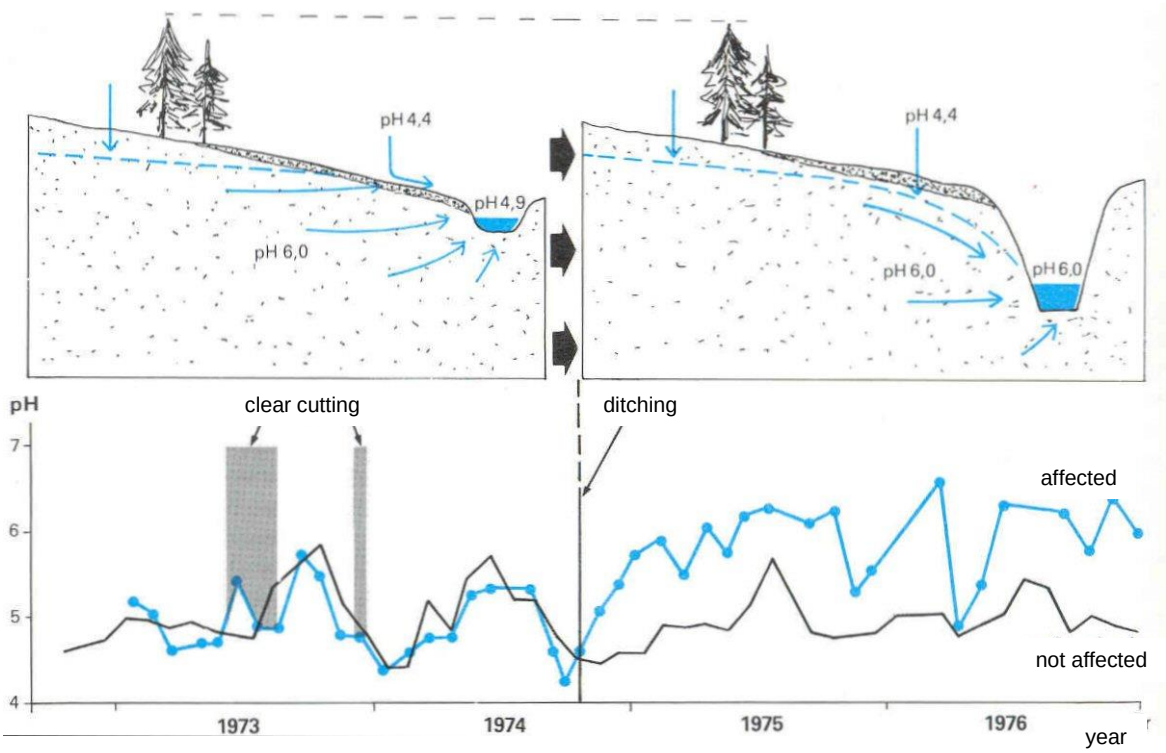


Fig. 57 The effect of a treatment can be identified by comparisons with a reference object, before and after a treatment. Clear-cutting a catchment in Kloten did not influence on the pH in the stream. After ditching the pH increased about one pH-unit. The upper part of the figure gives a suggestion for explanation of the phenomenon. Before ditching a part of the water reaches the stream without having been neutralized by weathering in the mineral soil. After ditching, all water passes the mineral soil.

Pathways for the fertiliser surplus from arable land

Farming in many ways implies more drastic changes in the soil water conditions than forestry. Soil management, drainage and sometimes irrigation are means to optimise the soil water

conditions in the root zone. Here we will only pay attention to one aspect on water in agriculture, water as a transporter of fertiliser.

If more fertiliser is applied to a soil than current vegetation is able to take up, the surplus may be leached from the soil profile. The fate of the leached fertiliser depends to a large extent on the hydrological situation, i.e., if the area is a recharge- or a discharge area.

Agriculture on recharge areas

The till clays in for example Skåne and Östergötland, Southern Sweden, and many soils with low clay content are, due to the topographical settings, often recharge areas. The risk for fertiliser leaching is especially high on these light soils. Leached nutrients will contaminate the groundwater if they are not taken up by the crop or microorganisms, or are transformed in the unsaturated zone.

Denitrification, i.e., reduction of nitrate to nitrous oxide gas or nitrogen gas, is an example of such a transformation. Nitrogen may be transferred to the atmosphere by this process. Microorganisms at anoxic conditions mediate the transformation when organic substrate is present in the soil. The pH-value of the soil should be within the limits 4.5 and 10 and the process is most efficient at pH = 7.5. The process is favoured by high temperature. Also in a soil profile with good aeration there may be some soil aggregates within which there are oxygen deficiency and by that prerequisites for denitrification. There may be anoxic conditions in a water saturated soil aggregate of only a few cm in diameter.

During the last decades high nitrate concentrations in groundwater have been observed in many areas where agriculture is practised on recharge areas, for example in Halland (Southwest Sweden), Gotland (East Sweden) and on the plain of Kristianstad (Skåne, South Sweden). Obviously the denitrification process has not been of such extent that the fertiliser surplus has been transformed. During the 1980's, 15% of the municipalities in the plains of Götaland (southern part of Sweden) had a median nitrate concentration above 30 mg/l nitrate in private wells. From a hygienic point of view these water supplies are considered "noteworthy".

Agriculture on discharge areas

A large part of the Swedish arable land is on clay soils. These are located in low parts of the landscape and are often typical discharge areas. By means of subsurface drainage much of this land has been transferred from saturated to unsaturated discharge area when the discharging groundwater was diverted to the drains. The water in these drains consists of both groundwater that infiltrated on the surrounding slopes, typically forested till slopes, and of water that infiltrated on the farmland. During dry periods the water quality in the drains is dominated by the groundwater from the forested slopes. During wet periods, when water is percolating through the arable soil profile, the water in the drains may show a strong influence of agricultural activities, notably nitrate leaching. In agricultural areas on discharge areas, streams get the fertiliser surplus directly via the drains, resulting in increased eutrophication of lakes and streams.

Acid rain

Combustion of fossil fuels increased wet deposition of acidifying compounds until the beginning of the 1970's. Conversion to other fuels and to flue gas cleaning has decreased the deposition of sulphate, but the deposition of nitrate and ammonium is still increasing (cf. p 100). Today the pH-value in the precipitation is about 4.2 in Southern Sweden and about 4.5 in the north. The

effects of the acid deposition are to a large extent determined by the water flow pathways through the landscape. Measures to counteract the effects must be planned with focus on water flow and the different functions of the different parts of a catchment. In this section we will give some aspects of the different functions of recharge- and discharge areas in this respect.

Acid rain releases metal cations from the soil's cation storage

We have earlier pointed out that the mor layer in a forest soil and the dead organic material in the root zone is a considerable storage of metal cations, a part of which exists in exchangeable form. The storage amounts to about 50 year's deposition from the atmosphere. We also stressed that the mor layer smoothes out short term variations in the composition of the infiltrating precipitation.

If the cation content in precipitation changes to a new level, the ion exchange system of the soil will eventually equilibrate with the changed input. Such a shift has slowly taken place as a consequence of the increased hydrogen ion concentration in precipitation (the pH-value has decreased). Chemical analysis of the mor layer on a large number of places in Central Sweden indicates that the amount of exchangeable calcium, magnesium and potassium decreased during the period of heavy acid deposition and that the amount of exchangeable hydrogen- and aluminium ions increased. If this is the case, the mor layer has been acidified. Since the acid deposition has decreased during the last decades the soil is recovering.

When a recharge area experiences acid deposition and the mor layer acclimates to this situation, basic cations are released to the same extent as acid cations are incorporated. The released cations that are not taken up by the roots leave the root zone with the percolating water. In the next stage, when the mor layer is in equilibrium with the acid precipitation, no further net release of basic cations takes place. The percolating water will then be as acid as the precipitation. Even if the weathering rate does not increase, the weathering in the unsaturated zone will increase and the weathering front will be adjusted downwards in the soil profile. If the soil is thin and the flow pathways for groundwater are short, it is possible that aluminium that was released by weathering will not be fixed in deeper soil layers. We then get acid groundwater containing aluminium.

Is the soil acidified by forest growth?

When we discussed the weathering of silicate (p. 111) we showed that hydrogen ions were consumed in the reaction. Normally carbonic acid delivers the hydrogen ions that are consumed in the weathering. For every hydrogen ion that is consumed in this way one bicarbonate ion is formed (cf. formula 3, page 107). To be able to move with the water, released basic cations must be accompanied with an equal amount of negative ions (anions), in this case bicarbonate ions. Because the root zone is, as we have shown, an open system regarding carbon dioxide, the carbonic acid concentration will not change, neither by weathering or percolation of bicarbonate ions accompanying the released basic cations.

In the case that a tree root delivers the hydrogen ion that is needed, and in exchange takes up the released basic cation from the weathering silicate, neither the basic cation nor the hydrogen ion appears in the soil water. Since the weathering rate seems constant, the leaching of basic cations and bicarbonate ions must decrease due to basic cation uptake by roots. As long as the root uptake of basic cations is limited to an exchange with weathering minerals, no acidification of the soil or the percolating water takes place.

It is also possible that the trees take up basic cations from the ion exchange storage in the soil, notably from the mor layer. Here the sources for basic cations are litter, i.e., from the weathering primary minerals, and wet and dry deposition of salt from the atmosphere. When the trees take up basic cations from the exchange storage, hydrogen ions will be added to this storage in exchange. The hydrogen ions have in this case no movable anion available and therefore stay in the mor layer thereby increasing its hydrogen ion saturation.

Basic cations in neutral salts from the atmosphere may in turn be exchanged into the exchange complex of the mor layer, and hydrogen ions from last step are released. These hydrogen ions will then be accompanied by the anions from the salt, i.e., anions from strong acids (Cl gives hydrochloric acid, SO_4^{2-} gives sulphuric acid and NO_3^- gives nitric acid). The strong acids that are now percolating through the soil profile cause a displacement of the carbonic acid system, and bicarbonate ions and hydrogen ions give carbon dioxide and water. The basic cations that were accompanied by bicarbonate ions in the percolating water are now accompanied by strong acids anions.

Nitrate- and sulphate ions are partly taken up by the tree roots since they are nutrients. They are accompanied by basic cations. Sulphate- and nitrate ions may also be reduced in anoxic environment and by that disappear from the water phase. In such case, equal amounts of basic cations are fixed in the soil.

The uptake of basic cations from the soil's cation exchange system by the trees thus is an acidification of the soil. If neutral salts are added from the atmosphere the percolating water will be less rich in bicarbonate and therefore more sensitive to acidification than it would have been otherwise.

Lower pH-values in the mor layer have been observed in old pine forests than in younger. In the Kloten area it has also been observed that the most acid runoff in streams (i.e., conditions at high flow) was more acid the older the tree stands that were growing in the catchments. One suggested explanation is that the dry deposition of acidifying substances is larger in tall stands, which more effectively clean the air, than in small stands, but more important is probably the higher hydrogen ion saturation of the mor layer in old stands.

Plants that get their nutrients exclusively from the atmosphere, as for example *Sphagnum*-mosses on bogs, contribute directly to acidification of streamwater by accumulating basic cations from salts deposited from the atmosphere in exchange for hydrogen ions. In this way they release strong acids, which in this case are not neutralised by weathering, but is directly delivered to streamwater.

Soil resistance to acidification decreases successively

Many buffer systems act in the soil and limit the pH-changes that would occur if hydrogen ions were added. The most important system is the carbonic acid system (carbon dioxide – bicarbonate – carbonate). This system is active at pH-values down to about 4.5. An acidification of water can often be identified more clearly as a displacement in a buffer system than as a change in pH-values. The restricted data that are available today does not show a general decrease in pH of groundwater in Sweden. On the other hand, data show that the resistance against pH-changes in groundwater has decreased, especially in South and Southwest Sweden. There are examples from wells where the bicarbonate concentration has decreased by more than 50% during the 1970's. The time series are, however, short and it could be temporary variations due to the dry years in the middle of the 1970's.

The acid precipitation also contains high concentrations of sulphate and nitrate. In almost all Sweden nitrate is a welcome nutrient for the vegetation. In most cases ecosystems in our country are nitrogen limited, i.e., nitrogen is the most limiting nutrient for plant growth. This does not hold true in some areas in south Sweden where nitrogen starts to exist in surplus and where nitrate is leaching to the groundwater.

The negatively charged sulphate- and nitrate ions move readily in the soil and bring cations when they follow the water through a catchment. If anoxic conditions prevail in the soil the sulphate, as we have stated, may be reduced and deposited as sulphides. In many investigations during the 1980's it has been found that the sulphur input was larger than the sulphur output via runoff. The sulphur storage in the soil is very large compared with the yearly deposition. Probably the sulphur storage was then increasing towards a new equilibrium corresponding with the high input. Now the sulphur deposition has decreased to less than half of its maximum value, and the sulphur export is larger than the deposition.

Acid streamwater at high floods

A general observation is that the pH-value of a stream decreases when the discharge increases. In many investigations very low pH-values have been found in connection with flow events. This phenomenon, called "acid surge", has above all been observed in connection with spring snowmelt or in floods after a longer period of low flow. How are the low pH-values in peak discharges connected with our statement that even flow events are dominated by groundwater?

The dry and wet deposition of acids on saturated discharge areas are more or less directly transferred to the streams and contribute directly to acid runoff. The same holds true for deposition on unsaturated peat land, where groundwater is formed without percolation through a mineral soil. But also the groundwater from mineral soils that contributes to flow events may be relatively acid. Earlier we have stated that the groundwater pH-value within an area varies a lot depending on the flow pathways for the different water particles and their residence time in different environments. The deep groundwater has a rather high pH-value, maybe around 6, or higher, the shallow often very low, around 5 or lower. In connection with flow events the proportion of shallow groundwater in the total groundwater discharge increases and the pH-value of the groundwater that contributes to the streamflow decreases. The groundwater that is formed when the groundwater level reaches the former unsaturated root zone has especially low pH-value. This earlier soil water probably constitutes a great proportion of the discharging groundwater at flow events (cf. page 94).

Depending on which pH-value the groundwater has and on its other chemical composition, different chemical reactions occur when the water comes in contact with the atmosphere and when it mixes with other water in the discharge areas.

We mentioned earlier that degassing of carbon dioxide to the atmosphere might give an increase in the pH-value of the water. The concentration of bicarbonate in groundwater is important when it is mixed with acid precipitation. If the bicarbonate concentration is high, a large part of the acids in precipitation will be neutralised. The mixture of groundwater and rainwater in the stream will therefore get a higher pH-value than what is indicated from the hydrogen ion concentrations and the relative volumes of respective water would give.

When discharging groundwater gets in contact with the atmosphere also reactions that tend to decrease the pH-value may occur. In anoxic groundwater iron is in the reduced form as iron (II).

When such groundwater gets in contact with oxygen in the discharge area the reduced iron oxidises it will be deposited as iron hydroxide. In this reaction hydrogen ions are produced and the pH-value decreases.

In silicate-weathering we have seen that among other things aluminium hydroxide ($Al(OH)_3$) is formed. In very acid environment (pH-values less than 4.5) free aluminium ions (Al^{3+}) and water are formed from the aluminium hydroxide and hydrogen ions. These aluminium ions may be transported down in the soil profile, but when the water reaches a less acid environment, for example the groundwater, aluminium hydroxide precipitates and new hydrogen ions are released. In this way aluminium ions replace hydrogen ions as transporters of acidity in acid environments. If there were no aluminium hydroxide, the water would have transported the hydrogen ions instead of the aluminium ions. In the same way the aluminium ions can transport acidity from the unsaturated zone to stream when the groundwater rises and a lateral flow occurs close to the soil surface.

Acid surges after a long time of dry weather is also assumed to be due to wash out of sulphuric acid and other sulphates oxidised from reduced sulphur when the groundwater level was low. These acid surges have been observed especially in acidified areas.

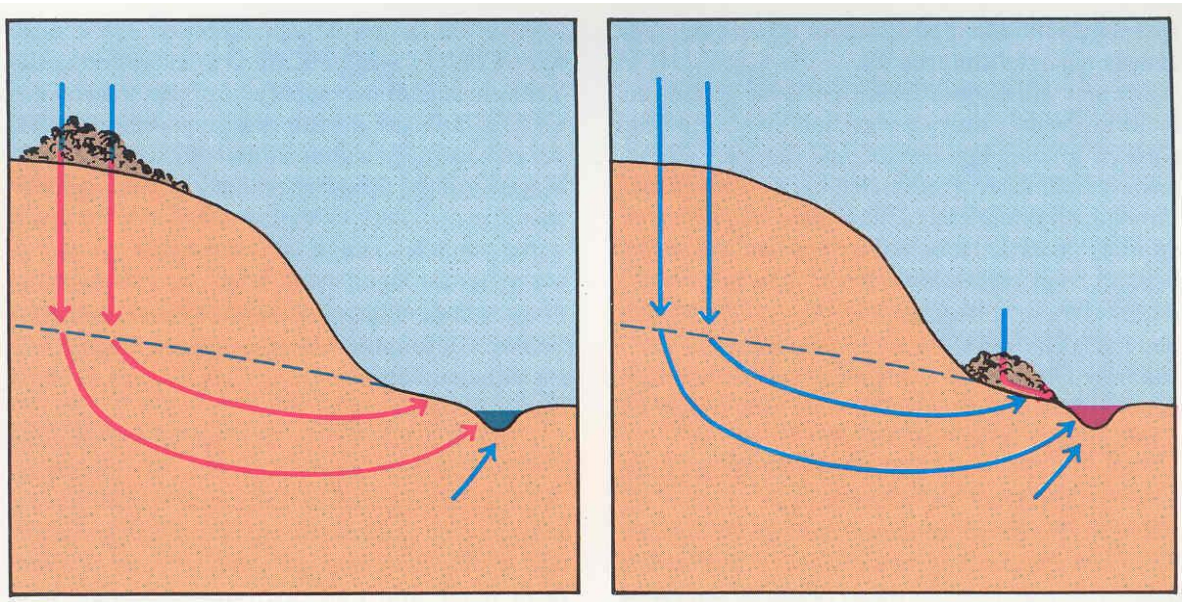
A phenomenon that contributes to acid surges in the spring is the re-crystallisation that takes place in the snow pack during winter. Water vapour is transported from small to large ice crystals, and then soluble contaminations are concentrated on the shrinking crystals while they are diluted on the large. The contaminated crystals have lower melting point than the clean ones and therefore salts, as well as acids, have a tendency to leave the snow pack with the first meltwater and give extra acid meltwater at the beginning of snowmelt.

All processes mentioned above give acid streamwater when the groundwater level is high or rising, i.e., when the runoff is large or increasing. The example with acid surges shows how important it is to know the water flow pathways and the chemical processes acting along them to explain observed water composition. The chemical reactions that occur when different water masses mix should also be considered.

Where in the catchment should a waste deposit be localised?

Now and then the news are reporting on waste deposits that have destroyed groundwater or surface water. Would it have been possible to avoid contamination of the water

Fig. 58 The choice of waste location is of great importance for the spread of contaminants. Waste that is decomposed in the unsaturated zone of the soil may be deposited in a recharge area. Here is shown how waste that does not decompose may influence the water quality.



by choosing an alternative location of the deposit? Is it possible to use our knowledge about how water is moving from recharge- to discharge areas to find the optimal location for different wastes?

If the waste is deposited on a recharge area, the leachage water infiltrates into the soil. Some types of contaminants are deposited or mineralised in the unsaturated zone, while others will be transported downwards with the water to the groundwater. The groundwater downstream from the deposit will get contaminated and ultimately the surface water in the discharge area. Only the deeper groundwater that infiltrated far upslope remains unaffected.

If the waste is deposited on a discharge area we lose the cleaning that may take place in the soil. But we will not get affect the groundwater and the contamination may be easier to observe. The water in drainage ditches will be a mixture of leachage water and clean groundwater.

Neither of these locations seems to solve the contamination problem in an optimal way. One therefore tries to deposit the waste in such a way that the leached water is possible to collect and purify. The purification process is simplified if there are small amounts of water that should be treated. The waste should therefore be placed in such a way that the water flow from the surroundings is small, i.e., the catchment should be small. One also tries to decrease the amount of leached water by minimising infiltration into the deposit.

In order to make it possible to collect the leached water, the waste on a recharge area should be stored on a watertight bed with a single outlet so that all drainage water is discharged from the bottom of the deposit as if there were a spring. Completely watertight beds for waste deposits are hard to find. Instead an upper limit of 10^{-7} m/s for the hydraulic conductivity of the bed is recommended. This low conductivity is still enough to infiltrate 10 mm per day. A large part of the rain or meltwater that infiltrates the waste deposit will therefore percolate down to the groundwater instead of allowing being collected at the base of the deposit.

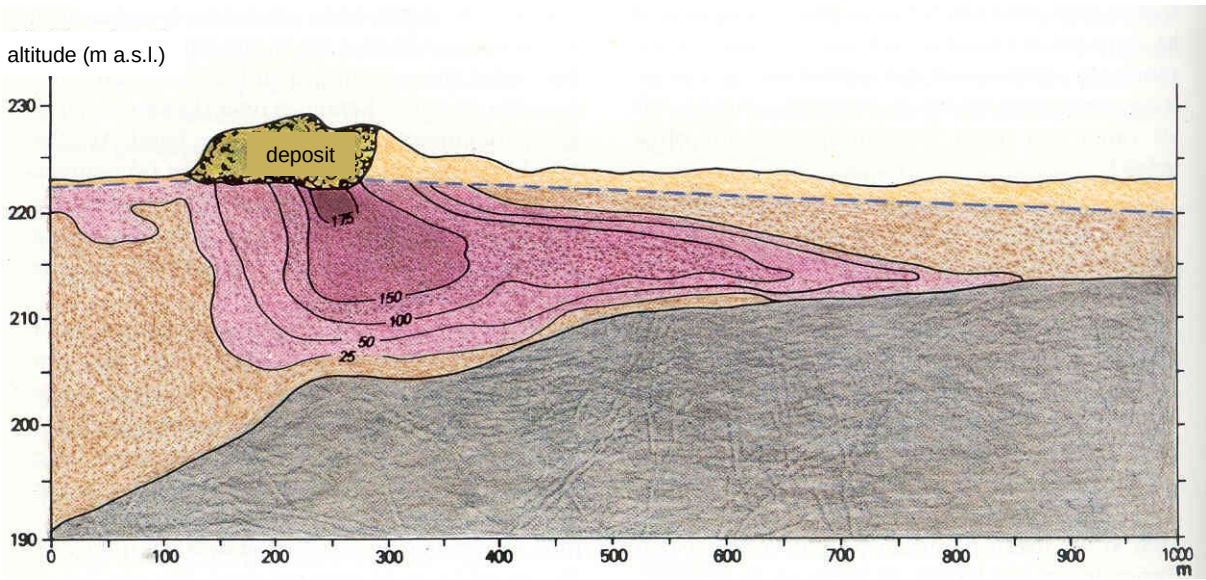


Fig. 59 A waste deposit was placed high up in a recharge area close to Toronto, Canada. By measuring the electric conductivity of the groundwater (given here in mS/m) it was possible to show how the salt rich leached water from the deposit was spread in the groundwater.

One way to reduce the flow of leached water through a deposit is to cover it with two soil layers: First a coarse layer and above that a relatively fine textured layer. If the sides of the deposit are

sloping one gets, in the same way as shown in Fig. 16, a flow along the lower boundary of the fine textured layer. As long as the rain intensity is moderate no water will penetrate the covering material down to the deposit. At large rain intensities break through may take place as illustrated in Fig. 17. As a whole the flow of leached water will be reduced.

Figure 59 illustrates, by the electric conductivity of the groundwater, how contaminants have spread from a waste deposit in Toronto, Canada. The waste deposit is situated on a recharge area close to the groundwater divide. The soil texture in the area is sand on relatively watertight clay. The groundwater is therefore flowing from the deposit toward the right hand side of the figure and brings leached water from the deposit. The hydraulic conductivity of the sand is $7.5 \cdot 10^{-5}$ m/s and its porosity is 35%. By using Darcy's Law and the expression for the velocity of the water particles the particle velocity can be estimated to 20 m per year. The figure shows the situation about 40 years after the first waste was deposited. The estimated extent of the contaminant plume, 800 m, is close to what can be seen from the measurements.

The conclusion to be drawn from this example is that it is inappropriate to locate a waste deposit far up in a recharge area because all groundwater downstream from the deposit will be contaminated. We have also seen that the contamination can continue for a long time and reach far away from the source.

The best place to locate a waste deposit may be at the boundary between the recharge- and discharge area, since the water that infiltrates there will soon discharge again without contaminating deeper groundwater. It will therefore be relatively simple to collect the leached water and groundwater in drainage ditches for later purification.

The restrictions on waste deposits are of cause different for different wastes. Greater demands are put on hazardous waste than on substances that can be mineralised. If the waste contains sulphides like pyrites from mines it is important to locate them out of reach for oxygen, i.e., below the groundwater surface. In the air sulphuric acid is formed, which can be harmful if leached to water courses in larger quantities.

Soil microorganisms can purify household waste water

Domestic sewage water can be purified when it is allowed to pass a soil profile. By means of the activities of microorganisms, and by adsorption to soil particles during the percolation, water gets rid of its contaminations. The purification takes mainly place in the contact surfaces between soil particles and water. It is therefore most efficient in a fine textured soil, but for the soil to admit enough percolation the hydraulic conductivity should not be too small. Even if sludge and larger particles have been separated from the sewage water before infiltration, the pores will get partly clogged. The soil therefore has to be rather coarse textured, like fine sand, gravely till or coarser.

The sewage water is spread over an area at a depth of about one metre by means of a layer of gravel put in the soil profile. If the soil is coarse enough and the distance to the water table is large enough, the water is allowed to percolate down to groundwater. In a fine textured soil, and when the groundwater level is shallow, the sewage water is infiltrated through an artificial sand bed. After infiltration the water is collected and diverted to a suitable watercourse. This construction is called an infiltration bed, to be distinguished from the former, which is called infiltration plant. The infiltration plant, with the infiltration taking place in the natural soil profile, normally gives a better purification of the water than the infiltration bed. The reason for this is

that the water in the infiltration plant percolates through a thicker soil layer and that the natural soil often consists of more fine particles than the sand in the infiltration bed.

An aerobic environment is sustained in the profile when the water is added at a suitable rate. This is a prerequisite for the activity of many microorganisms. Among the organisms that can live by purifying sewage water we find groups that oxidize ammonium- and nitrite-nitrogen to nitrate (in aerobic environment), groups that reduce nitrate to nitrogen- and nitrous oxide gas and sulphate to sulphides (in anaerobic environment). In experiments with infiltration beds almost complete oxidation of the organic material in the sewage water has been achieved, about half of the phosphorous content has been immobilized, while the decrease in total nitrogen has been 10 to 50%. In infiltration beds the number of coli bacteria is reduced, but has not been totally eliminated. In well-designed infiltration plants, the groundwater beneath the plant has been practically free from coli bacteria.

The private well

The groundwater issue most likely to confront a person is the amount and quality of a water supplied to their house, farm or summer cottage by a private well. Where should the well be dug or drilled, how should it be formed and how should it be protected to give clean, tasty and abundant water? The geology on the site is of course important, but let us give some general comments based on the discussion in earlier chapters.

The amount of water that can be pumped from a well depends on the groundwater recharge in the recharge area of the well and on the hydraulic conductivity of the soil or bedrock.

Swedish households use about 200 litres of water per person and day. As stated on page 61 the groundwater recharge is of about the same size as the specific runoff, i.e., the discharge in the stream per unit area of its catchment. In Sweden the specific discharge is lowest in the landscapes close to the Baltic Sea, with a long-term mean of about 180 mm per year. With this groundwater recharge each person needs a recharge area of $200 \cdot 365 / 180 = 400$ m^2. For a single household an area of 2000 m^2 may be needed, which is equivalent to a circle having a radius of 25 m. In rural areas the local groundwater recharge satisfies the household needs. But the problem is to extract the groundwater and to ensure that it is of good quality. Let us first look at the availability of the groundwater.

Groundwater exists everywhere

On most places in Sweden groundwater exists on reasonable depth below the ground surface. Groundwater is found wherever you dig, push down a tube, or drill. To be able to use the groundwater, the transmissivity of the soil or bedrock must be large enough to allow a sufficient amount of the groundwater that is recharged in the surroundings to flow to the well.

In sorted soils, with texture from fine sand and coarser, as well as in coarse till soils, the hydraulic conductivity does not limit the water supply of the household well. If the geological formation is reasonably large, and the groundwater zone reasonably thick, enough water can be transmitted to the well.

In fine textured till soils (sand-, silt- or clay till) it is necessary to find places with sandy or gravely lenses of sufficient horizontal extent below the groundwater surface. When groundwater is pumped from the well, such lenses, where the conductivity is relatively large, will collect water

from the layers above and transmit it to the well. The vertical flow from the less permeable layers is certainly small, but as it takes place over a large area the total recharge can be large.

Valleys having till soil- or sand layers below a clay layer are often suitable as water supply. In these confined aquifers (cf. page 59) the groundwater recharge in the elevated areas outside the area covered with clay is used.

When wells are drilled in solid rock one tries to find fissures that can transmit enough water to the well. Water bearing fissures in rock are usually, as in the case with the coarser lenses in till soils, in contact with groundwater that is formed over a large area. The groundwater recharge is therefore normally sufficiently large and the capacity of the well is determined by the ability of the fissures around the well to conduct water. Wells drilled in solid rock often give enough water for single households. Their capacity is moderate, and to satisfy larger needs wells in coarse soils are needed.

The water use in a household varies during the day, but in a soil with low conductivity the largest possible groundwater flow may only satisfy the mean consumption. To make temporary high water extractions possible, a storage in the well is needed. Such a storage decreases, i.e., the water level falls, when water is taken from the well. During the rest of the time the storage is filled again by the groundwater flow. Such well storages are usually needed in till soil wells. In a favourable coarse textured soil the groundwater flow to the screen of an extraction pipe may be enough even for large extractions.

The recharge area increases when groundwater is extracted

When water is extracted from a well, the water level in the well sinks below the original groundwater level. The groundwater surface surrounding the well will then slope towards the well, and a flow is induced towards the well. When there is equilibrium between water extraction and drawdown the groundwater flow through every imaginable cylindrical surface around the well is equal to the extraction rate. The cylindrical cross sectional surfaces for the groundwater flow gets smaller closer to the well, because their radius as well as their height decreases. The groundwater flow in the soil per cross sectional area (Q/A) is therefore largest closest to the well. Since the flow is proportional to the slope of the groundwater surface, the slope must increase closer to the well. The groundwater surface forms a so-called cone of depression, with the well as its tip. The shape of the cone depends on the extraction rate and the conductivity of the soil. In a fine textured soil, a large slope of the groundwater surface is needed to maintain a given flow. The cone of depression is therefore deep and steep. In a coarse textured soil the same extraction rate gives a shallow and flat cone. At the same drawdown in both wells, which implies a larger extraction rate from the coarse aquifer, the cone of depression covers a larger area in the coarse than in the fine textured aquifer. The advantage with high conductivity is that a high extraction rate is possible for a short time and that it is possible on a long term basis to extract groundwater from a large area around the well.

The storage coefficient is often used to characterize an aquifer. It is (as stated on page 63) the ratio between the groundwater storage change and the change in water level. The smaller the storage coefficient, the more rapid is the decline of the groundwater surface when an extraction starts. But it does not influence on the extraction rate at equilibrium, which is determined by the conductivity (see above).

Some wells occasionally get dry after a long dry spell with little groundwater recharge, for example during late winter or during summer. The reason is a combination of a natural recession

of the groundwater level (cf. Fig. 32) and drawdown from the extraction. In till soils it may happen that the groundwater level falls below a coarse lens that collects water to the well. Additional extraction will then lead to a rapid decline of the water level in the well. In a coarse textured soil, the groundwater level is often rather deep, but the natural variation of the groundwater level is small during the year and, further, it is not necessary that the well reaches much below the normal groundwater level to give a certain capacity. In a fine textured soil, the groundwater level is, on the other hand often shallow, but the yearly variations are large and the well has to reach far below the natural groundwater level to give a certain capacity. To secure the water supply during dry spells a deep well is therefore also needed in that case.

The recharge area of a well from which no water is extracted is often small. When the groundwater surface is a sloping plane it only consists a narrow strip upstream from the well. When water is extracted from the well the groundwater flow lines are diverted towards the well by the cone of depression that is formed in the natural groundwater surface, and the recharge area increases drastically. In similar geology the best capacity should be found in a well that have a large natural recharge area.

Some natural quality problems

With respect to the chemical changes of water along its path through a catchment it is not surprising that the chemical quality of water may vary substantially between wells. We have pointed out that the salt concentration increases and the hydrogen ion concentration decreases (pH increases) along the groundwater flow paths. Even if the water quality to large extent is determined by the local geology, some general comments can be made that are connected to the flow of water.

Water in many wells has too high iron content. It may lead to rust coloured bathtubs and sinks. Very high iron concentration gives the water smell and taste. Since iron is frequent in many minerals (cf. table 4 on page 105) it is abundant in soils and bedrock. This is clearly shown by the illuvial horizon of the podzol profile. But if dissolved iron is transported with the water to a well or not is largely dependent on the groundwater's oxygen conditions and contact with organic materials along the flow paths.

Compounds with reduced iron (Fe^{2+}) are easily soluble, and it is mainly as reduced iron ions that iron is transported with water in the soil. When oxygen is dissolved in the water, the reduced iron is oxidized to Fe^{3+} and forms sparingly soluble hydroxides (rust, $FeO(OH)$) that are deposited and not transported further. If organic acids are present they may form complexes with Fe^{3+} (cf. page 106) and it may remain even in aerobic water. That is the case in the eluvial horizon. In the illuvial horizon the Fe^{3+} is deposited when the organic acids decompose or are deposited. For Fe^{3+} to be reduced, anaerobic conditions are needed as well as organic material or some other substance that can be oxidized and that can use the iron as electron acceptor. Iron that is deposited in the illuvial horizon when the groundwater level is low may later be reduced when the groundwater level is high enough to reach the root zone. The oxygen gas that is consumed by the root respiration will then not be replaced by new oxygen gas, oxygen free conditions are formed and iron can be reduced. Wells, with recharge areas that totally or partly consist of such areas, therefore have often water rich in iron. If the contact between the water in the well and the free atmosphere is good and the water is oxygenated some iron will be oxidised and deposited as rust flocculates in the well.

The most common form of manganese in nature is manganese dioxide (MnO_2). The manganese here has the oxidation state +4. Mn (III) forms hydroxide ($MnO(OH)$) in water and is precipitated

as a black precipitate. In oxygen free environment both Mn (III) and Mn (IV) are reduced to Mn (II). It is, like Fe (II) water soluble. Manganese reacts in the same way as iron and high iron concentrations in a well are often accompanied by high manganese concentrations.

Organic substances in oxygen free groundwater can also reduce sulphate and nitrate ions. When sulphate is reduced, bad smelling hydrogen sulphide (H_2S) may form. Nitrate ions can be reduced to nitrous oxide or nitrogen gas.

There are, as we can see, many reasons to aerate an oxygen free groundwater. This is often done in municipal water works, for example by blowing in air at the bottom below the raw water. Reduced forms of iron and manganese are oxidised and precipitated, and dissolved gases as hydrogen sulphide are expelled from the water.

The concentration of calcium and magnesium in water, i.e., its hardness determines other important qualities of well water. Soft water dissolves concrete and increases the risk of corrosion in water pipes. Therefore calcium carbonate is often added to soft water.

Hard water requires more washing detergent and forms compounds with calcium, which have low solubility. When hard water is boiled, calcium and magnesium are deposited in the cooking-vessel. The water hardness can be decreased by means of a softening filter. Such filters act is ion exchangers, and calcium and magnesium are exchanged for sodium that is added as sodium chloride.

The reason for hard groundwater is the weathering of calcium- and magnesium minerals. The hardness of groundwater is therefore increased with its age, and the water is especially hard in areas with lime bedrock and lime rich soils.

Sometimes the water in a well may be coloured by humic substances. This often happens during spring and autumn, when the groundwater level is close to the soil surface and in contact with organic materials. Shallow groundwater can in such situations be rapidly transported to the well and enter through leaking walls. Such water can have high concentration of dissolved organic carbon and still contain dissolved oxygen. It may contain iron, since it can be transported as complex with organic acids.

A good well may easily be non-usable

The most common damage to a well is that the water quality declines, but also its capacity may decrease due to human impact.

The groundwater level may decrease and the capacity of a well may become insufficient because of digging or large water extractions in the vicinities of the well. The risk of influence is greatest in a coarse textured soil, where the groundwater level, as pointed out above, may be affected far away from the interference. Digging on the upstream side with respect to the groundwater flow causes greatest risk, since the natural groundwater flow could be diverted and the groundwater level around the well be lowered (cf. ditching of forest soils, page 122). Wells that get their water from a coarse layer in a fine textured soil can be very sensitive to interference. Ditching close to the well may, for instance, cut off the layer conducting water.

The variation of the groundwater level during a year and between years makes it difficult to judge if deteriorated water availability in a well is due to interference or natural variation in groundwater recharge (cf. Fig. 6). In many cases the well recovers when the precipitation again increases after a

period with precipitation deficit. The groundwater network at Swedish Geological Survey (SGU) monitors the groundwater level in a number of unaffected areas in different parts of Sweden. These observations can give some help for judging a possible human influence on the water availability in a well.

Earlier good wells in coastal areas may suddenly give salt contaminated water when the groundwater level has decreased because of increased tapping from the well or other wells in the surroundings. Along a coast there is always a surface where salt and fresh water meet. This surface is sloping from the shore in under land because the fresh groundwater, which has lower density, is floating on top of the salt water. The depth to the separating surface at a site depends on the groundwater level and on the densities of the salt and fresh water. On the Swedish west coast, the depth to the separating surface is about 40 times the groundwater level above the sea level. On the east coast the corresponding depth is about 250 times the groundwater level. The difference is due to differences in the salt content of the seas and thus in their density. When the groundwater level drops one metre the salt limit increases 40 respectively 250 m. If the intake to the well is below the sea level a decrease in groundwater level can therefore lead to salt water intrusion into the well, even if the groundwater level in the well is above the sea level. Groundwater extractions in coastal areas should therefore be planned in a coordinated way in order to keep the groundwater level in the area at sufficient level. (Also deep wells far away from the coast can give salt water. In our country it is then very old water from the time when the area was situated below the sea level.)

Since wells are located close to buildings and human activities the water has great possibilities to get contaminated on its way to a well. The water dissolves and transports many of the contaminations that are added to the soil in the recharge areas of the well. To protect a well from contaminations it is first necessary to find out from where the water in the well emanates, i.e., the recharge area of the well. For a well in soil this can be done rather safely by levelling the groundwater surface in a number of points around the well (cf. Fig. 18). If this is not possible, the natural flow directions can be inferred from the topography of the soil surface and the general geological situation. The water levels in ditches and in other free water surfaces are usually parts of the groundwater surface that are easy to determine. The appearance of the cone of depression is however difficult to estimate without measurements. The recharge area to a well drilled into the bedrock is always difficult to determine, since it depends on the extent and orientation of the fissures. Deep fissures can get their water from hills far away from the local recharge area, but they may also be recharged close to the well.

Some contaminants disappear after some time in the soil, as for example a large part of the domestic sewage (cf. page 130). The residence time for groundwater between the point of contamination and the well therefore plays an important role. After 60 days, groundwater is considered safe regarding spreading of diseases. The residence time for water in the unsaturated zone may be an even more important factor, since the decomposition is most efficient there. A thick unsaturated zone is a good protection against many contaminants. But other contaminants, as for example nitrate, are transported by water without large changes. For such contaminants, the whole recharge area is a risk area. By molecular diffusion, and by the relative difference in the water particle velocities, a contaminant that is added in a point will spread into a growing "cloud" that is transported with the groundwater flow in the same way as the smoke from a chimney is spread in a plume in the air. A long flow pathway results in a large dispersion and thus dilution of the contaminant as the cloud grows, and it takes a long time for it to reach the well. But the effect of a short contaminant input may last for a very long time after the contaminant has appeared in the well.

In chapter 3 and 6 we have discussed the velocities of water particles under natural conditions. The vertical velocity in the unsaturated zone is often low with mean velocities of about one or a few metres per year. These velocities are valid for natural water input to the soil surface, i.e., a water recharge that is relatively slow. If, on the contrary, a barrel of contaminated water is poured out on the ground, near water saturated conditions will be reached in the soil and the added water particles can soon reach many metres down into the unsaturated zone through coarser pore systems (cf. Fig. 20). The groundwater can in this way be contaminated within a short time.

Under natural groundwater flow conditions in soil the velocity of the water particles maybe about 0.1 – 1 m per day in soils that are used as aquifers. A coarse textured soil does not result in higher particle flow velocities than a fine textured soil under natural conditions. Since the porosity is about the same in different soils, perhaps 30%, a given flow rate gives the same particle velocity (cf. equation on page 38). At equal groundwater recharge, the groundwater flow, i.e., transported volume per time unit, is the same at a given distance from the water divide and independent of soil type. The flow, and thus the particle velocity is then determined by the cross sectional area, i.e., the thickness of the groundwater zone. If this is equal the particle velocity is the same in a coarse textured and a fine textured soil (but the slope of the groundwater surface is largest in the fine textured soil). If on the other hand the slope is the same the groundwater zone is thinnest and the particle velocity largest in the coarse textured soil. This is the case in a fine textured soil with a coarse textured layer. Here the total potential gradient is nearly the same in both soils and equal to the slope of the groundwater surface. This will give the largest flow and particle velocity in the coarse layer. It can also be the case in soils on impermeable sloping ground (cf. Fig. 49), where the thickness of the groundwater zone is determined by the hydraulic conductivity of the soil. In many cases the groundwater level in a coarse textured soil is determined by a threshold level (cf. page 74), which can give a thick groundwater zone but small flow and low particle velocity (and small slope of the groundwater surface) in such soils.

In solid rock the natural particle velocity of water can be relatively high. Because the porosity is very small compared with a soil the particle velocity for a given flow is many times larger than that in the soil. In a preliminary tracer experiment in the Kloten area a natural maximum particle velocity of about 10 m per day was measured in the upper part of the bedrock groundwater.

The particle velocity around a well that is in use is different from the natural velocity, both concerning direction and magnitude. Normal extraction in a household well in a soil aquifer seldom results in substantial increases in velocities, apart from within the nearest metres from the well. If the thickness of the groundwater zone at an extraction rate of 2000 litres per day is 1 m at a distance of 2 m from the well the particle velocity there will be 0.5 m per day. At 10 m distance the same extraction and thickness give a velocity of 0.1 m per day. The figures are mean velocities calculated from the equation on page 38, cylindrical cross sectional surfaces and 30% porosity. If flow occurs in coarser lenses or macropores the particle velocities in them will be much larger. Normally it is this maximum velocity that is of interest. In a well drilled into solid rock even relatively small extractions can result in large particle velocities. Pumping in wells in solid rock has given velocities of about 30 m per day when the distance between the point of tracer injection and the well was 30 m. (The figure is the mean velocity over the total length, but close to the well the velocity may be even higher.) Drilled wells in solid rock in areas with outcrops or shallow soil cover can be sensitive to contaminations hundreds of metres from the well.

When the extraction from a well increases, the recharge area increases because the cone of depression gets wider. Locations that earlier were situated "downstream" may now be situated "upstream", and earlier harmless contaminations may start to contaminate the water in the well. As long as the water was carried from a hand driven pump the water was clean, but when an

electrical pump was installed and the water use increased drastically, the water in the well may have become useless.

A common problem is that there may be some form of recirculation of domestic water. The well is contaminated by an infiltration plant or a leaking sewage pipe. An infiltration plant and a well should be placed in such a way that the groundwater level at the infiltration plant is always lower than the lowest water level in the well during extraction. If this condition is fulfilled, a minimum distance of 50 m is recommended in Sweden.

In many wells the water quality declines during periods with abundant water supply, i.e., during snowmelt and sometimes during the autumn. The reason may be shallow groundwater table and changed flow paths in the soil. The water that reaches the well during periods with shallow groundwater table has in the average had a shorter residence time in the soil than the water that enters the well during other times. The unsaturated zone is thinner and contaminants that were added to the soil surface have not been given time to decompose before they reach the groundwater. When the groundwater level rises and saturates a part of the unsaturated, zone some contaminants that earlier were relatively immobile may soon be mobilised and start to move with lateral groundwater flow. A leaking sewage pipe, that earlier was in the unsaturated zone, may for some time be included in the groundwater zone and contaminants may rapidly reach the well. In addition, the "natural" groundwater flow and the particle velocities increase and contaminants from more distant areas can reach the well area before they have been decomposed. This effect should be more pronounced when the high groundwater level persists for a longer time.

Hortonian overland flow from the near vicinity of the well, as well as melt and rainwater directly on the well cover can also transport contaminants to the well. The domestic well is often located where the soil is tramped and compacted and the infiltration capacity decreased.

In a Danish investigation on dug wells it was found that the bacterial content in the water could be predicted from the construction and location of the wells. In reasonably well located watertight wells with intake at the bottom only, no bacteria were found. The amount of bacteria increased with leaking lids and walls and decreasing distance to manure heaps and sewage wells.

Well in recharge- or discharge areas?

There are good wells in recharge areas as well as in discharge areas. Since the lowering of the water table in a well creates a recharge area, even a well located in relatively flat terrain on the former groundwater divide may be useful. A well that is in use will act as a small discharge area, a discharge point. Many older wells make use of springs, i.e., natural discharge points that have been formed by some geological irregularity, for example where a tight layer approaches the soil surface or at a hole in a tight clay layer on a confined aquifer.

A good location of a well is determined from the geology rather than from the topography. In some cases, for example in fissure valleys, favourable geology coincides with favourable topography. There the soil is thick and the soil layers are often good for groundwater extraction, the bedrock is often fissured and the recharge area large. At the bottom of hills and especially at concave terrain forms the groundwater level is also shallow and the natural groundwater level fluctuations small.

From the view of contamination the groundwater should not be too shallow around the well. Saturated discharge areas are therefore not suitable if the discharge is not very large, as in a richly

flowing spring. If the groundwater level can constantly be kept low, by extraction from the well or by drainage of the shallow groundwater, the water can be of good quality even in a saturated discharge area. In unsaturated discharge areas, with the groundwater surface some metre below the soil surface, the water quality should be good. There is especially low risk for contamination in wells in confined aquifers, for example in coarse textured soil below clay soil. In these discharge areas there are normally a flow from the coarse layer up through the clay layer. Contaminations in the groundwater in the clay layer, such as nitrate from agricultural activity, can in such case not affect the quality of the water in the well. At large extraction rates the flow direction in the clay may be reversed close to the well. But since the conductivity of the clay is very small and the clay layer often is thick, the flow will be small and contaminations will be diluted by flow of other, unpolluted groundwater to the well.

Some basic chemical concepts

Element	Matter that consists only of atoms having the same nucleus charge.
Atom	The smallest part of an element that still keeps the properties of the matter. The atom consists of a dense, central nucleus surrounded by a cloud of negatively charged electrons. The nucleus in turn consists of positively charged protons and non-charged neutrons.
Ion	An atom having an unbalance between positive and negative charges.
Cation	An atom that has lost one or more electrons and thus is positively charged (e.g. a sodium ion, Na^+).
Anion	An atom that has gained one or more electrons and thus is negatively charged (e.g. chloride ion, Cl^-).
Acid	A substance that can deliver protons, i.e., hydrogen ions (H^+).
Base	A substance that can take up protons. (There are more general definitions of acid/base.)
Basic cation	Potassium (K^+), Calcium (Ca^{2+}), Magnesium (Mg^{2+}), Sodium (Na^+).
Acid cation	Proton (H^+), Aluminium (Al^{3+}), Iron (Fe^{3+}).
Oxidation	Loss of electrons, e.g. $Fe^{2+}(aq) \rightarrow Fe^{3+} + e^-$.
Reduction	Uptake of electrons, e.g. $O_2\ (aq) + 2H_2O + 4e^- \rightarrow 4OH^-$.
Redox	An oxidation of an element always result in a reduction of another element and the above can be combined to $4Fe^{2+}\ (aq) + O_2\ (aq) + 2H_2O \rightarrow 4Fe^{3+}\ (aq) + 4OH^-$.
Ion exchange	A substance that has an ion loosely bound may get this ion exchanged by another ion of the same charge in an environment rich in the new ion. If the substance has positive ions loosely bound it is called a cation exchanger, while it is called an anion exchanger if it has negative ions loosely bound. Cation exchangers almost completely dominate in soils.
Chemical equilibrium	Many chemical reactions in the environment may go in both directions. Chemical equilibrium is at hand when the rate of reaction going from left to right in a chemical formula is of equal size as that going from right to left. The direction of the reaction is determined, apart from by temperature and pressure, by the concentration of the substances that take part in the reaction.
Base saturation	The relative amount of cation exchange sites occupied by base cations (the remaining sites are occupied by acid cations, especially by protons and aluminium ions).
Buffering	Weak acids or bases counteract pH-changes in a solution by a shift in balance if strong acids or bases are added. For example the bicarbonate ion dissociates into a carbonate ion and a hydrogen ion (proton); $HCO_3^- \leftrightarrow H^+ + CO_3^{2-}$. If protons are added (strong acid) the balance in the equation above is shifted towards the left and the increase in number of protons in the solution will be less than those added.
Equivalent weight	The amount in gram of an ion that its molecular weight states (i.e. the mol weight), divided by the charge of the ion.
pH	The logarithm of the hydrogen ion concentration in moles per litre with reversed sign, $-\ ^{10}\log [H^+]$.
Precipitate	Solid substance that separates from a solution. It is also the operation in which the substance is separated. A precipitate is usually the result of a chemical reaction, e.g., when ions in a solution join to form a new

	substance of low solubility. The texture of a precipitate may be crystalline, flocculent, fine grained, etc.
Solution	Homogeneous mixture of two or more solid, fluent or gas formed substances. The component that exists in greatest amount is called the solvent and the other dissolved substances.

Chemical terms used in the text

Al	Aluminium	Mg	Magnesium
AlOOH	Diaspor	Mn	Manganese
C	Carbon	MnO_2	Pyrolusite
Ca	Calcium	MnO(OH)	Manganese hydroxide
CO_2	Carbon dioxide	N	Nitrogen
CO_3^{2+}	Carbonate ion	Na	Sodium
Fe	Iron	NH_4^+	Ammonium ion
FeO(OH)	Iron hydroxide (rust)	NO	Nitric oxide
FeS_2	Pyrite	NO_2	Nitrogen dioxide
H	Hydrogen	NO_3^-	Nitrate ion
H^+	Hydrogen ion (proton)	O	Oxygen
HCO_3^-	Bicarbonate ion	OH^-	Hydroxyl ion
H_2CO_3	Carbonic acid	P	Phosphorus
H_2O	Water	S	Sulfur
H_2S	Sulfide	SO_2	Sulfur dioxide
H_2SO_4	Sulfuric acid	SO_4^{2-}	Sulfate ion
H_4SiO_4	Silicic acid	Si	Silicon
K	Potassium	Ti	Titanium

Mathematical symbols

A	Cross sectional area, catchment area
A_e	Effective cross sectional area
A_u	Size of saturated discharge area
b	Width
c_s	Tracer concentration in streamwater
c_g	Tracer concentration in groundwater
c_p	Tracer concentration in rain and melt-water
E	Evapotranspiration
g	Acceleration of gravity
h	Height, groundwater level
I	Electrical current
K	Hydraulic conductivity
M	Specific storage
m	Thickness
P	Precipitation
p	Porosity
PE	Potential evaporation
Q	Discharge
R	Runoff; Groundwater recharge; resistance
r	Radius
r_a	Aerodynamic resistance
r_s	Surface resistance
S	Storage content
s	Dry bulk density
T	Transmissivity, Temperature
U	Voltage (potential difference)
V_g	Discharged volume groundwater
V_p	Discharged volume fresh rain or meltwater
V_t	Total discharged volume
v	Velocity
v_d	Darcy velocity, flow
v_p	Velocity of water particles (mean velocity in macro scale)
w	Water content as weight ratio
X	Fraction of groundwater in streamwater
x	Distance, coordinate direction
z	Height, coordinate direction
α	Contact angle between meniscus (the curved surface of a liquid in a thin tube) and wall
γ	Surface tension. Electric conductivity
Δ	Change, difference
$\delta^{18}O$	Relative deviation in $^{18}O/^{16}O$ isotope ratio of a water sample from that of the international standard water SMOW (Standard mean ocean water)
θ	Water content as volume ratio
ϱ	Density of water
ϕ	Total potential of water
ψ	Water pressure potential;Tension

Literature

Chapter 1

Page 2. Kircher, A. 1664. *Mundus subterraneus.* Apud J. Janssonium & Weuerstraten, Amsterdam. 866 pp.

3. Perrault, P. 1678. *De l'origine des fontaines.* Paris. Published by A. LaRocque: On the origin of springs. Hafner, New York 1967. 209 pp.

4. Horton, R. E. 1933. The role of infiltration in the hydrologic cycle. *Trans. Am. Geophys. Union* 14:446:460.

5. Betson, R. P. 1964. What is watershed runoff? *J. Geophys. Res.* 69:1541-1552.

5. Dunne, T. & Black, R. D. 1970. Partial area contributions to storm runoff in a small New England watershed. *Water Resour. Res.* 6:1296-1311.

5. Dunne, T. 1983. Relation of field studies and modelling in the prediction of storm runoff. *J. Hydrol.* 65:25-48.

5. Sklash, M. G. & Farvolden, R. N. 1979. The role of groundwater in storm runoff. *J. Hydrol.* 43:45-65.

5. Rodhe, A. 1981. Spring flood – meltwater or groundwater? *Nordic Hydrology* 12:21-30.

5. Hewlett, J. D. & Hibbert, A. R. 1967. Factors affecting the response of small watersheds to precipitation in humid areas. I: W. E. Sopper & H. W. Lull (Eds.), *Int. Symp. on Forest Hydrology (1965)*, Penn. state univ. Pergamon press, New York. p 275-290.

5. Kirkby, M. J. (Ed.) 1978. *Hillslope hydrology.* John Wiley & Sons, Chichester. 389 pp.

5. Ward, R. C. 1984. On the response to precipitation of headwater streams in humid areas. *J. Hydrol.* 74:171-189.

5. Gustafsson, Y. 1970. Topografins inverkan på grundvattenbildningen. In: E. Eriksson, Y. Gustafsson & K. Nilsson (Eds.), *Grundvatten.* PA Norstedts & Söners förlag, Stockholm. p 15-33. (Earlier in E. Eriksson, Y. Gustafsson & K. Nilsson (Eds.), *Ground water problems.* Pergamon press, Oxford, 1968. p 3-19.)

Chapter 2

Eriksson, B. 1980. *Sveriges vattenbalans. Årsmedelvärden (1931-60) av nederbörd, avdunstning och avrinning.* SMHI Rapport RMK 18. 34 pp.

Fig 1 After *Skogliga fälstationen i Vindeln* 1983. Svartbergets och Kulbäckslidens försöksparker. Stencil nr 3. Inst. för Skoglig ståndortslära, SLU, Umeå. 33 pp.

Fig 3 After Eriksson, E. 1974. Vattnet – kemikaliebäraren. *Forskning och framsteg* 5:41-45.

9 Bolin, B. & Rodhe, H. 1973. A note on the concepts of age distribution and transit time in natural reservoirs. *Tellus* 25(1):58-62.

Fig 5 Data from SMHI.

12 Eriksson, B. 1983. *Data rörande Sveriges nederbördsklimat. Normalvärden för perioden 1951-80.* SMHI Rapport 1983-28. 92 pp.

12 *Vattenföring i Sverige.* SMHI, LF/ALLF, Stockholm, 1979. 403 pp.

12 Gottchalk, L. 1975. *Stochastic modelling of monthly river runoff.* Bull. Ser. A, No. 45, Inst. för Teknisk vattenresurslära, Lund. 203 pp.

12 Melin, R. 1970. *Hydrologi i Norden*. Utbildningsförlaget, Stockholm. 205 pp.

Fig 6 Data from SMHI.

Fig 7, upper panel After Persson, M. 1978. *Slutrapport över hydrologiska undersökningar i Lappträskets representativa område*. SMHI Rapport HB 33. 21 pp.

Fig 7, lower panel After Waldenström, A. 1977. *Slutrapport över hydrologiska undersökningar i Kassjöåns representativa område*. SMHI Rapport HB 29. 59 pp.

15 Eriksson, B. 1980. *Statistisk analys av nederbördsdata. Del II Frekvensanalys av månadsnederbörd*. SMHI Rapport RMK 17. 150 pp.

15 Eriksson, B. 1983. See ref. p. 12.

Chapter 3

Hillel D. 1982. *Introduction to soil physics*. Academic press, New York. 364 pp.

Koorevar, P., Menelik, G. & Dirksen, C. 1983. *Elements of soil physics*. Elsevier, Amsterdam. 228 pp.

Fig 11 After Andersson, S. & Wiklert, P. 1972. Om de vattenhållande egenskaperna hos svenska jordarter. *Grundförbättring* 25(2-3):53-143, and after Lundin, L. 1982. *Mark- och grundvatten i moränmark och marktypens betydelse för avrinningen*. UNGI Rapport nr 56. Naturgeografiska inst., Uppsala univ. 216 pp.

Fig 13 After Danfors, E. 1963. Marken ur fysikalisk synvinkel. *Jord, gröda, djur*. LT Förlag, Stockholm. p 98-120.

Table 1 Fagerström, H., Wiesel, C.-E. & SGFs laboratoriekommitté, 1972. *Permeabilitet och kapillaritet*. BFRs informationsblad B7:1972. 44 pp.

Tabele 2 Jansson, P.-E. 1980. *SOIL water and heat model. II. Field studies and applications*. Acta Univ. Upsaliensis 568. 26 pp.

35 Bear, J. 1979. *Hydraulics of groundwater*. McGraw-Hill, New York. 567 pp.

Fig 18 Uppsala univ., Avd. för Hydrologi, B-kursen 1983 (unpubl.).

41 Beskow, G. 1932. *Tjälbildningen och tjällyftningen med särskild hänsyn till vägar och järnvägar*. SGU Ser. C, No. 375, Årsbok 26(3). 242 pp.

Fig 21 After skogliga fältstationen i Vindeln 1982. *Referensmätningen 1980/81*. Red. C. Degermark. Svartbergets och Kulbäckslidens försöksparker. Stencil nr 2. Inst. för Skoglig ståndortslära, SLU, Umeå. 32 pp.

Fig 23 After Calder, I. R. 1979. Do trees use more water than grass? *Water Services* 83(995):11-14.

43 Monteith, J. L. (Ed.) 1975. *Vegetation and the atmosphere, Vol. 1 Principles*. Academic press, London. 278 pp.

Fig 24 After Linacre, E. T. 1973. A simple empirical expression for actual evapotranspiration rates – a discussion. *Agric. Meteorol.* 11:451-452.

Chapter 4

Fig 25 After Arnell, V. 1974. *Intensitets-varaktighetskurvor för häftiga regn i Göteborg under 45-årsperioden 1926-1971*. Geohydrologiska forskningsgruppen, CTH, Medd. nr. 5. 68 pp.

Fig 26 Data from SMHI.

50 Bengtsson, L. 1984. Snowmelt induced urban runoff in Nothern Sweden. *Proc. Stormwater and water quality model users group meating, Gainesville, Florida, January 27-28, 1983*.p 215-236. (Also Uppsala univ., Avd. för Hydrologi, Rep. Ser. A 1984:8. 22 pp).

50 Lundberg, A. 1974. *Infiltration och perkolation i det översta marklagret. En sammanställning av mätningar utförda i Norden*. IHD Report 35. Stockholm. 58 pp.

Fig 27 After Lindblad, A. 1981. *Infiltrationsmätningar utförda vid geologiska institutionen CTH/GU 1972-1980. Sammanställning och statistisk bearbetning*. Geohydrologiska forskningsgruppen, CTH, Medd. nr. 60. 80 pp.

53 Rodhe, A. Overland flow experiments (unpubl.).

53 Burt, T. P. & Williams, P. J. 1976. Hydrualic conductivity in frozen soils. *Earth Surface Processes* 1:349-360.

53 Kane, D. L. & Stein, J. 1983. Field evidence of groundwater recharge in interior Alaska. *Proc. 4th int. conf. on permafrost*. National academic press, Washington D.C. p 572-577.

Fig 28 Grip, H. Infiltration experimnets (unpubl.).

56 Andersen, L.-J. & Sevel, T. 1974. Six years' environmental tritium profiles in the unsaturated and saturated zones, Grönhöj, Denmark. *Isotope techniques in groundwater hydrology*. IAEA, Wien. p 3-20.

56 Dressie, Z. 1984. *Groundwater recharge to eskers*. Rep. Ser. A, No. 2, Avd. för Hydrologi, Uppsala univ. 141 pp.

57 IAEA 1981. *Stable isotope hydrology. Deuterium and oxygen-18 in the water cycle*. International Atomic Energy Agency (IAEA), Wien. 399 pp.

59 Tamm, C. O. & Troedsson, T. 1957. A new method for the study of water movement in soil. *Geologiska Föreningens i Stockholm Förhandlingar* 79:581-587.

Fig 29 After Sanner, H. 1984. *Markvatten- och grundvattenströmning i en moränsluttning*. Rep. Ser. B 1984:1. Avd. för Hydrologi, Uppsala univ. 43 pp.

Fig 30 After Saxena, R. K. 1984. Seasonal variations of oxygen-18 in soil moisture and estimation of recharge in esker and moraine formations. *Proc. Nordisk hydrologisk konferens, Nyborg Strand, Danmark, 6-8 Augusti 1984*, NHP-rapp. nr 5, Köpenhamn. Bind 2 p 39-49.

Fig 31 After Jansson, P.-E., Lindroth, A. & Perttu, K. 1984. Energy and water fluxes in Pine forest ecosystems. In: K. Perttu (Ed.) *Ecology and management of forest biomass production systems*. Rapport 15:63-75. Inst. för Ekologi och miljövård, SLU, Uppsala.

Fig 32 After Knutsson, G. & Fagerlind, T. 1977. *Grundvattentillgångar i Sverige*. SGU Rapporter och medd. nr 9, Stockholm. 140 pp.

Chapter 5

65 Tóth, J. 1963. A theoretical analysis of groundwater flow in small drainage basins. *J. Geophys. Res.* 68:4795-4812.

65 Freeze, R. A. & Witherspoon, P. A. 1967. Theoretical analysis of regional groundwater flow. 2. Effect of watertable configuration and subsurface permeability variation. *Water Resour. Res.* 3:623-634.

Fig 33 After Gustafsson, Y. 1970. See ref. p 8.

66 Carlsson, L., Winberg, A. & Grundfeldt, B. 1983. *Model calculation of the groundwater flow at Finnsjön, Fjällveden, Gideå and Kamlunge*. SKBF/KBS Teknisk rapport 83-45. Svensk kärnbränsleförsörjning AB, Stockholm. 208 pp.

67 Johansson, B. 1980. *Groundwater dating by means of isotopes*. KBS Teknisk rapport 80-08, Stockholm. 36 pp.

67 Saxena, R. K. 1984. Surface and groundwater mixing and identification of local recharge-discharge zones from seasonal fluctuations of oxygen-18 in groundwater in fissured rocks. I: *Hydrochemical balances of freshwater systems*. IAHS Publ. No. 150:419-428.

Fig 35 After Lundin, L. 1982. See ref. fig. 11.

67 Lundin, L., Jansson, P.-E. & Grip, H. 1981. *Mark och markförhållanden i Nickobacken*. Intern rapport 102. Barrskogslandets ekologi, SLU, Uppsala. 17 pp.

67 Carlsson, L. et al. 1983. See ref. p 66.

Fig 37 Calles, U. M. Groundwater studies (unpubl.).

Fig 38 After Calles, U. M. 1985. Deep groundwater contribution to a small stream. *Nordic Hydrology* 16(1):45-54.

Fig 39 Rodhe, A. Streamflow generation studies (unpubl.).

73 Rodhe, A. 1984. Groundwater contribution to steamflow in Swedish forested till soils. *Isotope hydrology 1983*. IAEA, Wien. p 55-66.

75 Eriksson, E. 1977. Ground-water recharge and surface run-off. *STRIAE* 4:11-12.

75 O'Loughlin, E. M. 1981. Saturation regions and their relations to soil and topographic properties. *J. Hydrol.* 55:229-246.

76 Andersson, M. G. & Burt, T. P. 1978. The role of topography in controlling throughflow generation. *Earth surface processes* 3:331-344.

77 Ivanov, K. E. 1981. *Water movement in mirelands*. Academic press, London. 276 pp.

77 Finnish peatland society 1982. *Peatlands and their utilization in Finland*. Soumen Graafinen Group Oy, Helsingfors. 139 pp.

79 Päivänen, J. 1973. *Hydrualic conductivity and water retention in peat soils*. Acta For. Fenn. 129. 70 pp.

Chapter 6

81 Chow, V. T. (Ed.) 1964. *Handbook of applied hydrology*. McGraw-Hill, New York. 1453 pp.

81 Pinder, G. F. & Jones, J. F. 1969. Determination of the groundwater component of peak discharge from the chemistry of total runoff. *Water Resour. Res.* 5:438-445.

81 Dinçer, T., Payne, B. R., Florkowski, T., Martinec, J. & Tongiorgi, E. 1970. Snowmelt runoff from measurement of tritium and oxygen-18. *Water Resour. Res.* 6:110-124.

Fig 46 After Rodhe, A. 1985. Groundwater contribution to streamflow in the lake Gårdsjön area. *Ecol. Bull.* (Stockholm) 37.

Fig 47 After Nilsson, T. 1981. *Sambandet grundvattenstånd – avrinning i Gårdsjöområdet, Bohuslän*. C-uppsats, Avd. för Hydrologi, Uppsala univ. 65 pp.

Fig 48 After Rodhe, A. 1984. See ref. p 73.

Table 3 Rodhe, A. 1984. See ref. p 73.

87 Herrmann, A. & Stichler, W. 1980. Groundwater-runoff relationships. *Catena* 7:251-263.

87 Sklash, M. G. & Farvolden, R. N. 1979. See ref. p 8.

Fig 49 After Johansson, B. 1985. A study of soil water and groundwater flow of hillslopes – using a mathematical model. *Nordic Hydrology* 16(2):67-80.

90 Ragan, R. M. 1968. An experimental investigation of partial area contributions. IAHS Publication No. 76:241-249.

91 Sklash, M. G. & Farvolden, R. N. 1979. See ref. p 8.

91 Freeze, R. A. 1972. Role of subsurface flow in generating surface runoff. 1. Base flow contributions to channel flow. *Water Resour. Res.* 8:609-623.

91 Freeze, R. A. 1972. Role of subsurface flow in generating runoff. 2. Upstream source areas. *Water Resour. Res.* 8:1272-1283.

92 Corbett, E. S., Sopper, W. E. & Lynch, J. A. 1975. Watershed response to partial area applications of stimulated rainfall. IAHS Publication No. 117: 63-73.

94 Lundin, L. 1982. See ref. fig. 11.

94 Knutsson, G. 1971. Studies of groundwater flows in till soils. *Geologiska Föreingens i Stockholm Förhandlingar* 93:1-22.

94 Beven, K. & Germann, P. 1982. Macropores and water flow in soils. *Water Resour. Res.* 18:1311-1325.

95 Mosley, M. P. 1979. Streamflow generation in a forested watershed, New Zealand. *Water Resour. Res.*15:795-806.

95 Grip, H. 1977. Recession – en tidsinvariant process? *Vannet i Norden* 10(2):50-53.

95 Kilpatric, P. A. 1980. Simulation of the effects of transpiration losses from the unsaturated zone on seepage from a soil slope. *J. Hydrol.* 46:30-309.

Fig 50 After Calles, U. M. 1982. Diurnal variations in electrical conductivity of water in a small stream. *Nordic Hydrology* 13:157-164.

96 Bergström, S. 1982. HBV-modellen after 10 years. *Nordisk hydrologisk konferens, Förde, Norge, 28-30 juni 1982*, NHP-rapp. nr2. p 212-222.

Fig 51 Bergström, S. Results from model sim ulations (unpubl.).

Chapter 7

Eriksson, E. 1985. *Principles and application of hydrochemistry*. Chapman & Hall, London, 187 pp.

Eriksson, E. & Holtan, H. 1974. *Hydrokemi – Kemiska processer i vattnes kretslopp*. Nordic IHD Report No. 7, Oslo. 124 pp.

Stumm, W. & Morgan, J. J. 1970. *Aquatic Chemistry. An introduction emphasizing chemical equlibria in natural waters*. Wiley-interscience, New York. 583 pp.

Troedsson, T. & Nykvist, N. 1974. *Marklära och markvård.* AW Läromedel, Stockholm. 403 pp.

100 Granat, L. 1988. *Luft- och nederbördskemiska stationsnätet inom PMK. Rapport från verksamheten 1988*. Naturvårdsverket Rapport 3649. 64 pp.

100 Monitor 12. *Försurning och kalkning av svenska vatten*. Naturvårdsverket, Informationsavdelningen. 144 pp.

102 Skyllberg, U. 1993. *Acid base properties of humus layers in northern coniferous forests.* Diss. Umeå. ISBN 91-576-4741-0. 54 pp.

Table 4 After FitzPatrick, E. A. 1980. *Soils – Their formation, classification and distribution*. Longman, London. 353 pp.

104 Loberg, B. 1980. *Geologi – Material, processer och Sveriges berggrund.* Norstedts, Stockholm. 302 pp.

104 Chou, L. & Wollast, R. 1984. Study of the weathering of albite at room temperature and pressure with fluidized bed reactor. *Geochimica et Cosmochimica Acta* 48:2205-2217.

105 Jacks, G. 1973. *Chemistry of some ground waters in silicate rocks.* Inst. för Kulturteknik, KTH, Stockholm. 73 pp.

113 Grip, H. 1982. *Water chemistry and runoff in forest streams at Kloten*. UNGI Rapport nr 58. Naturgeografiska inst., Uppsala univ. 144 pp.

114 Johnsson, N. M., Driscoll, C. T., Eaton, J. S., Likens, G. E. & McDowell, W. H. 1981. Acid rain, dissolved aluminium and chemical weathering at the Hubbard Brook Experimental Forest, New Hampshire. *Geochimica et Cosmochimica Acta* 45:1421-1437.

Fig 54 Data från Svenberg J. 1970. Avrinningsstudier inom Busktorpsbäckens och Vitalampabäckens avrinningsområden. C-uppsats, Avd. för Hydrologi, Uppsala univ. 38 pp.

116 Grip, H. 1982. See ref. p 113.

Fig 55 Rosén, K. pH-studies (unpubl.).

Fig 56 Data från Rosén K. 1982. *Supply, loss and distribution of nutrients in three coniferous forest watersheds in Central Sweden*. Report in Forest Ecology and Forest Soils 41, SLUT, Uppsala. 56 pp.

Chapter 8

119 Lundin, L. 1979. *Kalhuggningens inverkan på markvattenhalt och grundvattennivå*. Rapporter i skogsekologi och skoglig marklära 36, SLU, Uppsala.

119 Grip, H. 1982. See ref. p 113.

119 Rosén, K. 1984. Effect of clear-felling on runoff in two small watersheds in Central Sweden. *For. Ecol. Manage.* 9:267-281.

122 Heikurainen, L. 1973. *Skogsdikning.* Norstedts & Söner, Stockholm. 144 pp.

123 Nömmik, H. & Möller, G. 1981. *Nitrogen recovery in soil and needle biomass after fertilization of a Scots pine stand, and growth responces obtained.* Studia Forestalia Suecica 159. 37 pp.

Fig 57 After Ramberg, L. 1981. Increase in stream pH after drainage. *Ambio* 10(1):34-35.

123 Kloow, L. 1974. *En nitratinventering i tidigare ammoniumnitratgödslad skogsmark*. C-uppsats, Avd. för Hydrologi, Uppsala univ. 26 pp.

124 Thoms, C. & Joelsson, A. 1982. *Nitrat i grundvattentäkter i Sverige*. SNV PM 1589. 21 pp.

124 Granat, L. 1988. See ref. p 100.

125 Troedsson, T. 1980. Ten years acidification of Swedish forest soils. In: D. Drabløs & A. Tollan (Eds.), *Ecological impact of acid precipitation. Proc. Int. conf. Ecol. Impact acid precip.*, Norge 1980, SNSF project. p 184 (abstract).

125 Troedsson, T. & Nilsson, Å. 1984. pH in Swedish forest soils. SNV PM 1853. 24 pp.

125 Eriksson, E. 1985. Acidification of ecological systems. I: I. Johansson (Ed.), *Hydrological and hydrochemical mechanisms and model approaches to the acidification of ecological systems*. IHP Workshop, Uppsala 15-16 September 1984. NHP Report No. 10, p139-245.

126 Nilsson, S. I., Miller, H. G. & Miller, J. D. 1982. Forest growth as a possible cause of soils and water acidification: an examination of the concepts. *Oikos* 39:40-49.

126 Aastrup, M. & Persson, G. 1984. *Förändringar i grundvattnets kemi i Sverige sedan 1968*. Rapport. SGU, Jord- och vattenbyrån. 30 pp.

126 Jacks, G. & Maxe, L. 1984. *Surt grundvatten – en kunskapsöversikt*. SNV PM 1849. 46 pp.

129 *Lakvatten, karaktär – åtgärder – kontroll*. SNV Medd. 2/1982. Liber/ALLF, Stockholm, 1982. 68 pp.

129 Meijer, J.-E. 1981. Lakvatten och avfallsupplag. I: *Hydrologi, markanvändning och vattenkvalitet*. SNV PM 1455, Liber/ALLF, Stockholm. p 37-48.

Fig 59 After McFarlane, D. S., Cherry, J. A., Gillham, R. W. & Susicky, E. A. 1983. Migration of contaminents of a landfill: A case study. 1. Groundwater and plume delineation. *J. Hydrol.* 63:1-29.

130 *Infiltration av avloppsvatten*. SNV Medd. 4/1980. Liber/ALLF, Stockholm, 1980. 52 pp.

130 Nilsson, K. & Englöv, P. 1979. *Avloppsinfiltration*. VIAK, Malmö. 193 pp.

131 *Information om mindre vattentäkter*. Socialstyrelsen SN1, Projekt 1979:5, Stockholm, 1979. 50 pp.

131 Lundgrén, J. & Pousette, J. 1983. *Gårdens vattenförsörjning*. Praktiskt lantbruk 42, LT, Stockholm. 79 pp.

131 *Vatten i Sverige*. SNV PM 1105, Stockholm, 1979. 93 pp.

131 Eriksson, B. 1980. See ref. Chapter. 2.

132 Carlsson, L. & Gustafsson, G. 1984. *Provpumpning som geohydrologisk undersökningsmetodik*. BFR Rapport R41:1984, Stockholm. 124 pp.

133 Nordström, A. 1983. *Vattenförsörjning och avloppshantering*. Diagolos, Lund. 126 pp.

136 Sundby, M. Tracer studies in the Kloten are. (Unpubl.).

136 Gustafsson, E. & Klockars, C.-E. 1981. *Studies on groundwater transport in fractured crystalline rock under controlled conditions using non-radioactive tracers.*KBS Teknisk rapport 81-07. 50 pp.

137 *Information om mindre vattentäkter*. See ref. p 131.

Further reading

Dunne, T. & Leopold, L. B. 1978. *Water in environmental planning*. W. H. Freeman & Co., San Fransisco. 818 pp.

Eriksson, E. 1985. *Principles and application of hydrochemistry*. Chapman & Hall, London. 187 pp.

Eriksson, E. (Ed.) 1984. *Hydrochemical balances of freshwater systems*. Proc. of the Uppsala Symp., Sept. 1984. IAHS Publ. No. 150. 428 pp.

Falkenmark, M. 1983. *Känn ditt vatten*. STFs serie Känn ditt land, nr 14. 64 pp.

Fritzdotter, B., Karlström, U. & Åslund, P. 1984. *Dricksvatten*. Bokskogen, Göteborg. 79 pp.

Johansson, I. (Ed.) 1984. *Nordic glossary of hydrology*. A&W International, Stockholm, 224 pp.

Knutsson, G. & Morfeldt, C.-O. 1978. *Vatten i jord och berg*. Ingenjörsförlaget, Stockholm. 188 pp.

Nordberg, L. & Persson, G. 1979. *Vårt vatten – Tillgång, utnyttjande*. LT's förlag, Stockholm. 144 pp.

Otnes, J. & Raestad, E. 1978. *Hydrologi i praksis*. Ingeniørforlaget, Oslo. 314 pp.

Shaw, M. 1983. *Hydrology in practice*. Van Nostrand Reinhold, Wolkingham, UK. 569 pp.

Troedsson, T. & Nykvist, N. 1974. *Marklära och markvård*. AW Läromedel, Stockholm. 406 pp.

Vattenföring i Sverige. SMHI. LT/ALLF, Stockholm 1979. 403 pp.