CHAPTER 3

Water resources

The water resources used for human consumption are rain, surface water and groundwater. This classification, based on differing characteristics and exploitation methods, is arbitrary, since all these resources are part of the dynamics of the global water cycle.

1 The water cycle

The Earth works as a giant distillation plant, where water evaporates continuously and then condenses and falls again to the Earth's surface. This dynamic process is called the 'water cycle', and can be studied at various levels of time and space. Figure 3.1 shows the overall water cycle of the planet.

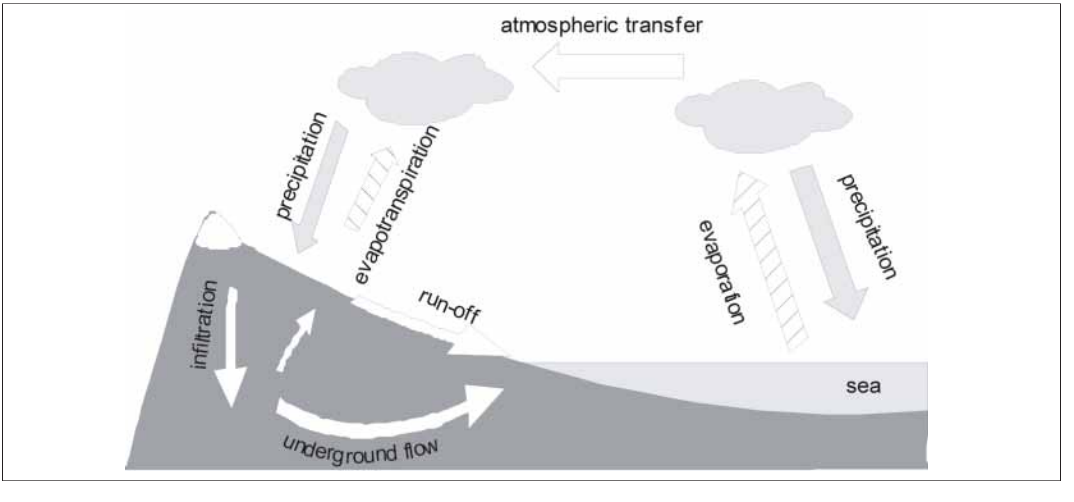


Figure 3.1: The water cycle.

Evaporation occurs mainly at the surface of the oceans, under the influence of solar energy. On the continents, all surface water, as well as shallow groundwater, can be taken up by evaporation. However, the biggest continental contribution is provided by plants in the form of transpiration.

In the atmosphere, water vapour is subject to various winds and transfer movements that feed precipitation.

Part of the continental precipitation is quickly evaporated or transpired, and part rejoins the oceans after having flowed into and fed streams and lakes. A third fraction of precipitation filters into the earth. This water, which becomes groundwater, is not static and continues to be part of the water cycle: it forms groundwater which flows out to feed springs and streams, or flows into the seas (the only exception is fossil sources which are no longer supplied).

The hydrological balance allows the various movements of water to be quantified. This can be written simply as follows:

$$\text{total flow} = \text{precipitation} - \text{evapotranspiration}$$
$$Qt = P - ETR$$

where Q_t = total flow; P = precipitation and ETR = real evapotranspiration (see Annex 6).

Considering the cycle on the planetary scale, De Marsily (1986) proposes the following annual average values:

– continental precipitations, 720 mm;
– evapotranspiration, 410 mm;
– flow of streams and groundwater to the oceans, 310 mm;
– oceanic precipitation, 1 120 mm;
– oceanic evaporation, 1 250 mm.

These figures nearly balance if it is considered that the oceans cover 70% of the Earth's surface, and the continents cover 30%.

The volumes, flows and retention times in the Earth's large reservoirs are shown in Table 3.I. However, local conditions of relief, climate and geology mean that the cycles vary considerably from one region to another. Note that there is an important difference between these figures and those provided by Castany (1982) and De Marsily (1986); they suggest that groundwater does not exceed 2% of the total volume of reserves, so that the ocean fraction therefore amounts to over 97%.

Table 3.I : The Earth's main water reservoirs (from Caron *et al.* 1995).

Reservoir	Capacity (%)	Flow (%)	Average retention time
Oceans	80	78	3 172 years
Atmosphere	0.3	7	4 months
Streams and lakes	0.01	7	5.6 years
Groundwater	19.6	7	8 250 years

2 Rainwater

The harvesting of rainwater is a common practice in many countries. Generally, two types of catchment systems are used.

Artificial and land surface catchments are used to collect rain. At a domestic level, this surface is generally provided by the roof of the house: gutters collect the rainwater and guide it towards storage vessels (jars, barrels or tanks). In the south of Madagascar and in Haiti, there are collective catchment surfaces made of reinforced concrete slabs laid on the ground, with a slope that allows the water

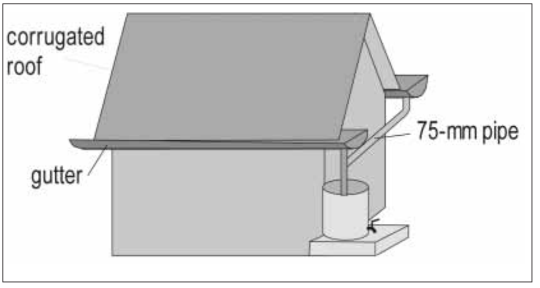


Figure 3.2: Harvesting of water from a roof.

to flow down to underground reservoirs. The bacteriological quality of such water depends on the cleanliness of the collection surface, channels and tank, and on storage and drawing methods

The harvesting of rainwater is also carried out directly: in Cambodia and Myanmar, domestic or collective ponds are dug. Rain fills up those ponds, which can be permanent or temporary, but are generally muddy and biologically contaminated.

For more details regarding rainwater quality, refer to Chapter 4.

2.1 The water layer concept

The concept of a 'water layer' is used to express the relationship between rainfall and flow that is produced on a drainage area. A 1-mm water layer corresponds to 1 l/m^2, so a 100-m^2 roof that collects 10 mm of rain is, in theory, capable of producing 100 x 10 = 1 m^3 of water.

In practice, it is not possible to collect all the rainfall, since part of it is evaporated and part of it is lost (leakage, overflow etc). For domestic or small collective rainwater catchments, Pacey and Cullis, 1986 propose the following harvested-water/rainfall ratios:

– tile: 0.8 to 0.9;
– corrugated metal sheet: 0.7 to 0.9;
– plastic sheets: 0.7 to 0.8;
– concrete: 0.6 to 0.8;
– brick: 0.5 to 0.6.

Therefore, 10 mm of rain falling on a 100-m^2 corrugated sheet roof produces: 100 x 10 x 0.7 = 700 litres of water (Figure 3.2).

In order to design a rainwater-harvesting system, it is advisable to collect the available rainfall data and/or to install a network of rain gauges or rain meters.

2.2 Rainfall measurement

The density of the network of rain gauges is selected on the basis of the particular measurement objective and the environmental conditions. For the same climate zone, rainfall can be affected by numerous factors: altitude, exposure of slopes, distance from the sea etc. In a reasonably small and relatively homogenous area, such as certain refugee camps, the installation of one or two rain gauges may be enough. Over a district, the network must be larger in order to incorporate environmental differences.

The measurement sequence must be as long as possible; up to several years for complete accuracy. It is therefore sensible to set up a network of rain gauges as soon as possible, since certain water-supply projects will exist for several years even if a permanent installation is not envisaged at the beginning of the programme. A station that automatically registers rainfall and temperature data can be installed in order to facilitate recording.

Then, it is necessary to give to each rain gauge a geographical area of representation. Several different methods can be used (see Annex 6), but the main ones are:

– the arithmetical mean, which is the least accurate but the easiest to use;
– the Thiessen method (surface weighting), usable in relatively flat areas;
– the isohyetal method, more accurate in the case of broken terrain.

3 Surface water

Surface water comes in many forms, and methods of exploitation are very varied.

Part of the rainwater that arrives on the ground runs off. Sometimes this water is intercepted by some man-made structure, especially in areas with a dry climate. For example, in a sahelian area, run-off that concentrates at low points is retained by a dam and used for human and animal needs. These ponds can be temporary or permanent, but they are generally muddy and polluted by faecal matter. They are difficult to protect, but being possibly the only available resource, they are also vital. Storage can be in underground reservoirs, which are called *birkad* in East Africa (Ethiopia, Somalia). Animals do not have a direct access to the reserve, to protect it from contamination, and evaporation is reduced.

Temporary watercourses, sometimes called by their Arabic name, *oued* or *wadi*, are characterised by their usually torrential flow, with a strong erosion capacity upstream and a sedimentation zone downstream. Perennial underground flow frequently accompanies these temporary watercourses, so their exploitation is generally linked to groundwater (see Chapter 5).

Permanent watercourses are used all around the world, since they are perennial and easy to exploit directly. Their quality varies greatly from one situation to another: they are very vulnerable to surface pollution, but have an auto-purification capacity linked to the biological conditions of the aquatic environment. They are usually closely linked to groundwater, and can be exploited indirectly via wells and boreholes in their alluvial deposits.

Lakes are extended areas of water without a direct link to the sea. They are formed in topographic depressions fed by surface-water flow and direct rainfall, or upstream of dams that block surface-water flow. Lakes are therefore considered to have catchment areas in the same way as watercourses. Lakes are widely-used resources of very varied quality. As with any other surface-water source, they are vulnerable to pollution.

For more details on surface-water quality, refer to Chapter 4.

3.1 The catchment area concept

Surface water flows by gravity and so the direction of flow depends on the topography. A catchment area is defined as the group of slopes inclined towards the same watercourse, into which they all contribute water. Catchment areas are divided by watersheds, and are identified on the topographic map by crest-lines. They are usually drained by an out-flowing watercourse, in which case they are known as exoreic basins, but can also be closed (without an outlet) when they are termed endoreic basins.

Exploitation of surface water can take place after assessing its quality (see Chapter 4) and the quantity of water available.

3.2 Run-off assessment

Run-off water is the fraction of rainfall that does not infiltrate and is not returned to the atmosphere by evapotranspiration, and it is therefore necessary to know the various balance terms to calculate it. Nevertheless, a quick assessment can be made as in Table 3.II, which establishes the orders of magnitude of run-off water on natural ground depending on average annual data. Note that these figures do not take into account the slope of the terrain or the intensity of the rainfall. It is therefore necessary to use them with caution, and favour the balance method or, even better, direct observation of the working systems.

Table 3.II : Run-off assessment.

Rainfall (mm)	Potential evapotranspiration (mm)	Run-off water (% of the rainfall) fairly impermeable terrain	very permeable terrain
> 1 100	–	16.5	6.5
900 to 1 100	–	13	6.5
500 to 900	< 1 300	10	5
500 to 900	1 300 to 1 800	10	3
400 to 500	1 300 to 1 800	8	1.5
250 to 400	< 1 800	5	0
250 to 400	> 1 800	3	0

A 100 hectare (10^6 m^2) rainwater catchment made of fairly impermeable compacted soil, in a climate with an average annual rainfall of 750 mm and potential evapotranspiration of 1 000 mm (for calculation of potential evapotranspiration, see Annex 6), can produce about 10^6 x 750 x 10% = 75 000 m^3 of run-off water.

3.3 Flow-rate measurement

Various methods are used to calculate the flow rates of watercourses. The technique chosen depends on the conditions of flow and the range of flow rates to be measured (Table 3.III).

Table 3.III : Techniques of flow-rate measurement.

Direct measurement of the flow rate (recipient and chronometer)	Flow rate < 35 l/s
Measurement of current velocity (float and chronometer) or current meter	Laminar flow
Measurement of the water height (weir)	Laminar flow
Salt-gulp method (salt dilution)	Turbulent flow 100 l/s < flow rate < 3 m^3/s

3.3.1 CHRONOMETER AND CONTAINER

For accuracy, the duration of collection must be between 30 and 60 seconds (Figure 3.3). The recipient is calibrated and its capacity is chosen depending on flow rate (Table 3.IV). It is advisable to carry out several measurements and to take an average.

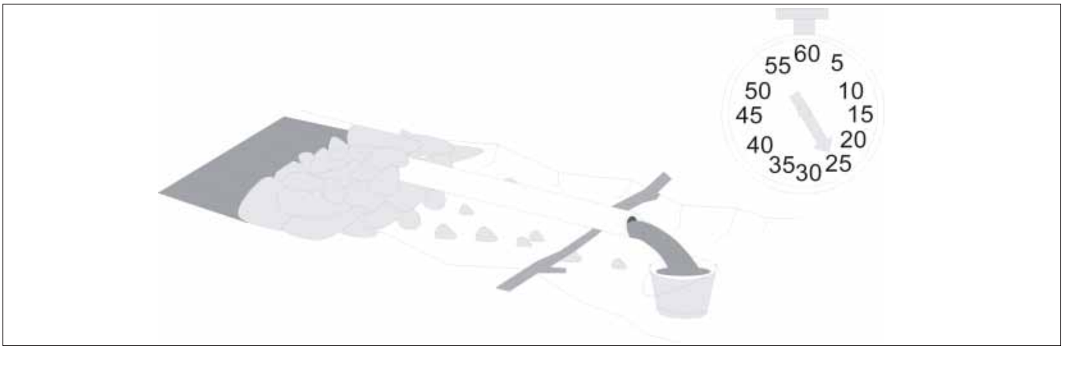


Figure 3.3: Direct flow measurement.

Table 3.IV: Volume of recipient depending on the flow rate.

Flow rates Q (m^3/h)	Volume of calibrated recipient (l)
Q < 3.6	20
3.6 < Q < 9	50
7.2 < Q < 18	100
14.4 < Q < 36	200

3.3.2 CHRONOMETER AND FLOAT

The flow rate Q is proportional to the velocity of the water and its flow section. This method is only valid for laminar flow (see Annex 6), and measurement accuracy is poor, since velocity is not constant across the section of the flow.

In practice, the flow section (taken perpendicular to the flow) is measured, and then, using a chronometer, the velocity of passage of a floating body (cork or wood) is measured over a known distance. The float must be launched in the direction of measurement, in the middle of the channel. The velocity obtained is a surface velocity, generally higher than the average flow-section velocity. Then the calculation is corrected with a coefficient, B, such that:

$$Q = B.V.S$$

where Q is flow rate (m^3/s), V is velocity (m/s), S is normal flow section (m^2) and B is a coefficient between 0.6 and 0.8.

3.3.3 WEIR MEASUREMENT

A weir is a device that enables flow rate to be calculated from the thickness of a water layer. The principle is to install a board or a metallic plate perpendicular to the flow. The thickness of the water layer measured above the weir is proportional to flow rate, and depends on the characteristics of the device. The flow must be laminar. If this is not the case, it is possible to smooth it by a using a relatively high plate. The thickness of the water layer must be measured at a distance from the weir equal to at least 5 times the maximum thickness of the water layer (Figure 3.4).

The shape of the weir is chosen depending on the range of flow rates to be measured: the weir must allow a large variation in water height to be obtained for a small variation of flow rate. The most usual shapes are triangular and rectangular, known as thin-plate weirs (see Box 3.1). Triangular weirs are generally classified depending on their opening angle. Table 3.V gives the application range of the most efficient thin-plate weirs depending on their shape.

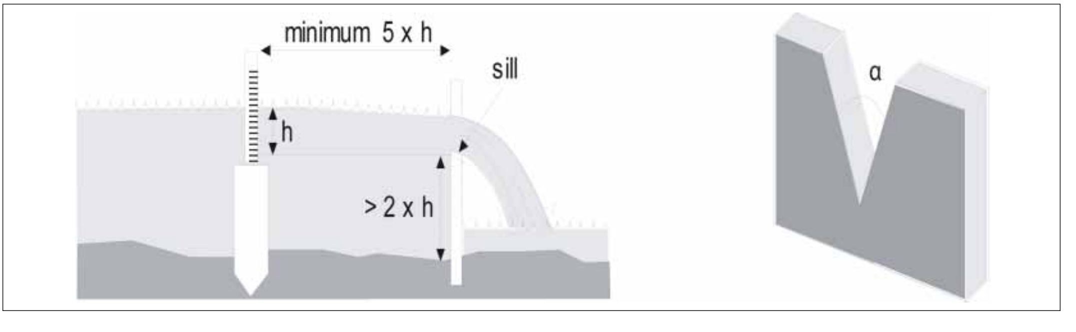


Figure 3.4: Thin-plate weir.

Table 3.V: Choice of thin-plate weir.
T: triangular. R: rectangular. L: threshold length.

Flow rates to be measured (l/s) min.	max.	Type of weir
0.3	40	T – a = 30°
1	100	T – a = 60°
20	150	T – a = 90°
20	200	R – L = 0.3 m
60	500	R – L = 1 m
300	1 500	R – L = 2 m
800	4 000	R – L = 5 m

In the field, the choice of weir must be followed by the construction of a design chart to read the flow rate quickly depending on the height of the water layer (Figure 3.5). Box 3.1 gives the most commonly used formulae.

3.3.4 SALT-GULP METHOD

This method is well suited to turbulent flow conditions. It consists of analysing the characteristics of passage of a volume of saline solution introduced into the watercourse (Figure 3.6 and Box 3.2). Two people are needed: the first prepares a solution of salt and then introduces it into the watercourse. Downstream, the second person measures the variation in conductivity induced by the passage of the salt, using a common conductimeter.

The NaCl solution is prepared without exceeding the solubility threshold of 300 g/l at 20° C. The quantity of salt to be used depends on the flow rate of the stream and its base conductivity: the objective is to double the conductivity of the stream in order to be able to measure the passage of the salt cloud. For flow rates of 100 to 3 000 l/s, a quantity of 1 kg of NaCl per 100 l/s allows the base

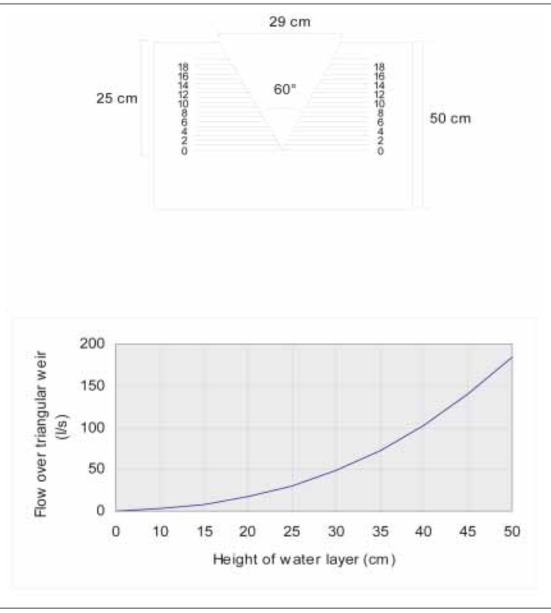


Figure 3.5: Design chart for triangular thin-plate weir (Bazin formula, 20-cm blade).

Box 3.1
Thin-plate weirs.

The weirs are referred to as ‘thin-plate’ since the thickness of the threshold is less than h/2, where h is the thickness of the water layer.

The general flow rate formula for weirs is:

$$Q = \mu \, . \, l \, . \, h \sqrt{2gh}$$

where Q is flow rate (m^3/s), μ weir flow-rate coefficient (non-dimensional), l weir width (m), h water layer (m) and g the acceleration due to gravity (9.81 m/s^2).

Rectangular weir without lateral constriction
When the width of the weir is the same as that of the supply channel, the weir does not reduce the width of the water layer.

Bazin’s formula, which is widely used, defines the coefficient μ:

$$\mu = 0.405 + \frac{0.003}{h}\left[1 + 0.55\,\frac{h^2}{(h+P)^2}\right]$$

where P is blade height (m). This formula is applicable for values of P lying between 0.2 and 2 m and h between 0.1 and 0.6 m.

The SIA formula gives:

$$\mu = 0.410\left[1 + \frac{1}{1000\,h + 1.6}\right]\left[1 + 0.5\,\frac{h^2}{(h+P)^2}\right]$$

It is applicable when P is higher than h and h lies between 0.025 and 0.8 m.

Rectangular weir with lateral constriction
When the supply channel is wider than the threshold, the SIA gives the following formula:

$$\mu = \left\{0.385 + 0.025\left(\frac{l}{L}\right)^2 + \left[\frac{2.41 - 2\,(l/L)^2}{1000\,h + 1.6}\right]\right\}\left[1 + 0.5\left(\frac{l}{L}\right)^4\left(\frac{h}{h+P}\right)^2\right]$$

Where L is the width of the supply channel and l the width of the weir, both in metres.

This formula is usable when $P \geq 0.3$ m; $l > 0.31$ L; 0.025 L/T $\leq h \leq 0.8$ m and $h \leq P$.

Triangular weir
The general formula is:

$$Q = \frac{4}{5}\,m\,h^2\sqrt{2gh}\;\mathrm{tg}\,\frac{\alpha}{2}$$

where Q is flow rate (m^3/s), μ flow-rate coefficient for Bazin’s rectangular weir without lateral constriction, h the thickness of water layer (m) and α the weir angle.

conductivity of a fairly low mineralised stream (conductivity about 200 μS/cm at 25 °C) to be multiplied by about 2.5. The solution is injected quickly, in an area where turbulence facilitates mixing with the stream water.

The distance between the point of injection and the point of measurement is chosen in order to obtain a cloud-passage curve [c = f(t)] with a Gaussian distribution (Figure 3.6). An average distance of 80 to 100 m usually gives good results. It is important to choose a section of watercourse where the losses and dead areas are as few as possible so as not to lose or immobilise part of the NaCl solution.

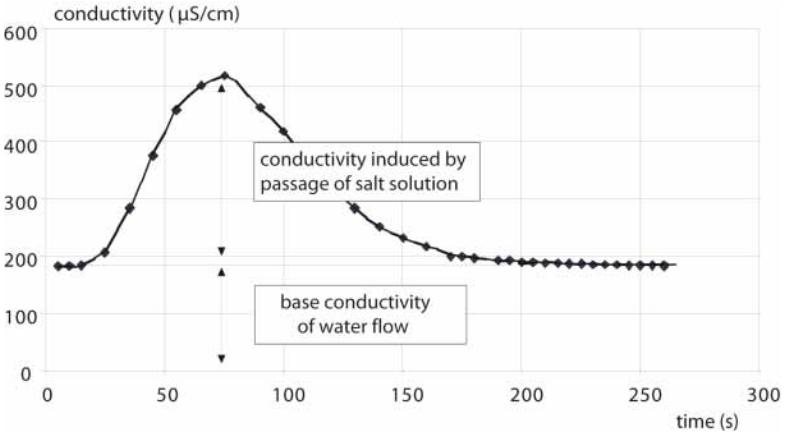


Figure 3.6: Salt-gulp method.

Box 3.2
The salt-gulp method.

During the period of time dt, the conductivity of the water measured during the passage of the cloud of salt is χ. This conductivity can be expressed in terms of concentration:

$$c = k\chi$$

where c is concentration (g/l), χ is conductivity (µS/cm) and k is a conversion factor.

In the same time dt, the volume of water passing is Q.dt. The average mass of salt that passes during dt is therefore:

$$\chi \,.\, k \,.\, Q \,.\, dt$$

Over the whole period of passage of the cloud of salt:

$$M_{NaCl} = \Sigma\, k \,.\, Q \,.\, dt$$

Or: $$M_{NaCl} = k \,.\, Q \,.\, \int \chi \,.\, dt$$

The flow rate of the watercourse is therefore given by:

$$Q = \frac{M_{NaCl}}{k \,.\, \int \chi \,.\, dt}$$

where Q is flow rate (l/s), M is the mass of salt (g), k is the conversion factor and $\int\chi$ dt is the integral over the whole passage period of the cloud [(µS/cm).s].

Conversion factor k
The factor k varies with the temperature and mineralisation of the water. For water at 25° C (automatic correction of the conductimeter) and slightly mineralised ($\chi_{\text{untreated water}} \approx 200$ µS/cm), $k = 5.48 \times 10^{-4}$.

Average concentration
The formula $Q = M_{NaCL} / k. \int\chi.dt$ is not very different from $Q = M_{NaCl} / \bar{c}.t$ when the curve $\chi = f(t)$ is Gaussian, where $\bar{c}$ is the average concentration of NaCl over the period t ($\bar{c} = k \,.\, \bar{\chi}$) and t is the time of passage of the cloud of salt.

It is advisable to draw the curve $\chi = f(t)$ on graph paper and to integrate it (small-squares method) in order to obtain a sufficient degree of accuracy. Calculation by average conductivity is not as useful as when the curve is Gaussian.

The flow rate of the watercourse is given by:

$$Q = \frac{M_{NaCl}}{k \,.\, \bar{\chi} \,.\, t}$$

where Q is the flow rate of the watercourse (l/s), M is the mass of salt used to make the solution (g), $\bar{\chi}$ is the average conductivity induced by NaCl over the period t (µS/cm), k is the conductivity/concentration conversion factor ($k \approx 5.48 \,.\, 10^{-4}$) and t is the time of passage of the cloud of salt (s).

3.3.5 PROPELLER DEVICES

Propeller devices, often called current meters, consist of a shaft with a propeller connected to the end. The propeller is free to rotate, and the speed of rotation is proportional to the stream velocity if the flow is laminar.

The number of rotations of the propeller produced by the flow, on a normal section of the stream at various depths and distances from the bank, is recorded. After conversion of the number of rotations into velocity, it is possible to calculate the flow rate passing through the measured section when the section area is known.

4 Groundwater

According to Castany (1982), nearly 60% of the drinking water reserves of the planet are stored in the form of ice or snow, less than 0.5% in continental surface water and 40% in the form of groundwater. The use of this resource is therefore a vital matter for many populations.

4.1 Water within rock

Rocks capable of containing water and allowing it to flow easily are called aquifers. An aquifer is not necessarily a homogenous geological group: it can be composed of different rocks or strata. An aquifer has an area saturated with water, and sometimes a non-saturated area. It is spatially restricted by an impermeable rock at its base (the wall or substratum), sometimes by an impermeable rock above it (the roof), and by lateral restrictions.

The groundwater is all the water contained in the aquifer, generally supplied by useful precipitation (the fraction of precipitation that infiltrates and feeds the aquifer) and the infiltration of surface water (streams and lakes).

Aquifers are not static: part of the water leaves the aquifer in the form of springs, feeding surface water (streams, lakes, seas), by pumping, or by direct evaporation (Figure 3.7).

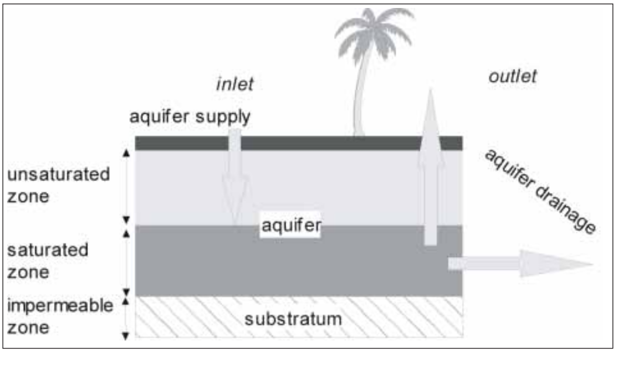


Figure 3.7: Storage and flow in an unconfined aquifer.

To describe an aquifer, a group of parameters related to its nature, geometry and functions (water storage and outflow capacity) are used (Box 3.3).

Box 3.3
Representative sample volume.

A porous environment can be physically defined by three characteristics:
– *continuity/discontinuity:* an environment is called continuous if its voids are interconnected in the direction of the flow;
– *homogeneity/heterogeneity:* an environment is homogenous if its characteristics are constant in the direction of the flow;
– *isotropy/anisotropy*: an environment is called isotropic when its physical characteristics are constant in the three dimensions.

The concepts of porosity and permeability, defined in the following sections, are linked to the scale of observation. In fact, a fissured rock is considered non-homogenous when observed at a centimetre scale, but it can be considered homogenous if observed at a kilometre scale.

The concept of representative sample volume (RSV) describes the characteristics of an aquifer: for example, a RSV at the cm^3 scale is used for sands or gravels, and at the m^3 or km^3 scale for fissured rock.

Even though this concept of RSV is often used, it presents considerable problems. Certain authors therefore define the concept of random function and study the environment as the development of random phenomena (De Marsily 1986).

4.1.1 WATER STORAGE

The quantity of water stored in an aquifer at a given instant depends on the volume of the reservoir and its capacity to bear water (Figure 3.8).

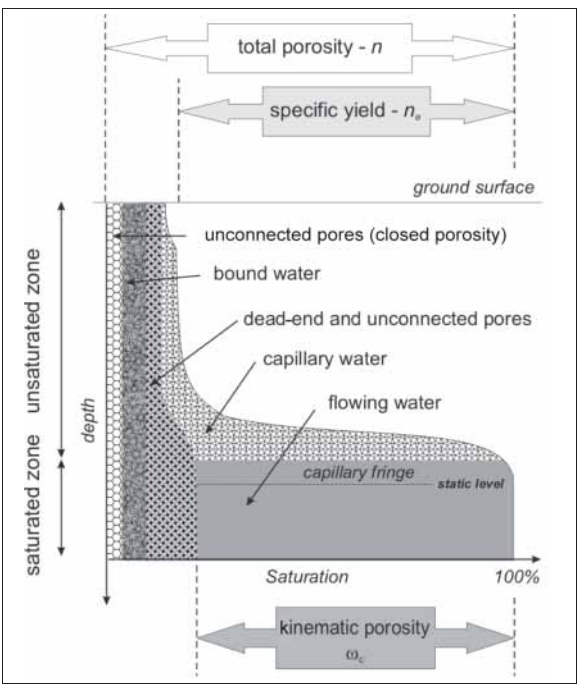


Figure 3.8: Groundwater type and storativity.

Total porosity

Most rocks naturally contain a certain percentage of voids that can be occupied by water. Total porosity (n) is the capacity of an aquifer to store water so that (Figure 3.8):

$$n = \frac{\text{void volume}}{\text{total volume}}$$

Generally, total porosity is expressed as a percentage.

The voids are not necessarily interconnected and the water is not always free to flow within the rock: total porosity is a necessary condition for water flow, but it is not the only one.

Generally, two types of porosity can be distinguished, depending on the geological nature of the aquifer. Microporosity (also primary or interstitial porosity) refers to the intrinsic porosity of the rocks, while macroporosity (fracture or secondary porosity) refers to the porosity induced by fissures, fractures or karstic developments of consolidated rocks. Certain aquifers, such as some sandstones or carbonated rocks, have both interstitial and fracture porosity simultaneously.

The total porosity of unconsolidated rock (sand, gravel) is controlled by its granularity, which is generally studied by screening a volume of rock, and is expressed in a normalised particle-size curve (Box 3.4). The more uneven the granularity, the lower the porosity. In the case of even granularity, grain size has no effect on total porosity; it is the arrangement of the grains which controls total porosity, which theoretically varies between 26 and 48%.

Relationship between water and rocks

Experience shows that part of the water contained in the aquifer cannot easily be extracted. A distinction can therefore be drawn between water bound to the rock by forces of molecular attraction and free water, which can move under the influence of gravity or pressure gradients.

According to De Marsily (1986), bound water corresponds to a layer of water molecules about 0.1 μm thick adsorbed onto the surface of the rock grains, and to a layer of water about 0.4 μm thick, known as a film, which is also subject to a measurable attractive force.

The bound fraction of the water is larger when the specific area (S_p) of the reservoir is greater:

$$S_p = \frac{\text{grain surface}}{\text{total volume}}$$

For example, medium sand presents a specific area of 10-50 cm^2/cm^3, and clay presents 500 to 800 cm^2/cm^3. This explains the fact that certain clays contain a great amount of water, but that water is bound and therefore cannot be moved by pumping; these clays are therefore regarded as impermeable.

For unconsolidated rock, the specific area is governed by the granularity. The diameter d_{10} and the uniformity coefficient are generally used to characterise the sample (Box 3.4).

The presence of capillary water suspended by capillary forces just above the saturated area is also of note. This ‘capillary fringe’ is well known by well diggers, as it is the first sign of water.

Kinematic porosity

In the saturated zone, free water consists of the water free to flow plus the water in the unconnected and dead-end pores (Figure 3.8). Kinematic porosity ω_c (often called effective porosity) represents the free-flowing water in the saturated zone; it is often used in flow and transport modelling, and is defined as the ratio between the Darcian velocity and the true linear velocity (see following sections).

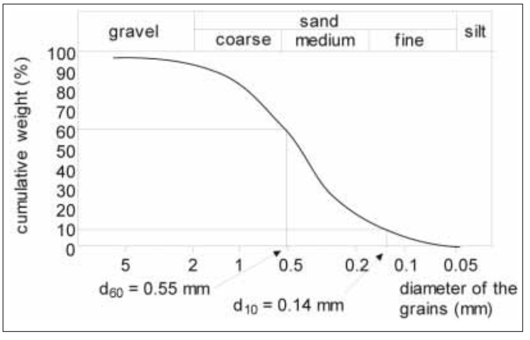
Note that unconnected and dead-end pores can play an important role in karstic and hard-rock contexts, but they are usually negligible in unconsolidated sediments.

Storativity

The groundwater fraction which is of interest for abstraction is quantified by the storativity.

Box 3.4
Particle-size analysis.

Particle-size analysis facilitates:
– study of the characteristics of the grains and voids of a porous environment;
– classification of the rocks according to a nomenclature;
– calculation of the characteristic particle-size parameters;
– accurate definition of the borehole equipment plan.



Particle-size analysis is based on a sample of the formation being studied. By successive screening, the various sizes of particle are separated and then weighed. The results are then represented in a standard curve (Figure 1), plotting the particle size in mm logarithmically on the horizontal axis and the percentage cumulative weight of successive screenings on the vertical axis.

Figure 1: Particle-size curve.

Rock classification
There are various classifications of unconsolidated rocks depending on their particle size, such as this one, proposed by Castany (1982):

Designation	Diameter of the particles (mm)
Pebbles, stones, blocks	> 16
Gravel	16 to 2
Coarse sand	2 to 0.5
Medium sand	0.5 to 0.25
Fine sand	0.25 to 0.06
Silt	0.06 to 0.002
Clay	< 0.002

Characteristic parameters
The majority of natural samples are a mixture of different particle sizes. The particle-size curve defines the whole group.

The uniformity coefficient, CU = d_{60}/d_{10}, allows the slope of curve to be calculated. If the CU < 2, the particle-size distribution (granulometry) is referred to as uniform, otherwise it is termed stepped.

The characteristic diameter, d_{10} is a conventional diameter, usually used to represent a sample.

In a confined aquifer (see Section 4.2), water release is related to the water-expansion and aquifer-compaction effect attributed to aquifer pressure changes due to pumping. This elastic water release is quantified by the storage coefficient as (De Marlisy 1986):

$$S \approx \left[\rho \,.\, g \,.\, n \,.\, \left(\beta_1 \,.\, \frac{\alpha}{n}\right)\right] .\, e$$

where ρ is the water mass/volume ratio, g is gravity acceleration, β_1 is the water compressibility, α is the aquifer compressibility, n is the total porosity and e is the saturated thickness.

When an unconfined aquifer is desaturated by well pumping, the quantity of extracted water is determined both by elastic water-release phenomena and by gravity-water phenomena. The storage coefficient is then:

$$S \approx \left\{ \left[\rho \ . \ g \ . \ n \ . \left(\beta_1 \ . \ \frac{\alpha}{n} \right) \right] . \ e \right\} + n_e$$

where the term [ρ.g.n.(β_1.α/n)] represents the elastic phenomena and n_e the specific yield parameter, which is:

$$n_e = \frac{\text{gravitational water volume}}{\text{total volume}}$$

Because the value of the storage coefficient is several orders of magnitude smaller than the value of specific yield, the storativity of an unconfined aquifer is usually taken to be equal to the specific yield. It means that the water release (or storage) is due to the gravitational drainage of the aquifer because of the falling water level while pumping. This is why specific storage is commonly called drainage porosity, gravity porosity or effective porosity.

Note that a confined aquifer that is desaturated by pumping (if the dynamic level falls below the bottom of the confining layer) becomes partially unconfined. The specific drainage (S_d) is the parameter that quantifies the volume of water that could potentially be released by a confined aquifer due to gravity (Lubczynski & Roy 2003).

In a unconfined aquifer, the phenomenon of seepage takes place when there is a drop in the water level (particularly due to pumping). In a confined aquifer, the release of the water is not caused by gravity, but by decompression of the water and the reservoir generated by a reduction in head (see Annex 12). As the compressibility of both aquifer and water is very small, we no longer refer to effective porosity but to a storage coefficient (S) defined as the volume of water released by a vertical prism of aquifer material of unit section as a result of a unit change in head; this factor is non-dimensional.

Assessment of porosity

Porosity can be measured by tests carried out in the laboratory. However, in ACF programmes, it is more useful to calculate the porosity by an *in-situ* method.

The most reliable method of calculating the storage coefficient (S) and/or the specific yield (S_y) is to carry out test pumping (see Chapter 6). Geophysics can also be used in favourable contexts to estimate the porosity of the medium (see Chapter 5). Electrical prospecting allows a relationship to be established between the resistivity of a rock and its porosity, using Archie's formula in unconsolidated sediments without clay (which is not common); MRS can estimate the storativity if calibrated with pumping-test data.

Table 3.VI gives some values of total porosity and specific yield for various reservoirs.

Table 3.VI: Total and specific yield, diameter d_{10} (from Castany 1982).

Reservoir	d_{10} (mm)	n (%)	n_e (%)
Medium gravel	2.5	45	40
Coarse sand	0.25	38	34
Medium sand	0.125	40	30
Fine sand	0.09	40	28
Silty sand	0.005	32	5
Silt	0.003	36	3
Clay	0.0002	47	–

4.1.2 WATER FLOW

In addition to its storage capacity, an aquifer is capable of conducting water. In the reservoir, the flow is determined by three groups of parameters: permeability and transmissivity; hydraulic head and gradient; speed rate and flow rate.

Permeability and transmissivity

Permeability is the capacity of a reservoir to allow water to pass through it under the influence of a hydraulic gradient. It is classified by intrinsic permeability k, expressed in m^2 or in a unit called the Darcy which takes into account the characteristics of the reservoir, and the hydraulic conductivity K, expressed in m/s, and which takes into consideration the characteristics of both the reservoir and the fluid, as follows:

$$K = \frac{g}{\nu} \cdot \frac{d^2}{a} = \frac{g}{\nu} \cdot k$$

where K is the hydraulic conductivity, which depends on both the liquid, by (g/ν) and the porous environment, by (d^2/a); k is the intrinsic permeability, which depends only on the porous environment, by (d^2/a); g is the acceleration due to gravity; ν is the kinematic viscosity of the fluid, d^2 is a dimension that characterises the environment, and a is a non-dimensional constant.

In hydrogeology, it is usually assumed that the characteristics of the water are constant (dynamic viscosity and specific gravity), so it is possible to work directly with hydraulic conductivity. Note that the main parameter that influences hydraulic conductivity in hydrogeology is the temperature of the water (through its viscosity); this has a great influence because a reduction of 40% in hydraulic conductivity is caused by a change of the water temperature from 25 to 5°C.

The coefficient K can be calculated by *in-situ* or laboratory methods. *In-situ* methods are carried out at a borehole:

– pumping test (see Chapter 6): this is recommended, as it takes into account the heterogeneity of the terrain;

– Lefranc tests for continuous environments and Lugeon test for fractured environments: these are methods that allow calculation of local permeability that applies only within the surroundings of the well;

– tests used in sanitation (see Chapter 13).

Laboratory methods involve the use of:

– a constant-level ($K < 10^{-4}$ m/s) or variable-level permeability meter ($K > 10^{-4}$ m/s), (see Darcy's experiment);

– Hazen's formula, which defines permeability for a granulometrically continuous environment as: $K_{Hazen} = (0.7 + 0.03t_F) \cdot d10^2$ where t_F is water temperature (°F) and d_{10} the grain diameter (cm) such that 10% of the elements are smaller.

Table 3.VII gives orders of magnitudes of permeability. The limit of impermeability is generally based on the coefficient $K = 10^{-9}$ m/s.

Finally, transmissivity is a parameter that expresses the productivity of an aquifer, so that:

$$T = K \cdot e$$

where T is transmissivity (m^2/s), K is hydraulic conductivity (m/s) and e is the thickness of the saturated aquifer (m).

Hydraulic head

In a porous environment, energy is generally expressed as 'head', or height, so that the unit is a length (see Annex 12). The hydraulic head is expressed in relation to sea level, whereas the piezometric head is expressed in relation to an origin that must be defined: for example, piezometric level may be measured in relation to the ground, the top of borehole casing etc.

Table 3.VII : Orders of magnitude of hydraulic conductivity K (m/s) (from Brassington 1998).

10^{-10} impermeable	10^{-9}	10^{-8}	10^{-7}	10^{-6}	10^{-5} semi-permeable	10^{-4}	10^{-3}	10^{-2}	10^{-1} permeable
Unconsolidated rocks									
solid clay		silt, clay, mixture silt/clay			fine sand		clean sand, mixture sand/gravel		clean gravel
Consolidated rocks									
granites, gneiss, compact basalts		sandstone, compact limestone, argillaceous schists				sandstone, rocks fractured			karst

Flow and velocity of water

In 1856, Henri Darcy experimentally established the formula that bears his name (Figure 3.9). It defines the rate of flow passing through a porous environment, so that:

$$Q = K \cdot A \frac{\Delta h}{L}$$

where Q is the flow rate through a porous environment (m^3/s), K is the hydraulic conductivity (m/s), A is the section of the porous formation normal to the flow (m^2), L is the length of the sandy formation transited by the fluid (m), and Δh is the difference between the hydraulic heads upstream and downstream of the length (m) under consideration.

The hydraulic gradient is defined as $i = \Delta h/L$, a non-dimensional factor, and produces the simplified Darcy's equation:

$$Q = K \cdot A \cdot i$$

Darcy's equation is valid under the following conditions:

– continuous, homogenous and isotropic environment (Box 3.3),
– laminar flow,
– permanent flow (see Annex 6).

In practice, underground flow often meets these conditions, except in a highly heterogeneous medium such as a karstic environment or in the immediate surroundings of a pumped well. The transition from laminar to turbulent flow can be estimated from the hydraulic gradient by an empirical formula. Thus, De Marsily (1986) defines the gradient limit for Darcy's equation validity so that $i = 1/15\ (K)^{1/2}$.

If Darcy's equation is now expressed per unit area, the filtration rate, or unitary flow rate, is obtained as $q = K \cdot i$ (m/s). The rate of flow passing through an aquifer can therefore be calculated using Darcy's equation. In fact, knowing K, it is simple to calculate the unit flow rate between two points from the static levels measured in two wells. Furthermore, if the geometry of the aquifer can be assessed (from boreholes, geophysics or geological characteristics), the flow rate passing through the reservoir can be determined (Figure 3.10).

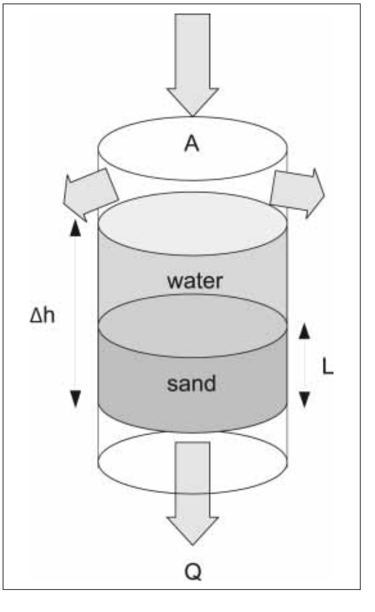


Figure 3.9: Darcy's experiment.

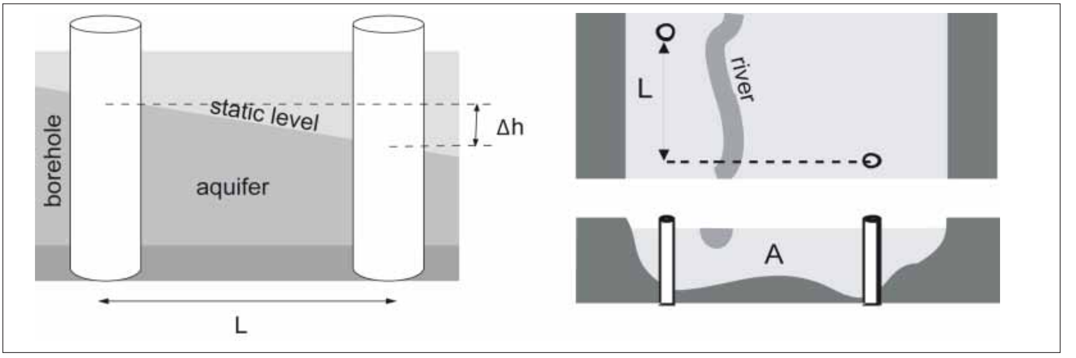


Figure 3.10: Application of Darcy's experiment.

Note that this filtration velocity is notional, since it assumes that the water uses the whole section of the aquifer to circulate. Part of this section is in fact occupied by the materials of the aquifer itself. If we assume that the water does not circulate through the whole section, but only through the space that can be estimated from specific yield, the effective velocity is obtained as:

$$V_e = \frac{q}{S_y}$$

These hydrodynamic velocities provide orders of magnitude and help to compare the various environments, but they remain theoretical. To obtain more real velocities, it is necessary to resort to hydrokinematics, in which tracing methods are used to reveal the realities of the flow.

Piezometry

The map of the piezometric surface is established using measurements carried out in a group of wells in an even context. It is a basic document that imparts an understanding of the dynamics of an aquifer. Such a map is valid only at a given moment, and is meaningful only when it is based on piezometric lines from measurements carried out in the wells, boreholes or springs corresponding to a single aquifer.

In humanitarian programmes, piezometric levels can be obtained by measuring, over a short period of time, the levels in wells at rest. The various measurement points must be located in space: GPS co-ordinates for x and y, and levels or topographic mapping for z. These values enable the piezometric map to be drawn (Figure 3.11): manual interpolation is the most reliable (see Annex 6).

Piezometric maps are interpreted as follows:

– the curvature of lines of the same level (isopiezometric lines or groundwater contours) differentiates the divergent and convergent flows. A divergent flow can indicate a line of separation of the groundwater or a recharge zone, whereas a convergent flow underlines a preferential flow axis. Rectilinear flows are rare, but are characteristic of a flat, homogenous aquifer, with a constant thickness;

– the spacing of the isopiezometric lines determines whether or not the flow is even. A narrowing of the lines means an increase in hydraulic gradient (according to Darcy, Q = KAi, so if i increases and the flow rate is constant, then A decreases – rise of the substratum or reduction in size, or K decreases – change of facies). On the contrary, if the isopiezometric lines separate and Q is constant, then A or K increase.

It is thus possible to calculate the variation of certain parameters of the aquifer on the basis of the study of the piezometric map. Indeed, according to Castany (1982), the lateral transmissivity variations (thus of K or e) are more frequent that the variations of flow rate in the aquifer (a radical

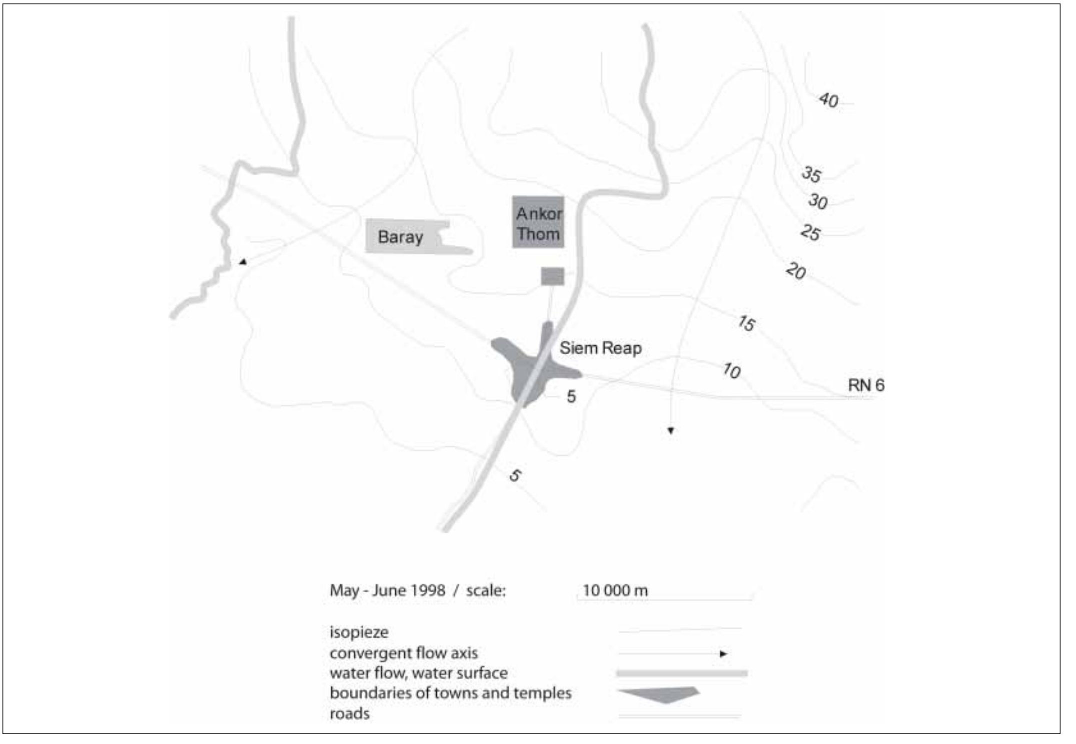


Figure 3.11: Piezometric map (Prasat Bakong, Siem Reap district, Cambodia, ACF, June 1998).

change of flow rate can be identified by a particular supply: stream, lake etc.). The result is that iso-piezometric lines with a curvature indicating a convergent flow and with increasing downstream spacing correspond to a drainage axis with a permeability or thickness increasing downstream. This is a favourable location for a well or borehole (Figure 3.12).

Monthly piezometric monitoring of newly constructed boreholes and wells also allows the impact of extraction on the aquifer to be measured, and seasonal variations of the piezometric level to be estimated (see Chapter 6).

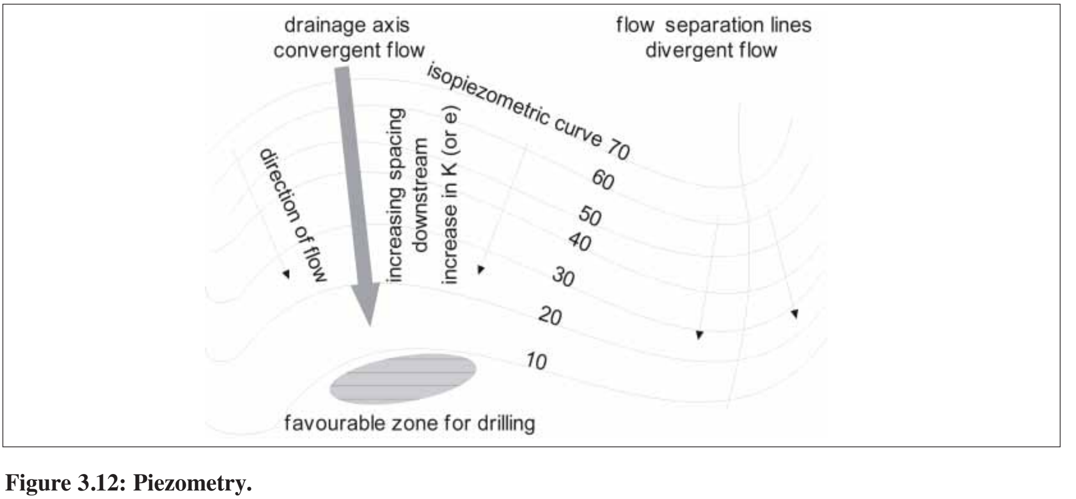


Figure 3.12: Piezometry.

4.2 Main aquifer systems

Table 3.VIII lists the principal hydrodynamic parameters that characterise aquifers. Aquifers are also defined by their geometry and by their limits. The base limit consists of an impermeable layer (in theory, $K < 10^{-9}$ m/s), called the wall or substratum. As for the upper limit:

– if the aquifer is covered by impermeable material (roof) or barely permeable material (sometimes called an aquiclude), and the piezometric head is higher that the elevation of the roof, the aquifer is referred to as 'confined'. A borehole made in a confined aquifer therefore presents a piezometric level higher than that of the aquifer's roof. If the piezometric level is higher than the topographic level then the borehole becomes artesian (Figure 3.13A);

– an aquifer of which the upper section is occupied by a non-saturated area is called a 'unconfined aquifer'. In this case, the static level of a borehole coincides with the water level in the aquifer (Figure 3.13B). A remarkable difference in the behaviour of these aquifers occurs in the course of a pumping process. The productivity of a unconfined aquifer decreases because the transmissivity (T = K x e) drops as e reduces. The transmissivity of a confined aquifer does not change during pumping, as long as the piezometric level remains higher than the roof. Aquifers that change from a free to a confined state, depending on topography, can sometimes occur.

The lateral limits of an aquifer can be geological (fault, lateral change of facies etc.) and/or hydrodynamic and thus variable in time and space. The limits in flow match incoming or outgoing flow rates (supply by a lake, drainage by a spring line etc.); the potential limits act on the aquifer by imposing a head level (level of a water surface, stream etc.). In some cases, these limits, usually called boundaries, will be mixed.

Table 3.VIII: Principal hydrodynamic parameters of aquifers.

	Parameter	Symbol	Usual unit	Notes
Storage	Total porosity	n	%	$n = V_{void}/V_{total}$
	Kinematic porosity (or effective porosity)	ω_c	%	ω_c = hydrokinetic speed/Darcy flow rate
	Specific yield (or drainage porosity)	S_y ou n_e	%	$S_y = V_{water\ grav} / V_{total}$
	Characteristic diameter	d_{10}	mm	10% of the smallest sample
	Specific surface	S_s	cm^2/cm^3	$S_s = S_{grain}/V_{total}$
	Storage coefficient	S	non dimensional	measure in situ (see Chapter 6)
	Thickness	e	m	
	Reserve	W	m^3	W = S x e
Flow	Hydraulic conductivity	K	m/s	measure *in situ* (see Chapter 6)
	Intrinsic permeability	k	m^2 or Darcy (D)	for water at 20°C :
			1 D = 0.98 10^{-12} m^2	k (D)= 0.966 10^{-5} K
	Transmissivity	T	m^2/s	T = K . e
	Hydraulic head	H	m of water column	
	Static water level	SWL	m	measured by reference to a datum
	Hydraulic gradient	i	non-dimensional	i = (H2 – H1)/L
	Unitary flow rate or filtration velocity	q	m/s	q = K . i
	Darcy flow rate	Q	m^3/s	Q = K . A . i
	Effective velocity	V_e	m/s	$Ve = q/S_y$

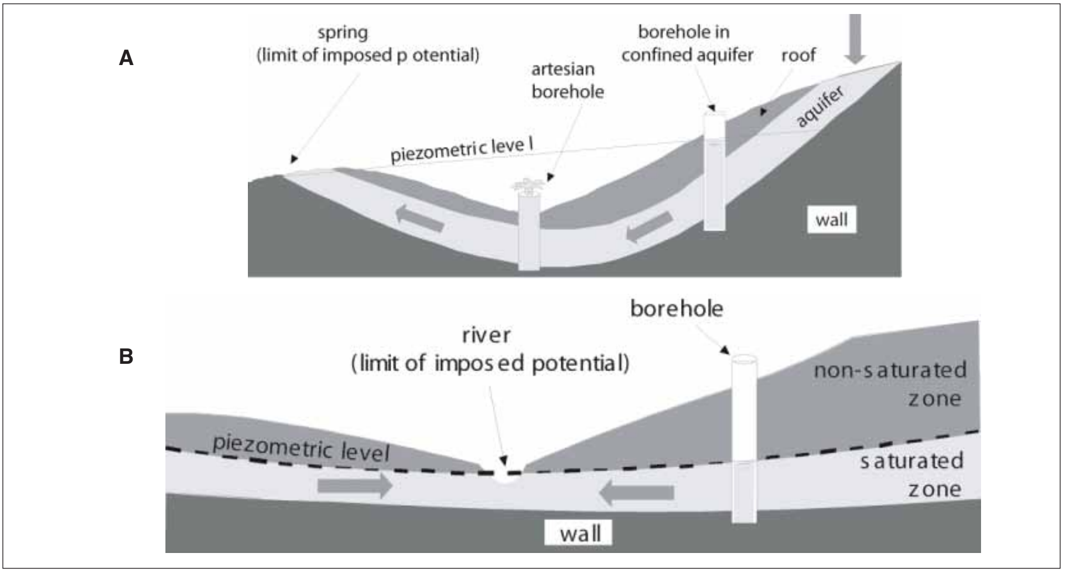


Figure 3.13: Aquifers.
A: confined. B: unconfined.

4.2.1 CRYSTALLINE BEDROCK AQUIFERS

Bedrock geological formations are very widespread and have been subject to major exploitation for about twenty years. These plutonic and metamorphic rocks are characterised by their compact nature, and by their very low porosity and permeability. Nevertheless, permeable horizons have sometimes been able to develop due to processes of physical-chemical weathering, and some phenomena of tectonic origin have allowed these rocks to acquire a secondary permeability that favours the formation of aquifers. Certain old sedimentary rocks resemble igneous and metamorphic rocks in their hydrogeological behaviour, and are therefore usually regarded as bedrocks also.

Geological formations

Three types of formation, each with its own characteristics, can be distinguished: granites, gneiss and migmatites; schists and greenstones; and quartzite sandstones. For simplicity, the terms granites, schists and sandstones will be used in this section, always bearing in mind the idea that these terms represent groups of rocks with usually similar hydrogeological behaviour.

Granites are compact, relatively inelastic rocks. Usually fractured, they present a network of fissures and open or closed fractures depending on orientation. Beyond about 50-70 m in depth, it is generally assumed that the majority of the fissures are closed due to the weight of the ground. Weathering phenomena that affect granites are essentially the result of the actions of water and temperature. Considerable variations in the nature and thickness of these weathered rocks are observed, depending on climatic areas and drainage conditions. Granites weather to a more or less argillaceous sand, with a thickness that can reach several tens of metres. At the base of these residues of weathering and at the contact layer with the solid rock, a coarse, sandy gravel is usually present (Figure 3.14).

Schists are deformed more easily than granites and thus do not always contain such a well-developed fracture network (Figure 3.15). They can nevertheless exhibit some fracturing linked to their structure. The weathering of these rocks can be very deep, but it is usually very argillaceous.

Quartzite sandstones are generally very consolidated and may have suffered significant fracturing. Usually less weathered than the other rocks, they sometimes have intercalations of a different nature (carbonates, pelites etc.) which can provide drainage.

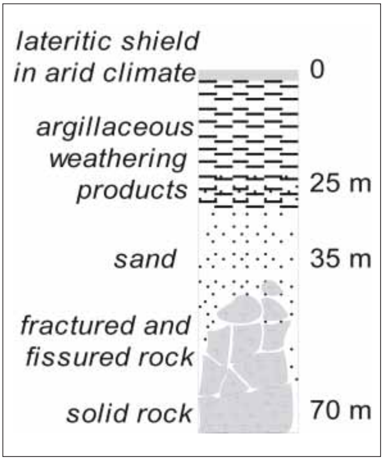


Figure 3.14: Weathering profile of granites.

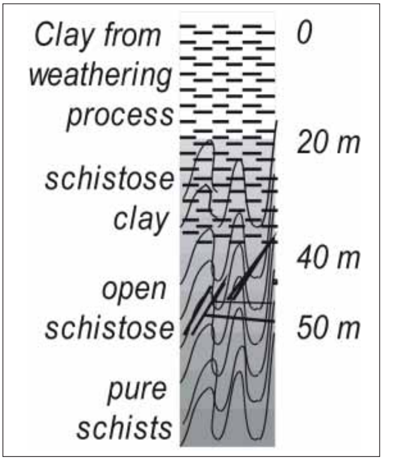


Figure 3.15: Weathering profile of schists.

Aquifers

There are three basic types of reservoir:

– *reservoirs composed of weathering products,* such as sandy-argillaceous gravels, of considerable extent in the areas of African basement zones. Their average thickness (10-20 m in granite-gneiss and 15-40 m in West African schists) is greater in equatorial zones and decreases towards the tropics;

– *fissure reservoirs situated just beneath the solid rock*. These are partially-weathered areas, made up of numerous fissures and joints, generally filled with weathering products. These reservoirs can extend over several tens of metres;

– *networks of faults or major fractures,* which can be the basis of groundwater movement.

Numerous studies carried out in West Africa (Lachassagne *et al.,* 2001, Wright & Burgees, 1992, University of Avignon 1990) have shown that these various reservoirs generally constitute a single aquifer, the storage function of which is ensured by the weathering products, and the flow function by the fissured or fractured area. The static level is generally situated in the weathering products. This group may in turn be drained by major fractures, an essentially conductive role. On the other hand, isolated aquifers of weathering products, fissures or faults, can also be found locally.

A statistical summary of the data from West Africa shows that the rate of success and productivity of boreholes increases with the thickness of weathering; and an empirical relationship between specific flow rate and the weathering thickness W_t can sometimes be found in a particular region. There is however a critical thickness threshold at about 35 m, beyond which productivity tends to decrease.

Much the same is true of the fissured and weathered areas: from about 30 m thickness, productivity does not seem to increase; beyond 50 m it becomes random. In all probability, this thickness is greater in the case of sandstone massifs. Figure 3.16 shows the types of instantaneous flow rates (slightly overvalued) obtained depending on the type of aquifer, in West African rural boreholes.

The productivity and the rate of success of a borehole depends equally on the quality of the weathering products and on the fissures: clay content, density and openness or blockage of fissures etc. The University of Avignon (1990) suggests transmissivities of 10^{-4} to 10^{-2} m^2/s and storage coefficients of 2×10^{-5} to 5×10^{-5} obtained in granite-gneiss, from pumping tests carried out in Ghana and Bénin. Wright (1992) gives an average transmissivity obtained by pumping test, in the granites and gneiss of Zimbabwe, of 5×10^{-5} m^2 and hydraulic conductivities varying from 1.15×10^{-7} to 3×10^{-5} m/s (data from more than 500 boreholes).

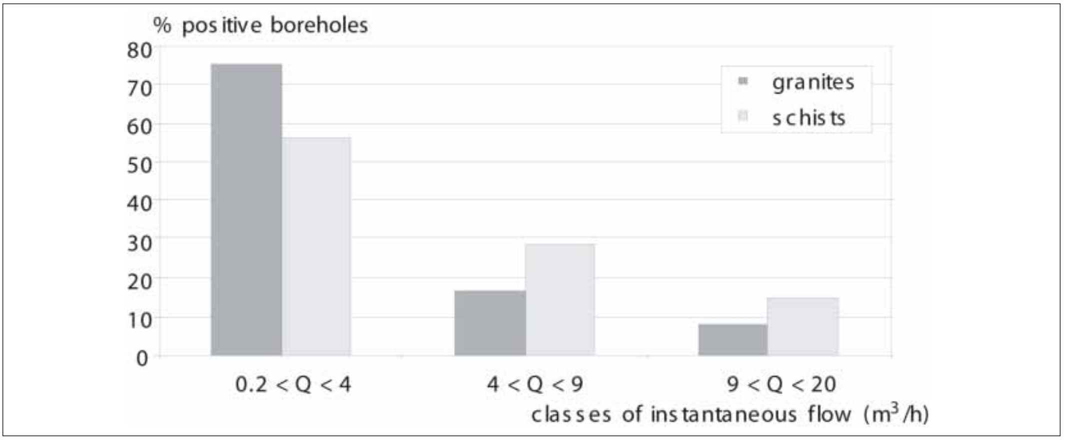


Figure 3.16: Frequency of instantaneous flows in basement rock aquifers (BRGM 1992).

Available resources

The storage capacity of a granite-gneiss zone can be calculated from its storativity. Taking a specific yield (or specific drainage) of 5% for the saturated weathering products and fissured zone with a 15 m thickness, and a specific yield varying from 0.1% at 15 m to 0% at 50 m for the bedrock, the distribution of potential reserves shown in Figure 3.17 is obtained.

Even though this distribution is notional, it reflects the storage role of the weathering products and the flow role of the fissured area. This reserve corresponds to an available volume of water: however, it also has to be supplied. For the weathering products reservoir, the direct supply threshold is usually taken to be 400 mm in the sahelian area, 600 to 800 mm in the savannahs and 1 000 to 1 200 mm in coastal forest areas.

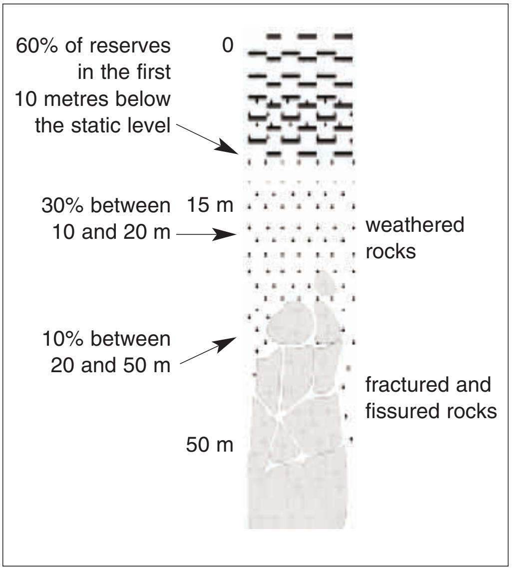


Figure 3.17 : Reserves in the bedrock zone.

Table 3.IX: Recharging in a bedrock area.

Climate	Rain (mm/year)	Exploitable recharge (mm/year)	Exploitable volume (m^3/km^2/year)	Exploitable volume (m^3/km^2/day)
Sahelian	500	50	50 000	130
Southern-sahelian	800	65	65 000	180
Sudanese	> 1 000	160	80 000	200

Box 3.5
Chotts and sebkhas.

Chotts and sebkhas are endoreic areas, true 'evaporating machines' that can be found in a sahelian area. When the piezometric level of the unconfined aquifer is located near ground level in a very arid climate, the water is taken up by evaporation. The mineral concentration dissolved in the groundwater is thus enriched up to the point of supersaturation and precipitation of those minerals. Depending on the initial mineralisation of the water, various types of development and various final conditions can be observed.

This type of phenomenon can be found either in continental or in littoral areas (supersaline lagoons).

A study carried out by BRGM from 1984 to 1991 allows orders of magnitude to be established for supply and exploitable resources in areas of schists and granites (Table 3.IX, Burkina Faso).

4.2.2 UNCONSOLIDATED-ROCK AQUIFERS

Valley aquifers

The effective rain that infiltrates into large areas of permeable ground generally supplies unconfined aquifers. These aquifers are sometimes called groundwater tables, because they are the first to be found when sinking a well.

Their outlets are the low points of the topography: springs, streams or the sea. In a sahelian climate, direct evaporation of groundwater is possible if the piezometric levels are near the ground (Box 3.5).

The boundary conditions vary, but usually the relation between the stream and the aquifer is essential: sometimes it represents a limit to the incoming or outgoing flow.

In arid areas, the flow is concentrated in the stream (or the lowlands), which therefore supply the aquifer; these usually temporary streams are called *wadis* or *oueds* (Figures 3.18A and 3.18C). In a temperate climate, the stream usually drains the aquifer (Figure 3.18B); nevertheless, for the same stream, periods of supply (wet season, high water) and drainage of the aquifer (dry season, low water) usually occur. In the same way, a stream can change from one situation to another through all or part of its course. Finally, a stream may not have any relation with the aquifer (Figure 3.18D); this is generally the case when the bed of the watercourse is blocked.

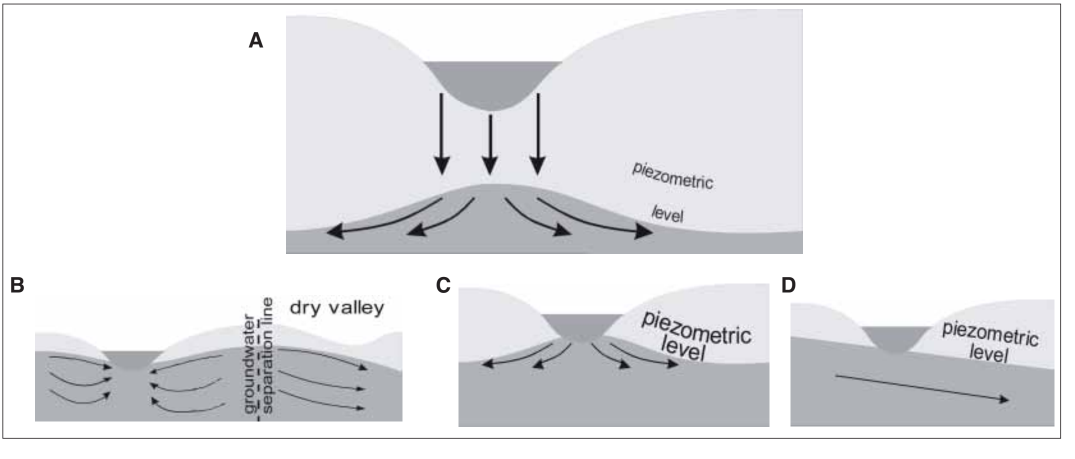


Figure 3.18: Relations between rivers and aquifers.
A: perched river, supplying aquifer. B: river draining aquifer. C: river supplying aquifer.
D: independent river, bed blocked.

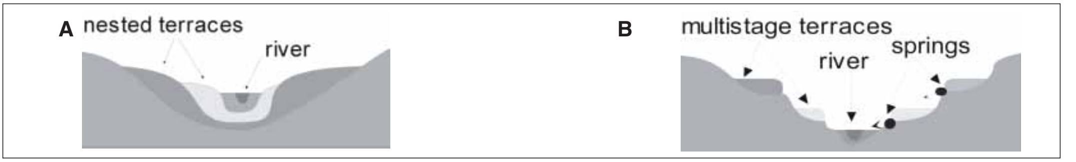


Figure 3.19: Alluvial terraces.
A: nested terraces. B: multi-stage terraces.

Alluvial aquifers

These aquifers are located in the alluvial deposits of a watercourse. They are in equilibrium with the stream water, which drains and feeds the aquifer depending on the seasonal high and low water conditions.

A particular form of fluvial deposit is represented by alluvial terraces. The formation of these terraces is due to the alternation of periods of erosion and deposition induced by climatic variations (glacial periods, rainfall) and by tectonic movements (uplift, subsidence). Depending on the conjunction of these various factors, two types of terraces occur. Nested terraces constitute a single but heterogeneous aquifer (Figure 3.19A), while multistage terraces constitute independent aquifers, sometimes underlined by lines of springs in contact with the substratum (Figure 3.19B).

When the alluvial deposits have not been recently disturbed, the coarsest sediments are usually found at the bottom of each terrace (positive granular classification). The thickness of these alluvial deposits can be very great. In general, the permeability of the coarsest materials is very good and determines a type of aquifer with a strong flow function but a moderate storage function. The rate of renewal of water is generally high, but the quality of the water must be checked when the aquifer is fed by a stream.

This type of aquifer is the most obvious, since it is indicated by the presence of the stream and associated vegetation. It is generally exploited in a traditional manner, but it is nevertheless necessary to pay attention to the stream/aquifer relationship in order to avoid an unpleasant surprise: the exploratory boreholes sunk by ACF in the alluvial deposits of the Juba River in Awdegle (Somalia 1993) turned out to be dry, because the stream bed was blocked and the alluvial aquifer was absent in this part of the region.

Perched aquifers

These aquifers are mainly formed in sedimentary deposits when a low permeability layer (often clayey) in the unsaturated zone creates a small aquifer situated above the main reservoir (Figure 3.20). The extent of these aquifers can be limited and reserves poor. They can also be perennial or seasonal. When sinking wells, it is important not to mistake a perched aquifer for the unconfined aquifer that is being sought.

One example is the perched aquifers of the sandstones of the Bateke plateaux of the Congo basin (1998 ACF exploratory mission). Created by horizons of impermeable siliceous rocks, these aquifers have, in some places, piezometric levels of several tens of metres higher than that of the deep aquifer of the sandstones.

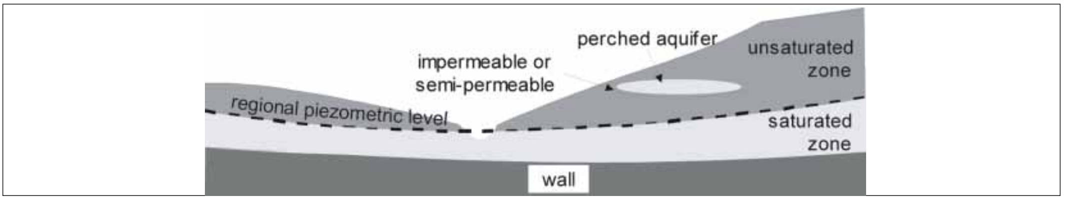


Figure 3.20: Perched aquifer.

Coastal sandbars

Sandbars have developed in numerous places along the African coast. The thickness of these sands can be sufficient to allow the creation of a freshwater aquifer, a sort of lens in contact with the salt water of the ocean on one side, and the supersaline water of the lagoons on the other. The geometry of these aquifers depends greatly on abstraction and supply conditions, but they are vulnerable because they are subject to saline intrusion.

In 1994 ACF sank a 12-m well on a beach of this type in Monrovia. The water remained fresh at an abstraction rate of 1 m^3/h (the annual rainfall is about 5 m).

4.2.3 AQUIFERS IN MAJOR SEDIMENTARY BASINS

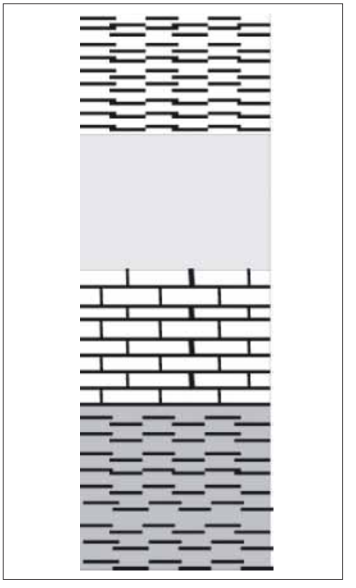


Figure 3.21: Stratified profile of a sedimentary basin.

These reservoirs are made up of a succession of more or less heterogeneous strata of various types corresponding to successive sedimentation episodes (Figure 3.21). All the formations can contain aquifers, but their function, quality and exploitation potential vary depending on depth.

There are three main types of system:

– *Unconfined aquifers* generally acts like water table aquifers or alluvial aquifers. Their supply area is proportional to their extent, which can be quite considerable, producing an aquifer with a large storage capacity.

– *The multi-layer aquifer* is formed by a group of permeable levels separated by less permeable or impermeable levels. The leaking phenomena, which allow a relation between two aquiferous levels divided by some semi-permeable material, can play an important role and thus enable the system to be considered as unified but multi-layered. The supply area of this group is generally limited to outcrop areas at the edge of the basin.

– *The deep confined aquifer* is usually poorly supplied, but can offer considerable storage capacity. The quality of the water does not always make it suitable for consumption, because of excessive mineralisation.

The limiting conditions, apart from the cases already mentioned, are sometimes created by fresh-water/salt-water contact. In fact, the flow of these aquifers moves towards the low points, oceans or seas, whereas salt water tends to infiltrate permeable terrain. If salt water is assumed to be immobile, and the fresh-water/salt-water interface without a significant mixing area, the simplified 'de Ghyben-Herzberg' equation can be developed:

$$H = 40\ h$$

where h is the elevation of the piezometric level above sea level and H is the depth of the fresh-water/salt-water interface below sea level. This simplified relation allows the profile of the saline wedge to be calculated for constant slope (Figure 3.22). Therefore, a drop of 1m in the piezometric level (pumping, for example) induces a rise in the saline wedge of 40 m.

In fact, the actual phenomena are more complex, and the fresh-water/salt-water interface can be very large, especially when the sea level changes: this was found by ACF in the coastal plain of Mangdaw (Myanmar, 1993-1998), where the high tidal range creates a mixing area which is all the greater because of the rise of the salt water due to rivers over several tens of kilometres. Only the upper section of the aquifer in that region can be used, because beyond a certain depth (20 to 30 m) it becomes saline.

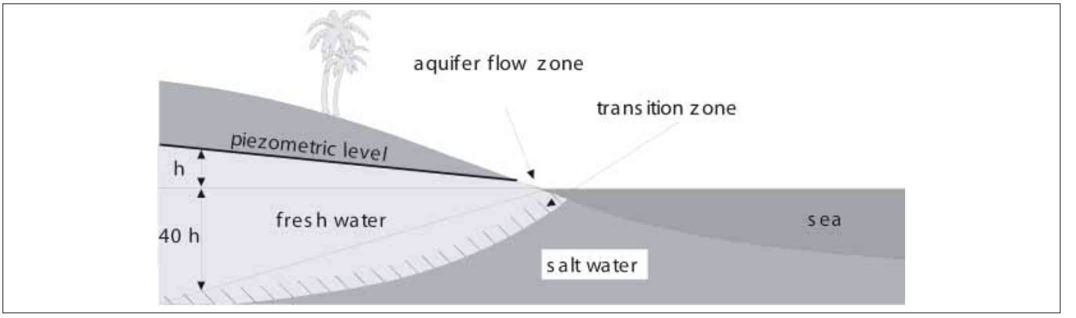


Figure 3.22: Saline wedge.

4.2.4 HIGHLY HETEROGENEOUS AQUIFERS

Karst and volcanic environments are characterised by their great heterogeneity, both within and between different aquifer systems.

Karst

There are numerous definitions and various concepts of karst. It is nevertheless possible to explain some important characteristics.

The reservoir consists of carbonated rock, which can have some permeability, even if in general it is not very significant. What gives karst its originality is its macro-permeability, due to the dissolution of the rock by movement of water through preferential passages (diaclases, fissures and fractures).

Generally, three areas can be distinguished in a karstic massif (Figure 3.23):

– *An outcrop area* (in general with little cover), sometimes called epikarst, where the rock is loose (open fissures), which can sometimes contain a small surface aquifer. Surface flow is almost absent in the epikarst because infiltration is significant. The biological activity developed charges the water with CO_2, which causes accelerated dissolving of the carbonated rock.

– *An unsaturated area,* sometimes very thick, which allows infiltration of the water coming from the epikarst. This infiltration occurs rapidly through fairly large channels, but it also occurs more slowly through fissures, diaclases or possibly interstitial porosity. This area has a flow function, but sometimes it also has a storage function when it is thick enough, and/or when large-capacity annexed systems have developed there (cavities).

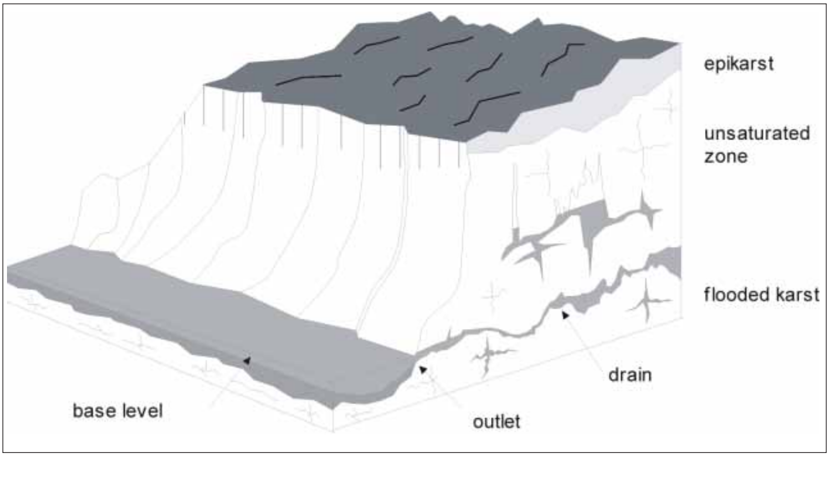


Figure 3.23: Karst.

– *A saturated area,* known as 'flooded karst', which has outlets of all types, though some are truly spectacular. They are karstic springs, with variable behaviour and sometimes with a very significant flow rate.

The karst reserves are generally important, even if their assessment is difficult. The permeability and porosity of such a strongly anisotropic environment are difficult to calculate, and the results of test pumping cannot be exploited with normal tools (see Chapter 6). The study of these environments is carried out with a whole range of tools, generally based on the study of the system's behaviour.

The management of karsts is difficult from a quantitative point of view, but even more difficult from a qualitative point of view. Indeed, very fast flows transfer pollution easily, in directions that are sometimes difficult to predict: the protection of the aquifer is therefore difficult.

Volcanic environment

Volcanic materials are very varied, though basically they can be divided into flows, extrusions, and intrusions. Lava flows can cover very large areas and have a thickness of several hundred metres. Sometimes compact, these flows are highly permeable if they are fissured; their weathering products are generally argillaceous. The creation of flows by 'carpet rolling' produces a certain stratification of the environment. In the same way, flows can be stratified with projection materials (ash or tufa), the micro-permeability of which is generally quite good. There are also lava tunnels created at the time of the differential cooling of the flow, which then constitute areas of great permeability. In general, the materials of volcanic flows are more favourable to flow than to storage of water.

The seams and dykes of volcanic materials (usually of a doleritic nature) are sometimes fractured and weathered. Injected into fractures, they then provide drainage. Conversely, if they are not weathered, they can form underground barrages to groundwater flow.

4.3 Aquifer recharge

The assessment of the characteristics of the aquifer is carried out at a given moment, and it is unwise to consider that they will remain stable over the whole year, or from one year to the next. The best example is, without any doubt, the thickness of an unconfined aquifer affected by its supply: there are many areas where productivity has fallen after an annual or inter-annual piezometric fluctuation. An example is the boreholes of South Angola (Kunene), where productivity has become almost zero following a general decrease in piezometric levels (ACF Angola 1998).

It is also necessary to consider groundwater as a renewable resource: in the aquifer, water pumped from wells and boreholes disturbs the water cycle's natural balance. It is therefore advisable to estimate the impact of withdrawals on the system and to make sure they do not dry up the resource. There are numerous examples of boreholes with a decrease in productivity after several months of use; as the pumped flow rate is higher than the supply rate, the system reserves are progressively used up. One example is a borehole sunk by ACF in 1996 in Sudan, where the flow rate of 5 m^3/h at the end of the development period progressively decreased, to become almost non-existent two months later.

In order to calculate the recharge rate of an aquifer, its hydrological balance must be drawn up. The volume of water that goes into the aquifer is assumed to be equal to the volume of water that comes out, plus or minus changes in the stock. The balance is expressed as follows:

$$\text{input} = \text{output} +/- \text{change of stock}$$

In short-term humanitarian programmes, it is not possible to determine complex hydrological balances, since this is a considerable task requiring a long period of observation. Furthermore, withdrawal flow rates are generally quite low. Nevertheless, in programmes spread over several years, it is possible to estimate this balance by using the available data, or by directly measuring simple parameters in the field. These investigations do not allow definite conclusions to be drawn, but some recommendations may be made (see the example in Chapter 5B, Section 4).

The measurement of recharge is carried out within a known system: its geometry and hydraulic mechanism must be known for the preliminary investigations (geology, boreholes, geophysics etc.). The working scale is the hydrogeological basin. The minimum scale of observation is one year, with a monthly or 10-day measurement interval. However, measurements carried over in a single year do not allow perennial regulation mechanisms to be estimated, and therefore the results have only an indicative value.

4.3.1 PIEZOMETRIC MONITORING

The simplest method of assessing the effects of exploitation of an aquifer is to carry out piezometric monitoring. Carried out monthly over a period of several years, piezometry tracks the behaviour of the aquifer, and is possible in the majority of programmes. The boreholes and wells must be chosen according to their location and the system under consideration. In practice, it is possible to start piezometric monitoring on a large number of wells in order to be able to draw a map as close as possible to reality. To alleviate logistics constraints, only the analysis of wells regarded as representative is then followed up.

4.3.2 SIMPLIFIED WATER BALANCE

Bearing in mind the constraints linked to humanitarian interventions, it is only possible to calculate the hydrological balance of relatively simple systems, where the inputs and outputs are easily quantifiable. Such systems present a balance similar to that shown in Table 3.X.

Two methods are used, and the balance is checked by comparing the two results. The upstream approach consists of calculating recharge by estimating the inputs to the system. The downstream approach estimates recharge from measurement of the outputs from the aquifer. The estimation can only be regarded as satisfactory if the two methods give similar results.

Table 3.X: Simplified balance.

Input	=	Output	+/–	Stock change
Useful rainfall (UR) + surface water infiltration (Is)	=	Flow rate of the springs + base flow rate of the streams (Qs) + pumping (Qp)	+/–	Stock change (Δs)
(UR + Is)	=	(Qs + Qp)	+/–	(Δs)

Moreover, the balance thus calculated can be verified by the hydrogeological approach: the underground flow rate calculated by Darcy's equation (Q = KAi) must correspond to the infiltration flow rate calculated by the balance method (+/– the changes in the stock).

Steps in the implementation of these methods are presented in Annex 6, and an example is given in Figure 3.24.

4.4 Groundwater quality

Groundwater has a reputation for being of good quality for human consumption. Biological, and especially microbiological contamination risk of this resource is extremely limited. However, the presence of mineral toxicity (natural pollution such as arsenic or fluorine) may occur, potentially leading to chronic health problems. When this type of natural pollution risk exists, rain or surface water become an alternative resource. Groundwater is generally less sensitive to various types of pollution than surface water (except in fractured contexts and karsts). The different types and interpretations of water analysis as well as measures for the protection of groundwater are presented in Chapters 4 and Annex 10.

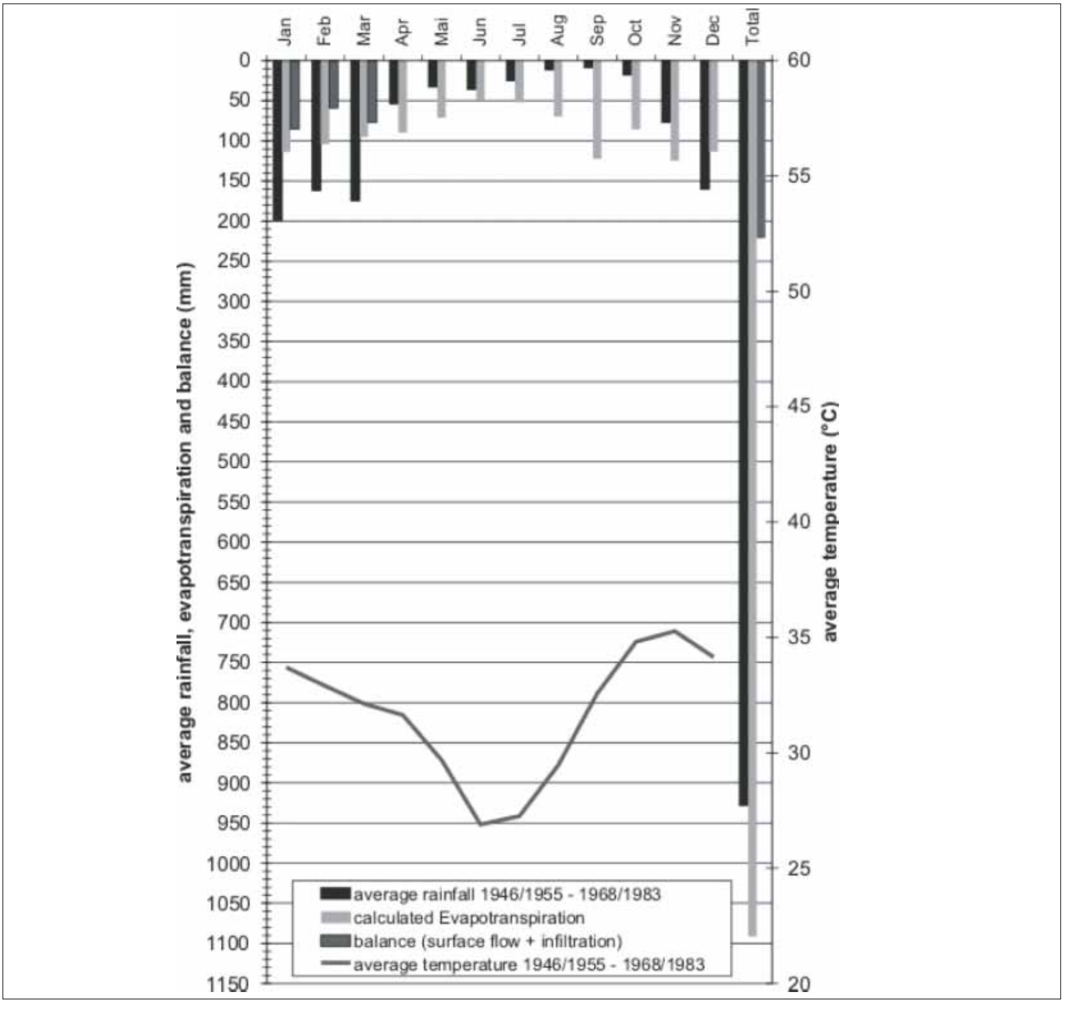


Figure 3.24: Water balance of Caia district prepared by ACF (Mozambique, 2000) with local data and using the upstream approach.
The calculation is based on the Thornthwaite method (see Annex 6) with a monthly calculation step.

4.4.1 HEALTH ASPECTS

The essential elements (major ions), as well as certain trace elements (iron, zinc, copper, manganese etc.), are considered important for human health. Groundwater can supply the essential daily intake of some of these minerals, but a varied and balanced diet is the best way to ensure sufficient intake.

Toxic elements may lead to acute health hazards if ingested at high concentrations, or may lead to chronic diseases if consumed at low levels over an extended period. Drinking water quality guidelines and regulations (e.g. WHO guidelines and national standards) have been developed on the basis of known or supposed risks to health.

Except in the case of agricultural contamination (pesticides, fertilisers or manure), chemical pollution of groundwater is mainly of natural origin, induced by the interactions between the rock and the water (leaching and dissolving). These interactions can create acceptance problems (e.g. in case of excessive conductivity or high iron presence), that can put people off a water source which is in fact potable from a sanitary point of view. It may also present a health risk (e.g. chronic diseases linked to mineral toxicity). Nevertheless, the biggest problem remains microbiological pollution of water.

4.4.2 CHEMICAL SIGNATURES

An analysis of the major chemical elements present in water enable it to be characterised. By comparing various analyses, it is possible to identify the chemical 'signatures' that correspond to different aquifer systems. It is thereby possible to differentiate various types of water as, for example, has been shown in numerous studies carried out in West Africa on bedrock groundwater.

Table 3.XI shows the average values of selected parameters for various types of aquifer in West Africa and Table 3.XII summarises the results of water-quality monitoring carried out by ACF in Uganda.

Table 3.XI: Quality of bedrock groundwater (from University of Avignon 1990).

Quality	Aquifer	pH	Conductivity (μS/cm)	Mineralisation (mg/l)
Medium	Granitic gravel	7.2	125	122
	Schist gravel	7.6	172	165
Inferior	Fissured schists	7.9	404	326
	Fissured granites	6.6	380	386
	Old sandstones	6.9	190	251

Table 3.XII: Quality of bedrock groundwater.
Annual average values 1996-1998 (ACF Uganda).

Water source	Quality	Aquifer	Conductivity (μS/cm)	Temperature (°C)
Boreholes (40)	Medium/inferior	Granite-gneiss sands	180-400	24-26
Springs (5)	Superior	Weathering products	100-130	24-27

The analysis of the intrinsic isotopes of the water (deuterium ^{2}H and oxygen ^{18}O) as well as the stable isotope ^{13}C enables the origin of the groundwater to be traced. Analysis of the radioactive isotopes (^{3}H and ^{14}C) gives the age of the groundwater (time of transit).

Humanitarian programmes are sometimes confronted with complex water-quality problems. Chemical analysis (major ions, trace elements, toxic minerals) is a necessary tool for understanding a hydrogeological context. The assistance of a laboratory is sometimes indispensable.

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