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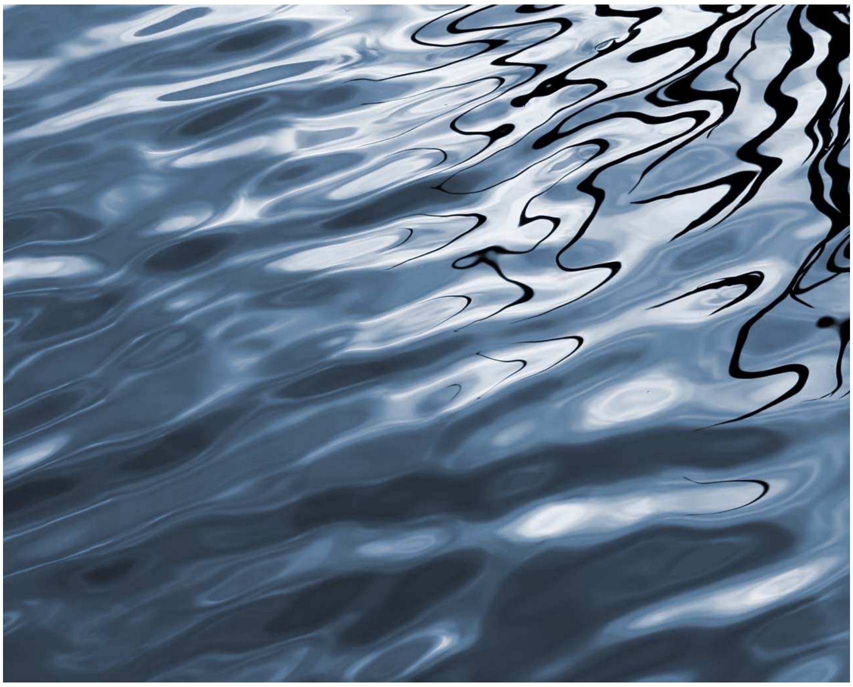
Measuring for water



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Contents

Chapter 1

Introduction

1.1 Need for measurements

There are several reasons why measurements for water are necessary, which in general can be grouped into four purposes:

- planning (how many ditches are required for a new neighborhood?)
- design (how high should a dike be? What is the minimum pipe diameter?)
- management (flood protection, maintaining water levels for agriculture)
- research (how is water transported from the land surface to the river? How are the water flows around structures?)

These objectives are leading in the way the measurements are carried out, because it determines what kind of data is required in terms of sampling frequency, sampling period, spatial distribution and quality (precision). For example, 'planning' requires long term records with usually a large time scale, 'design' requires also long term records, but on a smaller time scale. For the purpose of 'management' real-time data is required, so there can be anticipated on the consequences with forecast models. And for 'research' often high quality data with a high sampling frequency is essential.

Hence, it is important to clearly define what the measuring objective (or research question) is and what kind of data is required to meet this objective. Unfortunately, there are also limitations to the 'wish list' of measuring requirements. Often the most dominant limitation is the budget. Like often the case: the more advanced the equipment, the higher the price. Also labour is an important aspect. Some sensors are fully manual, which is really labour intensive, others store their data on a data logger, which only need to be read out on a regular basis (e.g. every month, or half year depending on the memory size), and other sensors fully automatically send their data wireless to the office, which is less labour intensive. Although the latter seems to be most desirable (even if we neglect the current high costs of these wireless systems), one should be careful to fully trust the remote data. It remains wise to regularly visit your measuring site to verify weather or not your sensor is working properly. For example, a rain gauge can be overgrown by a tree. If you never visit your rain gauge, it is very difficult to notice this from the measured data only.

Another consideration for choosing a measuring technique is the accessibility of the site. A difficult accessible or remote site requires more automatic sensors than for example laboratory experiments. Again the trade off between sensor price and labour costs are of importance for the sensor technique choice.

1.2 Measuring errors

Unfortunately, it is not always the case that when we measure, we measure the truth. When doing measurements errors are always involved. Without going into all the details of errors, it is important to realize that errors of observations are usually grouped as random (or stochastic), systematic and spurious.

Random Errors Random errors are sometimes referred to as experimental errors and the observations deviate from the mean in accordance with the laws of chance such that the distribution usually approaches a normal distribution. Repeating the measurements or extending the period of observation may reduce random errors. This will balance out the random effect and bring the result closer to an average value.

Systematic errors Systematic errors are those which cannot be reduced by increasing the number of observations, if the instruments and equipment remain unchanged. In streamflow, the systematic errors may be present in the water level recorder, in the reference gauge or datum, and in the current meter. It is possible that the crest of a weir is leveled incorrectly to the station datum, so producing a systematic error in the head measurement, which might have a serious effect on low values of discharge.

Spurious errors These are human errors or instrument malfunctions and cannot be statistically analyzed. The observations are recognized as outliers and must be discarded.

1.3 Reading instructions

These lecture notes give an overview of measurements in the field of water science and practice. It covers the fields of Water Resources, Sanitary Engineering, and Hydraulic Engineering. In Chapter 2 the basic measurement techniques of the most important hydrological processes will be described. The next chapter will focus on hydraulics and Chapter 4 will explain the measurements that take place in urban drainage and drinking water. The last chapter elaborates on filter techniques, sampling frequency and control techniques.

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Chapter 2

The hydrological cycle and the water balance

The hydrological cycle is the most fundamental concept for any water engineer. The hydrological cycle, which is depicted in Figure 2.1 shows how water finds it way from the sea to the atmosphere, to the land, to the river, and back into the sea. Depending on the scope of the engineer the focus is on a specific part of the hydrological cycle. For example, someone who is interested in polder behaviour, needs specific information on rainfall, evaporation, seepage and in- and outflow discharge of the polder area. But he is not interested in how the discharged polder water is transported to the sea. On the contrary, a hydraulic engineer, does not focus on the polder system. He especially wants to know how river water is transported to the sea. His focus is more on sedimentation processes and fluid dynamics. Hence which processes are important, the matter of detail, and where to put the boundaries depend on the objective of the study.

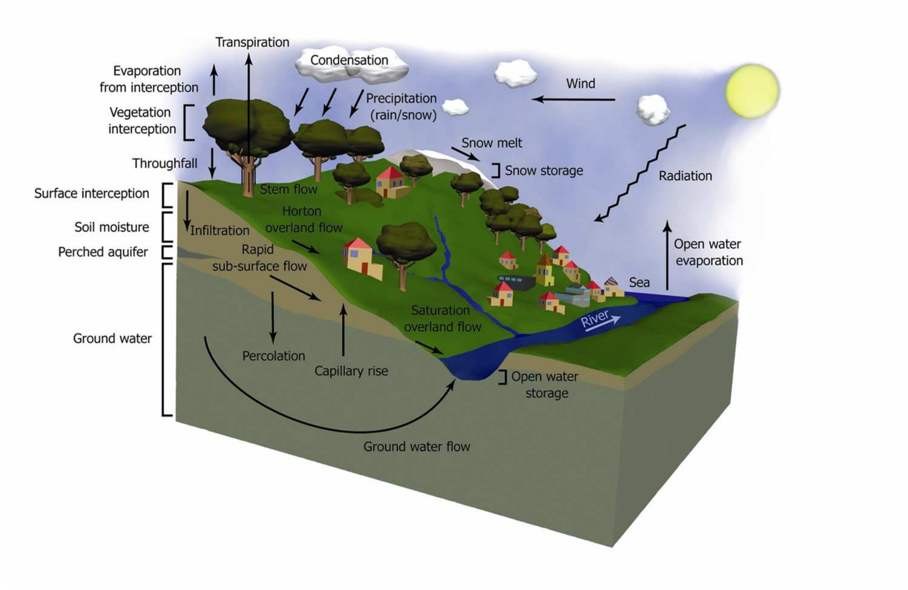

Figure 2.1: The hydrological cycle.

In this chapter the main processes of the hydrological cycle will be discussed, since these processes are generally important for all water disciplines. The basics behind these hydrological processes will be explained, but the focus will be on how these processes can be measured. The processes that will be discussed are:

- Rainfall, P
- Evaporation, E_a
- Discharge, Q
- Soil water storage (unsaturated zone)
- Ground water flow and storage

Rainfall, evaporation, discharge, and ground water flow are fluxes and have the dimension [L/T], while soil water storage and ground water storage are stocks with dimension [L]. For a closed system (e.g., a catchment), we can summarize the incoming and outgoing fluxes into the so-called water balance equation:

$$P - E_a - Q = \frac{\mathrm{d}S}{\mathrm{d}t} \tag{2.1}$$

In the storage term S both the soil moisture storage and the ground water storage are included. Of course, within the storage reservoir also internal ground water flow occurs, but this depends again on where one puts the boundary conditions.

2.1 Rainfall

From all the components of the hydrological cycle, the elements of precipitation, particularly rain and snow, are the most commonly measured. It would appear to be a straightforward procedure to catch rain as it falls and measure the depth of snow lying. However, climatologists and water engineers appreciate that making an acceptable precipitation measurement is not as easy as it may appear. It is not physically possible to catch all the rainfall or snowfall over a drainage basin.

The traditionally way is to sample the precipitation over the area by rain gauges (Section 2.1.1). The measurements are made at several selected points representative of the area and values of total volume (m^3) or equivalent areal depth (mm) over the catchment are calculated later. The measurements have to be done proper and standardized and therefore rules have evolved on equipment and placement of gauges.

Nowadays, more and more radar and remote sensing products become available which provide spatial patterns of rainfall. Nevertheless these products rely on verification on the ground. These products and why ground verification is necessary will be explained in Section 2.1.2.

2.1.1 Point observations of rainfall

Funnel

The most common way to capture and collect precipitation is by means of a funnel. The surface area at the orifice is the catching surface area. Obviously a large surface area reduces the measurement error. The opening surface area is usually standardized to an acceptable level of error. A common size for the aperture is 200 cm^2. The rims of the funnel are sharp to avoid turbulence, and the drop in the funnel is such as to minimize losses from out splash in heavy rain.

Tipping bucket rain recorder

Nowadays most used are tipping buckets (Figure 2.1.1). Rain is led down a funnel into a wedge-shaped bucket of fixed capacity. When full, the bucket tips to empty and a twin adjoining bucket begins to fill. At each tip, a magnet attached to the connecting pivot closes circuit and the ensuing pulse is recorded on a counter or electronically. This makes the tippet bucket suitable for electronic logging and/or telemetry. The capacity of the bucket is designed as to represent e.g., 0.2 mm, 0.5 mm or 1 mm of rainfall depth. What is registered electronically is the time of tipping, knowing that each tip represents accumulated rain equal to the capacity of the bucket since the last tipping. It has to be noticed that the tipping bucket does not properly register (low intensity) rainfall less than the bucket capacity and that there is a limit to high intensity rain that it can register accurately. It is also advised to arrange for water to be collected below the buckets for verification, so that totals can be measured if the recording fails.

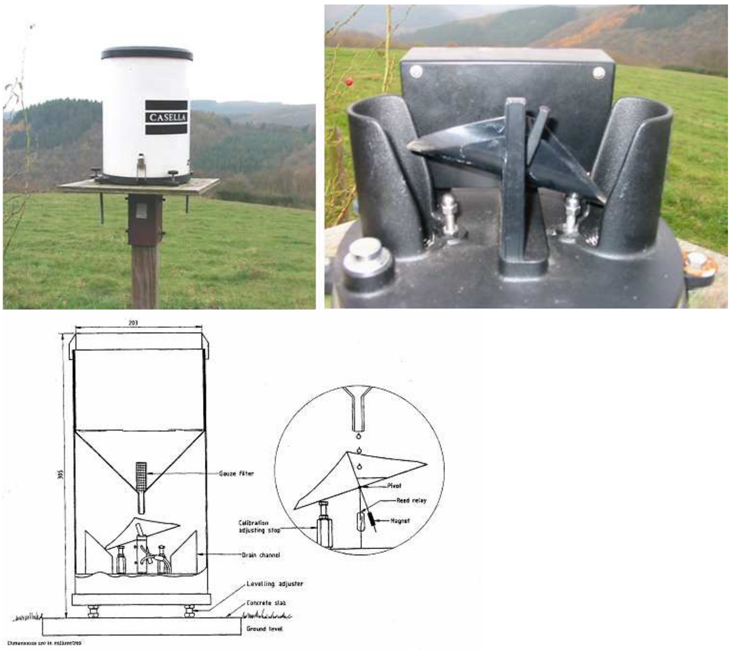

Figure 2.2: Tipping Bucket

The height of the aperture of the rainfall recorder above the ground surface has a significant effect on the catch of actual rainfall. Wind around the measuring device causes turbulence and hence lowers the catch. Generally, wind speeds increase with height above the ground. As a result, the catch reduces with increasing heights. At 1.50 m above the ground surface the measured rainfall can be 84%-96% from the actual rainfall at ground level.

Disdrometers

Disdrometers measure rain by counting drops. There are various underlying principles for disdrometers: video disdrometers, laser disdrometers, acoustic disdrometers, to name a few. Compared to other type of rain gauges, disdrometers have the advantage that, in addition to rain rates, they also measure drop size from which drop size distributions can be derived. Also, by measuring 'every' drop, disdrometers have a higher resolution. A drawback of disdrometers is that they usually do not measure rainfall directly, so they have to be calibrated. Because of the nature of disdrometers, this is not something easily done in the field.

2.1.2 Areal observations of rainfall

Although traditionally precipitation over the area is sampled by rain gauges, modern techniques have been applied to obtain areal distributed information on precipitation. Nowadays, more and more products become available (often for free) to give areal estimates for rainfall. The most common are radar and satellites. Although one might think that these areal products replace the necessity of point observations, radar and satellite data are always calibrated and validated with ground observations. Hence point observations remain important.

Radar

A well known method to measure rainfall is a weather radar. Radar is an acronym for **Ra**dio **D**etection **a**nd **R**anging and measures the reflectivity of radio waves. An antenna (see Figure 2.3a) transmits a pulse, which bounces off on any object resulting in an echo. Clouds without rain do not produce an echo, but clouds with rain reflect the pulse. The higher the reflectivity, the higher the rain intensity. By comparing the radar images (see Figure 2.3b) in time, the direction of the shower can be derived. Modern radar systems do this by making use of the Doppler effect. These new systems also give information on the wind speeds inside the storm.

The temporal resolution of radar data can be as small as 5 minutes. Theoretically, the radar can produce rainfall estimates for a circle with a radius of 300 km from the antenna. However, in practice this is limited to a radius of 180-200 km, because of the curvature of the earth. Due to this curvature, the pulse is traveling through higher elevations with increasing distance from the antenna. Since rainfall clouds are usually present in the lower 5 km of the atmosphere, the radar can not observe showers anymore when the pulse is higher than that 5 km. Another problem with radars occurs when large showers block all radio waves, so showers behind the large shower can not be observed anymore.

Rainfall from satellites[1]

Remote sensing data becomes more and more important to estimate rainfall. One of the first empirical relations used was the relation between cloud top temperatures(measured with a thermal band) and rainfall, also called cold cloud duration. It is now widely used with geostationary

1. From Winsemius, H. C., 2009. Satellite data as complementary information for hydrological models. Ph.D. thesis, Delft University of Technology Winsemius [2009]

a)  b) 

Figure 2.3: a) Radar of the KNMI in De Bilt, the Netherlands. b) Example of a radar image.

satellite imagery from Meteosat, Meteosat Second Generation (MSG) and Geostationary Operational Environmental Satellite (GOES). For instance the GOES Precipitation Index (GPI, Arkin and Meisner [1987]) is a precipitation algorithm based on this relation. Cold cloud tops are caused by release of latent heat to the atmosphere during convective rain storms and it is assumed that when a temperature threshold is underspent, a certain rain rate will occur. Estimating daily rainfall is therefore a matter of counting the number of times that underspending of this temperature occurred. This approach is therefore especially suitable for tropical areas, where most of the rainfall is of a convective nature.

Microwave imagers (e.g., Special Sensor Microwave/Imager (SSM/I), Advanced Microwave Soundings Unit (AMSU-B), TRMM Microwave Imager (TMI) and Advanced Microwave Scanning Radiometer - EOS (AMSR-E)) may also be used to retrieve precipitation fields. Two types of algorithms are developed: one based on scattering and one on emission (e.g., Ferraro and Marks [1995]; Zhao and Weng [2002]). The algorithm based on scattering is where the amount of scatter of the microwave signal is dependent on the quantity and size of ice particles and the algorithm based on emission is where the variations in brightness temperature are an indication of present water vapour and rainfall. Problems with the first type of algorithms occur where the rainfall is relatively warm such as oceanic rainfall events due to orographic lifting or shallow convective storms, because there are no or only little ice particles present. For both algorithms, there are some problems with scale. The range or footprint of a SSM/I sensor is in general too large to detect small convective events, since the algorithms are based on footprint-averaged exceeding of thresholds. TRMM is the first mission that carries a Precipitation Radar (PR) to estimate 3-dimensional cloud properties and concurrent rainfall.

The difference between rainfall estimates from some different algorithms is given in Figure 2.4. It becomes clear that microwave algorithms see different rainfall properties than cold cloud duration. Therefore, the best results are expected when different estimates are merged. In general, end user rainfall products are combinations of independent estimates, where the weight of the independent estimate is usually somehow based on local groundtruth rainfall. Examples of two popular rainfall estimates are FEWS RFE 2.0 and TRMM.

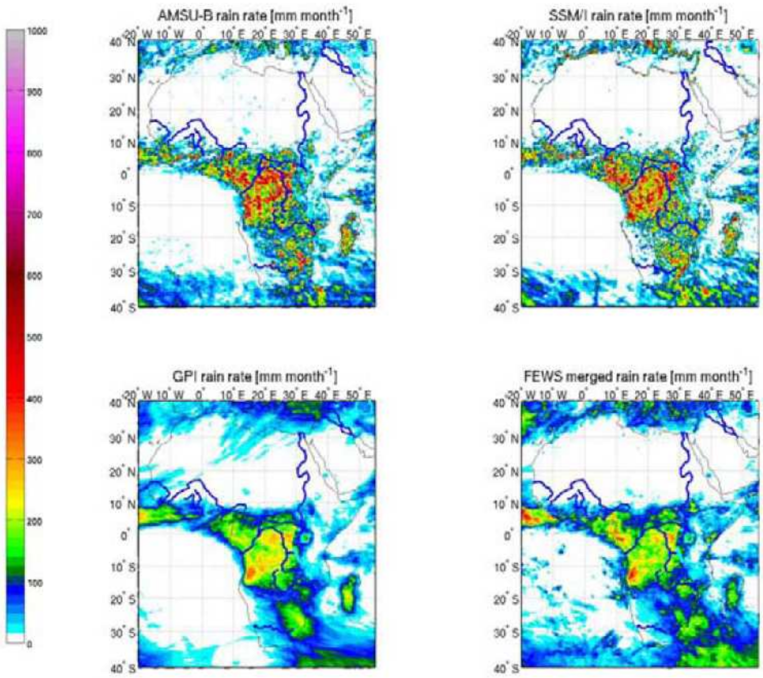

Figure 2.4: Three independent accumulated rainfall estimates and station data based estimates for a 30-day time window (November 1st until 30th 2007) over Africa. From left to right: AMSU-B microwave, SSM/I microwave, Meteosat GPI, and a merged estimate. The merged estimate shows the impact of rain gauges, especially over the Atlantic Ocean.

2.2 Evaporation

2.2.1 Types of evaporation and definitions

Evaporation is the transfer of water from liquid state into gaseous state. Within the framework of the hydrological processes it normally means the average vertical vapour transport from a surface area. Evaporation can occur by direct evaporation of water and by transpiration.

- Direct evaporation:
  - Open water evaporation, E_o [L/T]. This is the amount of vaporized water within a certain time from an open water surface. The availability of the water is hereby not restricted.
  - Soil evaporation, E_s [L/T]. The water that is vaporized from the unsaturated zone. When the soil moisture content is limited, the evaporation is more constrained.
  - Interception evaporation, E_i [L/T]; Evaporation from the wet surface after rainfall. This is more then just interception by leaves, it contains evaporation from all wet surfaces; vegetation, paved surfaces, dropped leaves, bare soil, etc.
  - Snow or ice evaporation, also called sublimation, E_{sub} [L/T]. In cold climates were there is snow or ice sublimation can be considerable. From the solid state water molecules transform into a vapor state.
- Transpiration (E_t [L/T]): The transfer of water vapor into the atmosphere through the stomata of living plants. Just a very small amount of the water absorbed by plants from

the soil remains inside the plant itself. The major part vaporizes from the leaves. Because of the influence of solar radiation, transpiration occurs primarily during daytime. At night the pores close and little water leaves the plant. An exception is for instance the cactus (the stomata only open at night).

Summarizing, direct evaporation is the transfer of liquid into vapour by molecules which escape the water surface; transpiration is a physiological process of water molecules transferring by gas exchange in the stomata from the vegetation into the atmosphere.

The sum of all occurrences of evaporation is called 'actual evaporation', E_a [L/T]. This is the total amount of vapour that is fed back to the atmosphere and is the flux of interest in the water balance equation (Equation 2.1). So the total or actual evaporation is a combination of all previous forms:

$$E_a = E_o + E_t + E_i + E_s + E_{sub} \tag{2.2}$$

The open water evaporation is important to add to the summation when lakes, swamps and rice paddies are involved. Sublimation is only important for areas where snow and ice are dominantly present.

Hence actual evaporation is the flux we need for our water balance, but unfortunately it is often very difficult to determine (as shown later) because it is constrained by the availability of water. Therefore, the 'potential evaporation', E_p, is often used with or without correction factors. The difference between potential evaporation and actual evaporation is that potential evaporation is not constrained by the available water. Resulting in the fact that actual evaporation is thus always lower or equal to the potential evaporation. The potential evaporation indicates the upper boundary of the possible evaporation.

2.2.2 The process of evaporation

Evaporation, the transfer of water from liquid state into gaseous state, uses energy. The reversed process produces the same amount of energy. This energy is known as the latent heat of vaporization; λ. The latent heat λ for a set temperature is defined as the amount of Joule which is used at that temperature to transfer 1 gram of water from liquid state into water vapor. The dependence of λ on temperature proves to be minor. That is why λ is in practice taken as a constant: $\lambda = 2.45 \cdot 10^6$ J/kg $= 2.45$ MJ/kg. The latent heat of vaporization per cubic meter becomes $\rho\lambda = 2.45$ GJ/m^3.

The energy for evaporation is, directly or indirectly, provided by solar radiation. The amount of radiation which is available to evaporation depends on the latitude, the time of the year, atmospheric conditions like the cloudiness, reflection by the earth's surface, atmospheric absorption and storage capacity of the ground or the water. To keep the process of evaporation going, the water vapour has to be removed from the evaporating surface (e.g., by wind). This is a process of diffusion and turbulent transport which is affected by the humidity and temperature of the air, the roughness of the surface and the wind speed.

The resistance to flow of water inside the plant and the percentage of the soil covered by crop plays an important role in the process of transpiration.

Nature of surface	Albedo (r)
Water[1]	0.02-0.90
Moist dark soils; ploughed fields	0.05-0.15
Gray soils, bare fields	0.15-0.25
Dry soils, desert	0.20-0.35
White sand; lime	0.30-0.40
Green grass and other short vegetation (e.g. alfalfa, potatoes, beets)	0.15-0.25
Dry grass; stubble	0.15-020
Dry prairie and savannah	0.20-0.30
Coniferous forest	0.10-0.15
Deciduous forest	0.15-0.25
Forest with melting snow	0.20-0.30
Old and dirty snow cover	0.35-0.65
Clean, stable snow cover	0.60-0.75
Fresh dry snow	0.80-0.90

a. The albedo of water is dependent on the depth of water and the angle of the suns rays

Table 2.1: Albedo (r) for different surfaces (from Brutsaert [2005]).

Summarizing, the (potential) evaporation is affected by factors which are determined by:

A Evaporation surface
- A.1 Reflection coefficient (albedo)
- A.2 Roughness of the surface
- A.3 Fraction covered
- A.4 Crop resistance (resistance to flow of water inside the plant itself)

B Atmospheric conditions
- B.1 Wind velocity
- B.2 Relative humidity
- B.3 Temperature
- B.4 Solar radiation

In the following section these factors will be further elaborated and explained how to measure them.

Factors influencing evaporation

A.1. Reflection coefficient (albedo); Typical values for the reflection of solar radiation reaching the earth's surface are given in Table 2.1. For the influence of reflection on evaporation see the paragraph about solar radiation.

A.2. Roughness of the surface; The turbulent transport of water vapor from the evaporating surface into the atmosphere is largely determined by the roughness (e.g., crop height) of the surface. The resistance to this transport is called the aerodynamic resistance and may for a specific surface (roughness) be written as a function of the wind speed, see the paragraph about wind velocity.

A.3. Heat storage capacity; In climates with a distinct summer and winter period, part of the energy that becomes available in spring is used to warm up the surface. This energy is released during the next autumn and winter. A deep lake or wet soil has a large heat storage capacity. Hence, the evaporating surface will remain relatively cold during spring which affects the evaporation. The subsequent release of heat during the autumn and the winter causes a phase shift in the evaporation cycle of deep lakes as compared to shallow lakes.

A.4. Fraction covered; The fraction of soil that is covered by the crop directly affects the transpiration of the area.

A.5. Crop resistance; Transpiration of a cropped surface is usually less than the evaporation of an open water surface due to the additional resistance of water transport in the plant and the transfer of water vapor through the stomata. For tall crops, however, the increased turbulence lowers the aerodynamic resistance which may result in higher values for the transpiration as compared to open water evaporation. The crop resistance is often taken from literature.

The atmospheric conditions affecting evaporation comprise:

B.1. Wind velocity; ; The aerodynamic resistance to water vapour transport r_a [d/m] is a function of the wind speed. Wind speed is measured with anemometers and is a function of the height above the surface. Values which apply for a height of 2 m are generally used to estimate evaporation. The wind speed at other heights can by approximation be conversed to wind speed at 2 m. The aerodynamic resistance in [d/m] to water vapour transport is computed with the formula:

$$r_a = \frac{245}{0.5u_2 + 0.5}\frac{1}{86400} \tag{2.3}$$

with u_2 as the wind speed [m/s] at a height of two meters measured with an anemometer (Figure 2.5).

Stagnant air in contact with the water surface will eventually approach the (saturation) vapour pressure at the surface and evaporation will cease.

B.2. Relative humidity; Evaporation is very sensitive to the humidity of the air at constant temperature. That is why the actual humidity relative to the saturated state is important. The actual humidity is expressed as the actual vapor pressure e_a. Vapour pressure is the pressure of water particles in the atmosphere. This can be determined in different ways, for example with a psychrometer or by measuring the relative humidity. Before going into details, the saturation vapor pressure has to be explained. The saturation vapor pressure $e_s(T)$ is the maximum vapor pressure of water particles before condensation. Saturation vapor pressure [kPa] is a function of the temperature T and is expressed in the formula:

$$e_s(T) = 0.61 \exp\left(\frac{17.3T}{237 + T}\right) \tag{2.4}$$

Figure 2.5: Anemometer

where T is the temperature in °C.

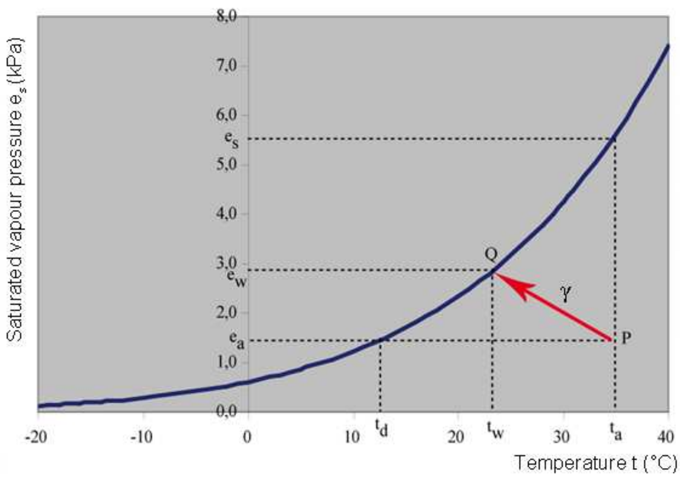

Figure 2.6: Saturation vapour pressure as a function of the air temperature

In Figure 2.6 we see the equation as a graph. Each specific point P in the temperature-vapour pressure space has an actual temperature T_a and an actual vapour pressure e_a. The saturation vapour pressure is reached when the temperature drops to the dew temperature T_d and fog, dew or condensate evolves.
For some computations the slope of the saturation vapour pressure curve ($s = \mathrm{d}e_a/\mathrm{d}t$) is important. This is easily determined by the derivative of Equation 2.5 in [kPa/°C]:

$$s = \frac{4100 e_s}{(237 + T)^2} \tag{2.5}$$

There are two ways to determine the actual vapour pressure e_a:

a. *Determining the actual vapour pressure by the relative humidity;* In case of direct measurement of the relative humidity (h [-]), e.g. by a hygrometer (Figure 2.7) the actual humidity e_a is determined by the relation:

$$h = \frac{e_a(T_a)}{e_s(T_a)} \tag{2.6}$$

where $e_s(T_a)$ is the saturation vapour pressure at the current air temperature T_a according to Equation 2.4.

b. *Determining the actual vapour pressure by a psychrometer;* A psychrometer consists of two thermometers, a dry one and one in a wet state. By means of for example a fan, an air flow is forced past the thermometers. This results in a lower temperature of the wet thermometer than of the dry one due to the extraction of energy of latent heat for the evaporation of water from the wet thermometer. The actual vapour pressure [kPa] is computed with the formula:

$$e_a(T_a) = e_s(T_w) + \gamma(T_a - T_w) \tag{2.7}$$

where:
$e_s(T_w)$ = computed with Equation 2.4 [kPa]
T_a = temperature 'dry' thermometer [°C]
T_w = temperature 'wet' thermometer [°C]
γ = psychrometer constant (0.066 kPa/°C)

In Figure 2.6 the line PQ indicates the situation of a humid medium (e.g., a wet patch) cooling down under the influence of ventilation from the actual temperature T_a to 'wet' temperature T_w. The evaporation involved in this process causes a drop in temperature until a new equilibrium is reached in point Q. At this point the vapour pressure is increased to the saturation vapour pressure of a wet medium $e_s(T_w)$. Note that $-\gamma$ gives the slope of the line PQ.

B.3. Temperature; To measure the temperature properly, only air temperature has to be measured, hence the energy should be supplied by convection and not by radiation, condensation or conduction. For that the thermometer needs to be naturally ventilated and sheltered from sun and rain. This can be done according the WMO standard in a 'Stevenson shelter' (Figure 2.7).

Figure 2.7: Stevenson Shelter with combined hygrometer and temperature meter

B.4. Solar radiation; The latent heat of vaporization is, direct or indirect, provided by solar energy; the latent heat for evaporation λ for water is 2.45 MJ/kg. Solar radiation is therefore the most dominant factor for the determination of evaporation.

The net radiation, R_n, is important for the estimation of the evaporation. This is the difference between the net short wave radiation ($R_{s,n}$) and the net long wave radiation ($R_{l,n}$). The net short wave radiation is the difference between the incoming short wave radiation $R_{s,in}$ and the fraction that is reflected $rR_{s,in}$ (r is the albedo). The net long wave radiation is the difference between the emitted radiation from the surface ($R_{l,out}$) and the long wave radiation that is scattered back by e.g., clouds ($R_{l,in}$).

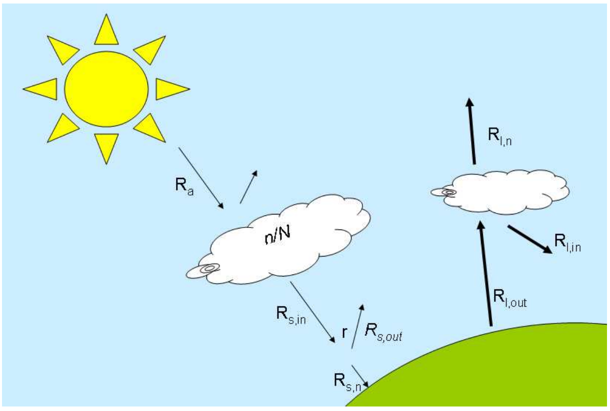

Figure 2.8: Radiation balance.

The net radiation energy is defined as [J $d^{-1}m^{-2}$]:

$$R_n = (R_{s,in} - R_{s,out}) - (R_{l,out} - R_{l,in}) = R_{s,in} - rR_{s,in} - R_{l,n} = (1-r)R_{s,in} - R_{l,n} \quad (2.8)$$

The net radiation can be measured with a radiometer (see Figure 2.9).

Figure 2.9: Radiometer

In the absence of radiation measurements R_n is to be estimated from standard meteorological data using empirical formulae. The procedure is the following. The short wave radiation energy (wave lengths 0.3 - 3 μ m) R_a that is received at the outer limits of the atmosphere, may be read from tables for a given latitude and time of the year. R_a [J $d^{-1}m^{-2}$] divided by the latent heat of vaporization λ [J kg^{-1}] yields equivalent evaporation values R_a/λ in [kg d^{-1} m^{-2}], which is with a density of 1000 kg/m^3 equal to mm/d

(Table 2.2).

Table 2.2: Short wave radiation expressed in equivalent evaporation; R_a/λ (kg m^{-2}d^{-1})

	Lat	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sept	Oct	Nov	Dec
	NORTHERN HEMISPHERE												
	60	1.4	3.6	7.0	11.1	14.6	16.4	15.6	12.6	8.5	4.7	2.0	0.9
	52	3.2	5.5	8.8	12.5	15.4	16.6	16.0	13.6	10.2	6.7	3.9	2.6
	50	3.7	6.0	9.2	12.7	15.5	16.6	16.1	13.7	10.4	7.1	4.4	3.1
	40	6.2	8.4	11.1	13.8	15.9	16.7	16.3	14.7	12.1	9.3	6.8	5.6
	30	8.1	10.5	12.8	14.7	16.1	16.5	16.2	15.2	13.5	11.2	9.1	7.9
	20	10.8	12.4	14.0	15.2	15.7	15.8	15.8	15.4	14.4	12.9	11.3	10.4
	10	12.8	13.9	14.8	15.2	15.0	14.8	14.9	15.0	14.8	14.2	13.1	12.5
Equator	0	14.6	15.0	15.2	14.7	13.9	13.4	13.6	14.3	14.9	15.0	14.6	14.3
	10	15.9	15.7	15.1	13.9	12.5	11.7	12.0	13.1	14.4	15.4	15.7	15.8
	20	16.8	16.0	14.5	12.5	10.7	9.7	10.1	11.6	13.6	15.3	16.4	16.9
	30	17.2	15.8	13.5	10.9	8.6	7.5	7.9	9.7	12.3	14.8	16.7	17.5
	40	17.3	15.1	12.2	8.9	6.4	5.2	5.6	7.6	10.7	13.8	16.5	17.8
	50	16.9	14.1	10.4	6.7	4.1	2.9	3.4	5.4	8.7	12.5	16.0	17.6
	60	16.5	12.6	8.3	4.3	1.8	0.9	1.3	3.1	6.5	10.8	15.1	17.5
	SOUTHERN HEMISPHERE												

$R_{s,in}$ is the short wave radiation that is received at the earth's surface. Its value depends on the local atmospheric conditions (e.g., smog) and the cloudiness n/N. Where n is the number of actual hours of sunshine and N is the number of possible hours of sunshine. The actual hours of sunshine are easily measured with a Cambell-Stokes sunshine recorder (see Figure 2.10). The possible hours of sunshine are read in Table 2.3.

Figure 2.10: Cambell-Stokes sunshine recorder.

Table 2.3: Mean daily duration of maximum possible sunshine hours N

North Lats. South Lats.	Jan July	Feb. Aug.	Mar. Sept	Apr. Oct.	May Nov.	June Dec.	July Jan.	Aug. Feb.	Sept Mar.	Oct. Apr.	Nov. May	Dec. June
60	6.7	9.0	11.7	14.5	17.1	18.6	17.9	15.5	12.9	10.1	7.5	5.9
58	7.2	9.3	11.7	14.3	16.6	17.9	17.3	15.3	12.8	10.3	7.9	6.5
56	7.6	9.5	11.7	14.1	16.2	17.4	16.9	15.0	12.7	10.4	8.3	7.0
54	7.9	9.7	11.7	13.9	15.9	16.9	16.5	14.8	12.7	10.5	8.5	7.4
52	8.3	9.9	11.8	13.8	15.6	16.5	16.1	14.6	12.7	10.6	8.8	7.8
50	8.5	10.0	11.8	13.7	15.3	16.3	15.9	14.4	12.6	10.7	9.0	8.1
48	8.8	10.2	11.8	13.6	15.2	16.0	15.6	14.3	12.6	10.9	9.3	8.3
46	9.1	10.4	11.9	13.5	14.9	15.7	15.4	14.2	12.6	10.9	9.5	8.7
44	9.3	10.5	11.9	13.4	14.7	15.4	15.2	14.0	12.6	11.0	9.7	8.9
42	9.4	10.6	11.9	13.4	14.6	15.2	14.9	13.9	12.6	11.1	9.8	9.1
40	9.6	10.7	11.9	13.3	14.4	15.0	14.7	13.7	12.4	11.2	10.0	9.3
35	10.1	11.0	11.9	13.1	14.0	14.5	14.3	13.5	12.4	11.9	10.3	9.8
30	10.4	11.1	12.0	12.9	13.6	14.0	13.9	13.2	12.4	12.0	10.6	10.8
25	10.7	11.3	12.0	12.7	13.3	13.7	13.5	13.0	12.3	12.0	10.9	10.6
20	11.0	11.5	12.0	12.6	13.1	13.3	13.2	12.8	12.3	12.0	11.2	10.9
15	11.3	11.6	12.0	12.5	12.8	13.0	12.9	12.6	12.2	12.0	11.4	11.2
10	11.6	11.8	12.0	12.3	12.6	12.7	12.6	12.4	12.1	12.0	11.6	11.5
5	11.8	11.9	12.0	12.2	12.3	12.4	12.3	12.3	12.1	12.0	11.9	11.8
Equator 0	12.0	12.0	12.0	12.0	12.0	12.0	12.0	12.0	12.0	12.0	12.0	12.0

In Table 2.4 examples of empirical expressions for $R_{s,in}$ as a function of R_a and the pro-

Table 2.4: Empirical relations for $R_{s,in}$

The Netherlands	$R_{s,in} = (0.20 + 0.48\ n/N)R_a$
Average climate	$R_{s,in} = (0.25 + 0.50\ n/N)R_a$
New Delhi	$R_{s,in} = (0.31 + 0.60\ n/N)R_a$
Singapore	$R_{s,in} = (0.21 + 0.48\ n/N)R_a$

portion n/N are read. The net short wave radiation is equal to $(1-r)R_{s,in}$. To determine the net radiation R_n the net outgoing long wave radiation has to be known. The net outgoing long wave radiation $R_{l,n}$ [J d $^{-1}$ m^{-2}] may be estimated from the following empirical formula:

$$R_{l,n} = \sigma(273 + t_a)^4(0.47 - 0.21\sqrt{e_a})(0.2 + 0.8\frac{n}{N}) \tag{2.9}$$

Where σ is the Stefan-Boltzmann constant ($\sigma = 4.9{\cdot}10^{-3}$ J d^{-1} m^{-2} K^{-4}), T_a is the temperature of the air in °C and e_a is the actual vapour pressure of the air in kPa.

2.2.3 Measuring evaporation

There exist several ways to measure evaporation directly or more often indirectly. In Table 2.5 an overview of some of the techniques are shown. It is important to realize what kind of evaporation is measured. Does the method gives you an estimate for the potential evaporation or for the actual evaporation. And what type of evaporation is measured? Evaporation by open water (E_o), transpiration (E_t), soil (E_s), or evaporation of intercepted water (E_i)?

In the following paragraph the most common methods are explained. A more detailed overview on actual evaporation measurements is given in the course CIE4440 'Hydrological Measurements'.

Penman Equation

The most well known method to calculate open water evaporation E_o is the method of Penman (Penman [1948]), which has found world-wide application because it has a strong physical basis. This formula, which may be used for estimating the potential evaporation E_p, is written as:

$$E_p \approx E_o = \frac{\frac{sR_n}{\rho\lambda} + \frac{c_p\rho_a}{\rho\lambda}\frac{(e_s-e_a)}{r_a}}{s+\gamma} \tag{2.10}$$

With:

Table 2.5: Overview of methods to measure evaporation.

	Potential		Actual			
	E_o	E_t	E_o	E_t	E_s	E_i
Penman	x		x			
Penman-Monteith		x				
Pan evaporation	x	x	x			
Lysimeter				x[1]	x[2]	x[1]
Sapflow				x		
Bowen ratio			x	x	x	x
Eddy correlation			x	x	x	x
Scintillometer			x	x	x	x
Energy balance			x	x	x	x

[1] If vegetation in lysimeter.

[2] If soil layer in lysimeter.

R_n net radiation on the earth's surface [J d^{-1} m^{-2}]
λ latent heat of vaporization [J/kg] (2.45 MJ/kg)
s slope of the saturation vapour pressure-temperature curve [kPa/°C] (see Eq.2.5)
c_p specific heat of air at constant pressure [J kg^{-1} K^{-1}] (1004 J kg^{-1} K^{-1})
ρ_a density of air [kg/m^3] (1.205 kg/m^3)
ρ density of water [kg/m^3] (1000 kg/m^3)
e_a actual vapour pressure in the air at 2 m height [kPa]
e_s saturation vapour pressure for the air at 2 m height [kPa]
γ psychrometer constant [kPa/°C] (0.066 kPa/°C)
r_a aerodynamic resistance [d/m]

The method described above only needs four standard meteorological parameters:

- Net radiation (or at least sunshine hours)
- Wind velocity
- Relative humidity
- Air temperature

The required data of the meteorological observations are 24 hour means at a height of 2 meters above the soil surface.

Penman-Monteith Equation

The formula of Penman for open water evaporation is also used as a reference to estimate the potential transpiration ($E_{p,t}$) of a crop; therefore the crop resistance is introduced (Monteith [1965]):

$$E_{p,t} = \frac{\frac{sR_n}{\rho\lambda} + \frac{c_p\rho_a}{\rho\lambda}\frac{(e_s-e_a)}{r_a}}{s+\gamma(1+\frac{r_c}{r_a})} \tag{2.11}$$

where r_c [d/m] is the crop resistance. The crop resistance depends on the availability of soil moisture. If a crop is supplied with an abundant amount of water, the crop resistance reaches a minimum level; the transpiration is equal to the potential transpiration $E_{p,t}$.

The relation between crop resistance and soil moisture availability is crop dependent and difficult to evaluate. To come to the total evaporation, the evaporation of interception water and the direct evaporation of water on the ground still has to be taken into account.

To compute the water requirements for irrigation purposes the maximum (potential) transpiration of a crop $E_{p,t}$ is normative. It is not always possible to use Equation 2.11 because some e.g., meteorological values are missing or r_c can not be estimated. In that case it is common practice to use the reference evaporation E_{ref}. The reference evaporation is defined as the evaporation from an idealized grass crop with a fixed crop height of 0.12 m, an albedo of 0.23 and a r_c=70 s/m:

$$E_{p,t} = k_c E_{ref} \tag{2.12}$$

where k_c is the crop factor. The crop factor expresses the factors that are related with the evaporation surface, for instance the type of crop and the stage of growth the crop is in, so k_c is actually time dependent.

More information on the Penman-Monteith equation and calculation examples can be found in the FAO manual (Allen et al. [1998]).

Pan evaporation

The most straightforward method to measure open water evaporation directly is with a pan (Figure 2.11). The pan is filled with water and every time interval (e.g., daily) the water level is measured. The drop in the water level divided by the time interval gives the rate of open water evaporation. Of course, the water level difference should be compensated for rainfall. Therefore, next to the pan a rain gauge should be installed.

Various types of pans are in use, such as the sunken pan, floating pan, and surface pan. The use of the surface pan is most widely spread, despite its shortcomings. The most common is the U.S. Weather Bureau Class A-pan, with a diameter of 4ft and depth of 10 inch. The water level is maintained 2-3 inches below the rim (Figure 2.11).

Although the Class A-pan is a direct method, one should be careful with assuming that the pan evaporation is an estimate for the open water evaporation. Often the pan evaporation is higher than the open water evaporation due to warming of the sides by the sun. To reduce this effect the pan can be installed on floats in a lake or one can make use of correction factors (see e.g., Grismer et al. [2002] for an overview).

2.3 Soil water

Soil water is the water that is stored in the unsaturated zone. Thus the zone between the surface and the ground water table, where the pores in the soil are partly filled with water and partly

Figure 2.11: Class A-pan

with air. The unsaturated zone (or vadose zone) is the buffer between infiltrated water and percolated water to the ground water and is mainly important for plants.

The water content in the soil can be measured directly by the gravimetric method, whereby a soil sample is taken from the field and taken to a lab. The sample is successively weighted in the 'wet' state (M_{sample}) and afterwards completely saturated and again weighted (M_{sat}). Next, the sample is dried in an oven and weighted for the last time (M_{dry}). The volumetric soil moisture content θ ($0 < \theta < 1$) is then expressed as:

$$\theta = \frac{M_{sample} - M_{dry}}{M_{sat} - M_{dry}} = \frac{V_w}{V_t} \tag{2.13}$$

Although this method is cheap and straightforward, it has the disadvantages that it is destructive and that it can not be used for continuous measurements, which is often required. The indirect methods are more appropriate for this purpose.

There are several ways to measure soil moisture indirectly. In Vereecken et al. [2008] and Robinson et al. [2008] a good overview is given. The most common methods are those that make use of the dielectric permittivity (Capacitance Probe, Time Domain Reflectometry (TDR), and Frequency Domain Reflectometry (FDR)) or the thermal conductivity (Heat Pulse Sensors). Furthermore, also remote sensing techniques become more important for soil moisture measurements.

2.3.1 Capacitance Probe

The capacitance probe uses the capacitance to measure the dielectric permittivity of the surrounding soil, which is a function of soil water. The probe measures the charge time of a capacitor which uses the soil and water as a dielectric. The charge time is related to capacitance, which is again a function of dielectric permittivity. The dielectric permittivity of water is 80, while soil has a dielectric permittivity in the order of 2-5. Hence when the soil contains more water the dielectric permittivity will increase. This relation is used to measure the soil moisture content.

The dielectric permittivity is measured with a so-called capacitor sensor, which consists of a number of electrodes: either two circular rings or an array of parallel metal spikes (Figure 2.12).

The rings or spikes form the plates of the capacitor with the soil in between acting as a the dielectric.

Figure 2.12: Capacitance probe (*10HS Soil Moisture Sensor - Decagon*)

2.3.2 Time Domain Reflectometry (TDR)

A TDR probe consists of a device with two or tree pins attached to it (Figure 2.13). The devices transmits a high frequency electromagnetic wave and measures the travel time between transmitting and the reflected signal. The shorter the travel time, the lower the dielectric permittivity, and thus the lower the water content.

TDR probes are relatively expensive, but highly accurate if properly calibrated. Compared to the capacitance probes, the TDR probes are less sensitive for salinity and temperature.

Figure 2.13: Time Domain Reflectometry (*Trime Data Pilot System - IMKO*)

2.3.3 Frequency Domain Reflectometry (FDR)

The FDR-method (Figure 2.14) uses a similar principle as the TDR; however, while the TDR measures the travel time of an electromagnetic pulse, the FDR-method uses the difference in transmitted and reflected frequency of radio waves. The advantages of the FDR in comparison to the TDR are the price and the sensitivity to temperature and salinity.

Figure 2.14: Frequency Domain Reflectometry (*PR2 Profile Probe - Delta-T Devices*)

2.3.4 Heat Pulse Sensors

Heat pulse sensors are relatively a new methodology to measure soil moisture. Campbell et al. [1991] introduced the method where two needles are inserted into the soil. One of the needles is heated with a short heat pulse and the temperature response at the second sensor is measured. In this way the soil's volumetric heat capacity can be determined, which is a function of soil moisture. Campbell et al. [1991] used two needles, but current heat pulse sensors have more needles to improve the accuracy of the measurement and are often also combined with dielectric measurements.

2.3.5 Remote Sensing

Nowadays, soil moisture is more and more measured by remote sensing techniques, like passive microwave radiometers, synthetic aperture radars, scatterometers or thermal methods (Wagner et al. [2007] and Drusch et al. [2004]). Although remote sensing data is promising for the future, currently is suffers from problems with spatial averaging and the small penetration depth. The spatial resolution of the satellite is so large that the measurements do not reflect the heterogeneity of the soil moisture which is important. Furthermore, the satellite is only able to detect the moisture in the top layer of the soil, while measurements of the moist in the root zone are much more important.

Further readings

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2.4 Ground water

Measurements in groundwater always require the drilling of a hole and the installation of an observation well or piezometer (pronounce: piezómeter). The main measurement in groundwater is the hydraulic head, which is defined as the elevation to which water rises in an observation well. Heads are commonly measured with respect to the top of the observation well. To be able to compare heads at different observation wells, the elevations of the tops of the observation wells need to be measured relative to eachother, or, better yet, relative to a datum (in the Netherlands: NAP). Historically, heads were measured by hand with a tape measure and measurements were taken infrequently. The standard measurement frequency in the Netherlands used to be twice a month, while once a year or twice a year is more common in many other places in the world. In the past decade, it has become popular to use small pressure transducers. These pressure transducers are sometimes called “divers”, which is the brand name of a popular pressure transducer fabricated by Schlumberger Water Services (see Fig. 2.15). Pressure transducers are lowered in the observation well below the water table and may be programmed to take and store pressure measurements at a specified frequency. As groundwater generally moves very slowly, it is rarely necessary to measure more frequently than every hour, while it is often sufficient to measure once a day or once a week.

A pressure transducer measures the absolute pressure, which needs to be converted to head. The relationship between absolute pressure and head depends on the elevation, the atmospheric pressure and the density of the water, which in turn is a function of the temperature of the water. The temperature is automatically recorded by the pressure transducer. Atmospheric pressure needs be be measured with a separate pressure transducer. The spatial variation of atmospheric pressure is small, so that one atmospheric pressure transducer may be sufficient to correct pressure transducers in a larger area. It is also possible to use special vented pressure transducer, which are connected to a small plastic tube that extends till above the water table.

Figure 2.15: A pressure transducer used to measure groundwater heads

The density of the water may also be affected by the salinity of the water, especially in coastal areas. Conversion of a pressure to a head requires knowledge of the density (and thus the salinity) over the entire depth of the observation well. For accurate measurement in coastal areas, a constant salinity in the observation well may be obtained by pumping water from the observation well until the salinity of the water doesn't change anymore. Pressure transducers are relatively accurate devices, but they need to be calibrated before installation and it needs to be realized that they may depart from the calibrated state over time.

Measurements of groundwater flow is difficult if not impossible because the velocity is generally very low (less than a meter per day) and measurements can only be taken in an observation well. Aquifer parameters, especially the hydraulic conductivity and the storage coefficient, may be measured with a pumping test. In a pumping test, water is extracted at a known rate from a well and the head response is measured in one or more nearby observation wells. Many approaches have been developed to infer the aquifer parameters by relating the discharge of the well to the head response.

2.5 Discharge

Stream flow is the result of all climatological and geographical factors that operate in a drainage basin. It is the only phase of the hydrological cycle in which the water is confined in well-defined channels which permit accurate measurements to be made of the quantities involved.

Good water management is founded on reliable stream flow information and the final reliability of the information depends on the initial field measurements. The hydrologist making these measurements has therefore the responsibility of ensuring that raw data of acceptable quality is collected. The successful processing and publication of the data depend largely on the quality of the field measurements. This Section will focus on the following methods.

- Velocity area method

  The area of the cross-section is determined from soundings or manual measurements, and flow velocities are measured using current- meters, electro-magnetic sensors, or floats. The mean flow velocity is deduced from velocities measured at points distributed systematically over the cross-section. In principle the measurements should be made under steady conditions.

- Dilution method
  A suitable selected tracer (salt) is injected at the upstream section of the measurement reach with the Mariotte Bottle or Gulp Method. In a downstream sampling section, where the tracer should be uniformly distributed throughout the cross-section, samples are taken at regular time intervals. The discharge of the stream can be calculated from the concentration of the tracer injected at the upstream section of the reach and the concentration measured at the downstream end of the reach.
- Stage-discharge method
  The stage-discharge method makes uses of the empirical relation between water level (stage, h) and discharge, Q. The Qh-relation is found by measuring the discharge and water level for a wide range of discharges. This relation is called the rating curve. If the rating curve is known only the water level has to be measured to estimate the discharge. The Qh-method can be applied for both natural rivers as for weirs and flumes (structures).

In this Section just a short overview of discharge measurements is given. For more details the STOWA manual (Hartong and Termes [2009] in Dutch) is a very good reference or Boiten [2008].

2.5.1 Velocity area method

The velocity area method is a method to obtain a near instantaneous value of the discharge of a river. The discharge Q of a river is being measured by means of measuring velocities (u_i), representative for the part A_i of the area (A), in a suitable cross-section. The method of approach is:

$$\int^{A} u \mathrm{d}A \approx \sum^{n} u_i \cdot \Delta A_i \tag{2.14}$$

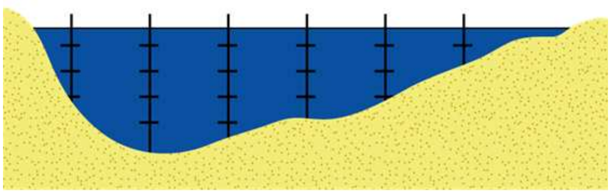


Figure 2.16: Example 'velocity-area-method'

The cross section is divided into segments by selecting verticals at a sufficient number of points to ensure an adequate sample of bed profile and velocity distribution. The spacing of the verticals can vary from site to site and depends on the width, the character and unevenness of the bed and the variations in velocity distribution throughout the cross section. In general a reach or cross section is preferred having the following characteristics:

- A straight reach with the threads of velocity parallel to each other
- A stable streambed free of large rocks, weeds or other obstacles that create turbulence

- A flat streambed profile to eliminate vertical components of velocity

The velocity area method uses point information of flow velocity in the considered cross-section. Various instruments have been designed to obtain these velocities. The following instruments will be described:

- Propeller current meter
- Floats
- Doppler current meter

Propeller- type current meters

The propeller types of current meters use the principle of proportionality between the local flow velocity and the resulting angular velocity of the rotor of the instrument. An electrical circuit closed intermittently by contact points measures the number of the revolutions of the rotor. The interruptions are transmitted to an acoustic or electric counter.

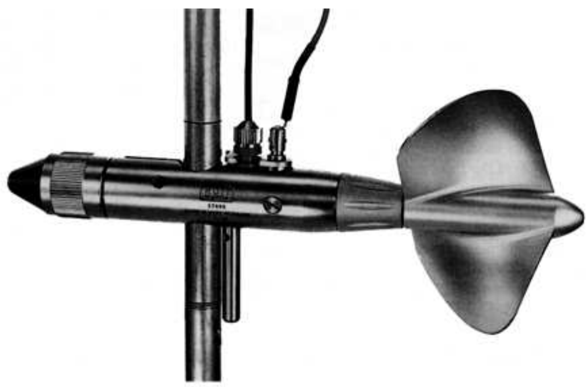


Figure 2.17: Propeller type current meter

Calibration of these types of current meters is carried out by experimental determination of the curves representing the relationship between the velocity of flow and the propeller or rotor speed, usually expressed in revolutions per second. Current meters may be calibrated individually or a group rating may be supplied by the manufacturer for a particular type of meter. A calibration curve, a formula, or a rating table is supplied by the manufacturer and the actual limits of validity are also indicated.

The propeller current meter has a horizontal axis to which a propeller is attached. The flow makes the propeller turn and thus the axis.

Current meter measurements by wading are preferred if conditions permit. Wading measurements with the current meter supported on a graduated wading rod, which rests on the bed of the stream are normally more accurate than those from cableways or bridges as the operator has more control over the general gauging procedure. The position of the operator is important to ensure that the operator's body does not affect the flow pattern when approaching the current meter. The best position is to stand facing one of the banks, slightly downstream of the meter and an arm's length from it. The rod is kept vertical throughout the measurement with the meter parallel to the direction of flow.

Figure 2.18: Measuring stream flow by wading

A measuring tape or tag line is stretched across the river at right angles to the direction of flow. The positions of successive verticals used for depth and velocity are located by horizontal measurements from a reference marker (initial point) on the bank, usually defined by a pin or a monument. The gauging starts at the water edge of the near bank, where depth and velocity may or may not be zero. At each chosen vertical the depth is measured and the value used to compute the setting or settings of the current meter depending on the method to be used (e.g. at 0.6 or 0.2 and 0.8 depth).

The cross section is divided into segments by selecting verticals at a sufficient number of points to ensure an adequate sample of bed profile and velocity distribution. The spacing of the verticals can vary from site to site and depends on the width, the character and unevenness of the bed and the variations in velocity distribution throughout the cross section.

The spacing of verticals is determined by the required accuracy. An elaborate study in ISO-connection has shown that it is preferred to have many verticals in the cross-section rather than many measuring points in each vertical. For large rivers at least 20 verticals are preferred. For small streams the actual distance (75 cm-100 cm) between verticals determines the number of verticals.

It is advised to have several (2-3) measuring points in each vertical when there is sufficient depth (e.g. at 0.6 or 0.2 and 0.8 depth measured from the surface). Especially for small watercourses it is usual to limit the number of measuring points in the vertical.

The average velocity $\overline{v_w}$ for a three point measurement in the vertical is calculated with equation 2.15 or 2.16.

$$\overline{v_w} = \frac{1}{3}(v_{0.2} + v_{0.6} + v_{0.8}) \tag{2.15}$$

$$\overline{v_w} = \frac{1}{4}(v_{0.2} + 2 \cdot v_{0.6} + v_{0.8}) \tag{2.16}$$

Alternatively the velocity is measured at one point at 0.6 of the depth measured from the surface or at 0.5 of the depth with equation 2.17 or 2.18.

$$\overline{v_w} = v_{0.6} \tag{2.17}$$

$$\overline{v_w} = 0.95 \cdot v_{0.5} \tag{2.18}$$

When the current meter is installed at the proper depth in a vertical, a point measurement of the current can start. As the current usually is turbulent the measurement of flow velocity is averaged over a certain period of time. Usually a time of measurement for approximately one minute will be sufficient. Some counters will immediately specify the averaged flow velocity; others will specify the number of revolutions of the propeller over the time of measurement. In the latter case this has to be converted into an average flow velocity, according to the method as described by the supplier of the apparatus.

With some equipment different propellers are applied to different flow velocities. Therefore a first estimate of flow velocity is required. Each propeller will have its own conversion equation or table.

Calculations are based on computing a partial discharge through partial sections. Thus, for example, the discharge through partial section 4 (outlined in Fig. 2.19) is

$$q_4 = \overline{v_4}(\frac{b_5 - b_3}{2})d_4 \tag{2.19}$$

q_4 = Averaged velocity in vertical number 4 (m/s)

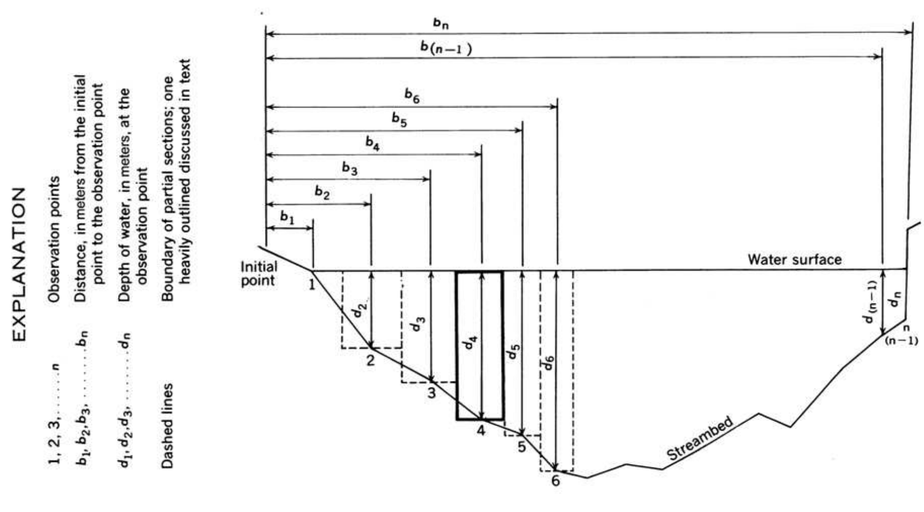


Figure 2.19: River sections

For the first vertical one applies

$$q_1 = \overline{v_1}(\frac{b_2 - b_1}{2})d_1 \tag{2.20}$$

For the last vertical one applies

$$q_n = \overline{v_n}(\frac{b_n - b_{n-1}}{2})d_n \tag{2.21}$$

Floats

Floats are the simplest tools for measurement of flow velocity. The time it takes the float to drift over a known distance between two previously fixed transversal lines is a measure of the flow velocity. Examples of surface floats and subsurface floats are shown in Fig. 2.20.

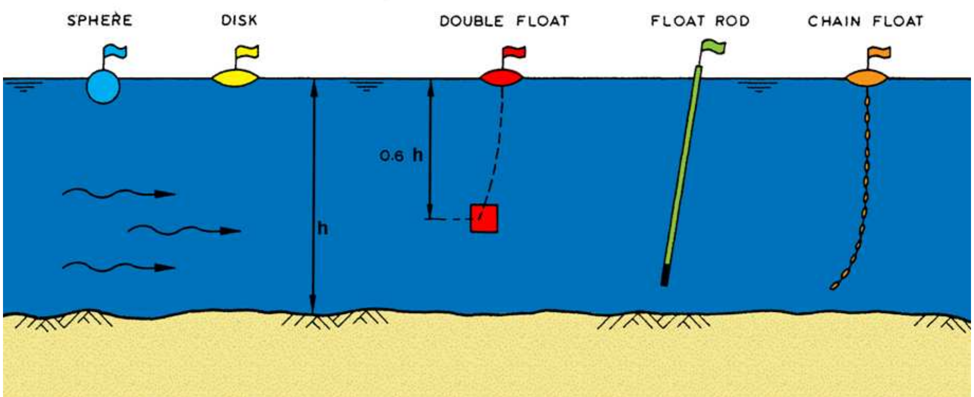


Figure 2.20: Floats

Two disadvantages of surface floats are that they measure the flow velocity at the water surface only and that they are influenced by the wind. Better information on the mean velocity in the vertical can therefore be obtained using subsurface floats which can be constructed for example by two bottled partly filled with sand. The only problem remains getting the average velocity over the cross-section as the variation over the cross-section may be substantial and appears to be the largest source of errors. Therefore, one should select multiple measuring positions in the cross-section.

Doppler current meter

An instrument, that measures the flow velocity by means of the Doppler effect, is the Acoustic Doppler Current Profiler or ADCP. ADCP's are used to measure discharges in open channels, according to the velocity-area method. There are either downward looking ADCP's, measuring velocity profiles from near the water surface to the bed, or upward looking ADCP's, installed on the bed. An example of a downward looking installation is the vessel mounted ADCP, which is in use since the mid-1980's.

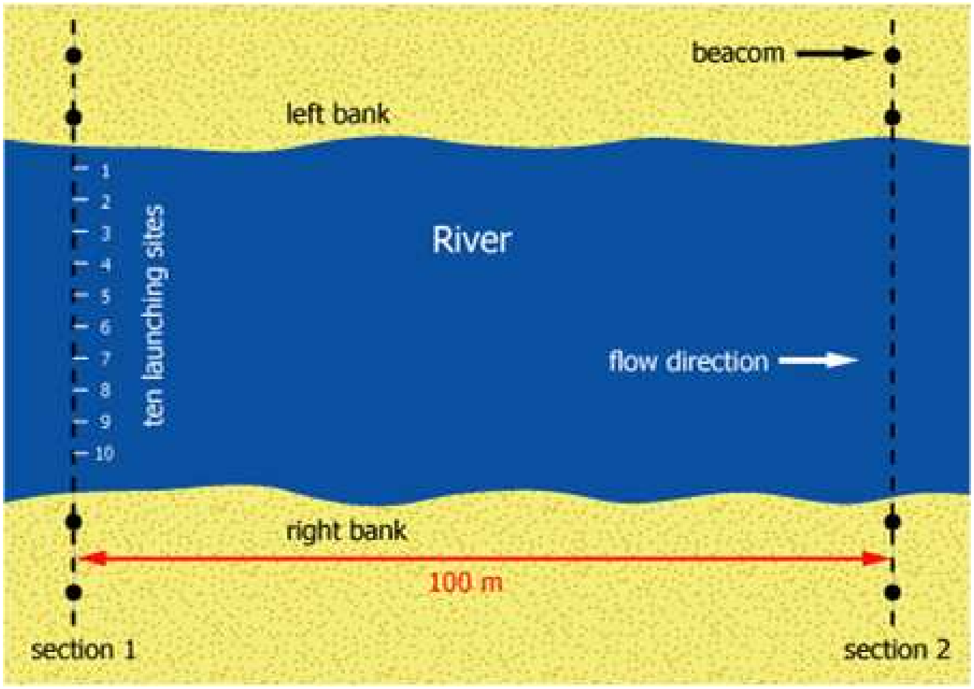


Figure 2.21: River section

The Doppler effect is a change in the observed sound pitch that results from relative motion. Doppler shift is the changed frequency observed from a moving sound source, compared to the stagnant situation.

ADCP's use the Doppler effect by transmitting sound at a fixed frequency and listening to echoes returning from sound scatters in the water. These sound scatters are small particles (suspended load) that reflect the sound back to the ADCP. Scatters are everywhere in rivers. They float in the water and on average they move at the same horizontal velocity as the water (Note that this is a key assumption!). Sound scatters in all directions form scatters. Most of the sound goes forward, unaffected by the scatters. The small amount that reflects back is Doppler shifted. When sound scatters move away from the ADCP, the sound you hear is Doppler shifted to a lower frequency proportional to the relative velocity between the ADCP and the scatter.

2.5.2 Dilution gauging

Dilution gauging is, as is the velocity area method, a method to obtain a near instantaneous value of the discharge of a river. The principle of dilution gauging is simple. Added to a constant unknown discharge Q, is a volume of water with a known concentration ϕ_0. The diluted concentration ϕ_1 is measured downstream. The degree of dilution is a measure for the unknown discharge Q. Depending on the way of injection two methods can be distinguished, i.e., constant rate injection or sudden injection. The dilution gauging is preferably suitable for small watercourses with a lot of turbulence as well as for pumping stations. Nevertheless, accurate measurements were achieved with discharges up to 2000 m3/s. With precise execution the relative error in the discharge can be in the order of 1%.

Constant rate injection method

The injection consists of a constant discharge ΔQ, with concentration ϕ_0 that is added to the river discharge Q, during a long enough time-period. If on the reach L there is sufficient mixture, then after a certain time period a constant concentration ϕ_1 will be measured at the end of the reach. From the mass-balance of the concentrations it follows by approximation (when $\Delta Q << Q$):

$$\phi_1 = \frac{\Delta Q}{Q}\phi_0 \tag{2.22}$$

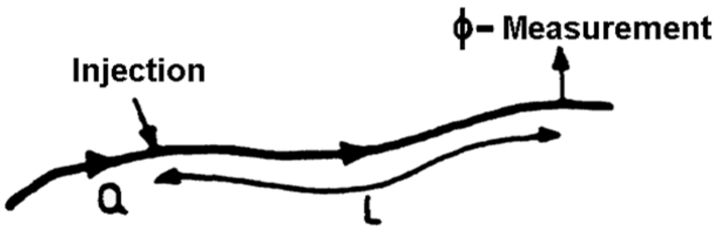


Figure 2.22: Dilution gauging for reach L

For the constant rate injection method it is important to keep ΔQ constant. This is possible by injecting the tracer into the river with adjustable pump. It is also possible to use the principle of the bottle of Mariotte (1620-1684).

Sudden or Gulp injection method

By this method on $t = 0$ a tracer mass (M) is instantaneously injected in the river. On a distance L downstream (where the concentrate is assumed uniformly diluted) the mass concentration $\phi(t)$ is being measured. The following formula applies here:

$$M = Q\int_0^\infty (\phi(t) - \phi_1)\,\mathrm{d}t \tag{2.23}$$

Where ϕ_1 is the possible existent natural concentration in the watercourse, which can be determined upstream from the injection point. Theoretically ϕ_1 can also be a function of time. The condition is that $\phi_1 << \phi(t)$ should apply. When M is known and $\phi(t)$ en ϕ_1 are measured the discharge Q in equation 2.23 is the only unknown.

Mixing length

Dilution gauging works best in strong turbulent water (and thus good mixture). For mountain creeks this method is ideal. In fact, other methods based on velocity measurements in mountain creeks are usually less suitable, due to the irregular cross-sections. A theoretical derivation for the mixing length, L, is given by the equation of Rimmar (Rimmar [1953]):

$$L \geq \frac{0.13b^2C(0.7C+2\sqrt{g})}{gd} \tag{2.24}$$

$$\text{for } 4.8 < \frac{C}{\sqrt{g}} < 16$$

With:
C Chézy roughness ($m^{1/2}/s$)
b River width (m)
d River depth (m)

The mixing length assumes adding of the concentrations from either bank of the river. Obviously the necessary length of the reach is shorter if one injects in the center of the watercourse. In that case the mixing length is determined by applying half the width of the river. See also the lecture notes of CIE4400 Water Quality Modeling.

A way to test whether or not the water has mixed completely is by applying a coloring agent. One can use the coloring agent uranine (sodium fluorescein). This can be toxic for the environment. The LC50 (value at which half of the population is killed) for fish is 2200 mg/l during 96-h, 165 mg/l for Cladocera (water fleas) during 48 hours and less than 10 mg/l for algae during 96 hours (Field et al. [1995]).

Measuring concentration

The concentration applied for the dilution gauging method depends on the used tracer and measuring equipment. For coloring agent (e.g., rhodamine WT or fluorescent) a colorimeter is being used. For salt solutions (e.g., cooking salt) the conductivity is being measured. The conductivity meter has to be calibrated against concentrations of e.g. salt solutions for a range of values that are to be expected (see below).

In order to prepare the concentrations of salt to add to the river, one should estimate the discharge Q of the river and have an idea of the measurable concentrations. For the constant rate injection method the rate of injection ΔQ has to be determined as well. A comfortable concentration of NaCl that can be measured with a conductivity meter is 0.1g/l.

When the sudden or gulp injection method is applied the total amount M (g) of salt that will be added to the river will result in an increased concentration wave at the measuring point L downstream. Approximately the concentration wave would last 5 to 10 minutes (300-600 sec). To obtain an average concentration $\phi(t) = 0.1$g/l for, say 10 minutes, the total amount of salt that is required is:

$$M(g) = Q(l/s) \cdot 600(s) \cdot 0.1(g/l) \tag{2.25}$$

With e.g. Q=10(l/s), M=600 g. This amount should be dissolved in the proper amount of water before it is added to the stream.

Calibration of conductivity meter

The conductivity meter has to be calibrated for the expected range of concentrations. If the expected range of average concentration $\phi(t)$=0.1g/l and the Gulp method is applied the calibration range could be from 2 g/l to 0.06251 g/l. This could be done by diluting 8 gram of salt in 1 liter of water, divide in half, fill up to 1 liter and repeat this seven times.

When the dilution is done with the river water there is no need to correct for the background concentration in the calculation of the discharge (assuming a constant background concentration). Alternatively the calibration is done with demineralized water and a correction is made for the background concentration ϕ_1 of the river water.

Conductivity is temperature dependent. Therefore, most conductivity meters have a temperature sensor and in this way automatically convert the measured conductivity to the conductivity at a standard temperature (usually 20 or 25 °C).

2.5.3 Stage-discharge method

Rating curve

Determination of the discharge of a river by direct measurement is either laborious and time consuming or expensive. It is therefore common practice to establish a relationship (rating curve) between the stage at a particular gauging station and the discharge. A continuous record of discharge can than be obtained from a continuous record of stage. The relationship is based on the correlation between discharge and stage. This chapter deals with discharge curves at natural control sections, i.e., undisturbed reaches of a river or canal. Rating curves for structures e.g., weirs, flumes etc. can be derived from the characteristics of the structure.

Site selection
Often the use of the discharge data roughly determines the location of the gauging station. Nevertheless, one must take river properties into account in order to establish a reliable and stable relation.

If the channel is stable, relatively a few measurements may be required although very few rivers have completely stable characteristics. The calibration therefore cannot be carried out once and for all, but has to be repeated as frequently as required by the rate of change in the stage-discharge relation. In order to define the relation in sand-bed channels, for example, several discharge measurements a month might be required. In particular, surveys are required after flood flows, when not only the cross-sectional area might have changed, but even the course of the river. This indicates that selecting a stable site e.g. on rock bottom and between rock outcrops is preferred. A stable section where the stage-discharge relation does not change with time is called permanent. When it does change with time it is referred to as a shifting control.

Furthermore, a stage-discharge relationship has to be as unique as possible. Therefore, e.g. a varying backwater curve from downstream effects has to be avoided. This might be the case when downstream of the gauging station a tributary with variable flow confluents with the main river. In this respect a control section upstream of a waterfall or rapids, where critical flow occurs, is preferred.

Ideally, the site is selected at a location where the flow is uniform. A guideline is to select a 100m straight reach, with little turbulence, in such a way that the stage is representative for the entire section. This is in contrast, e.g., to a section where the channel bends. In a uniform flow section also the discharge can be established with larger accuracy, as the velocity profile will be according to expectations.

Composition of Rating curves

A rating curve gives the relation between discharges and gauge readings (also referred to as stages or water level readings. See next section on how to measure stage) in a certain cross section of a river at a fixed geographical location (gauging station). The rating curve can be approximated by the formula:

$$Q = a \cdot (h - h_0)^b \tag{2.26}$$

where, Q = discharge in m^3/s
h = stage reading in m
h_0 = stage reading at zero flow
a,b = coefficients

For a channel control this equation is compatible with the Chézy formula, where the cross sectional area A and the hydraulic radius R are functions of $(h - h_0)$. Assuming a wide river, where approximately $A = B \cdot (h - h_0)$ and $R = (h - h_0)$ applies to, it can be shown that:

$$Q = C \cdot B \cdot (h - h_0)^{1.5} \cdot i_b^{0.5} \tag{2.27}$$

where,
i_b = slope of the energy level
C = the Chézy roughness

The coefficient b has a value of 1.59 in a rectangular channel, a value of 1.69 in a trapezoidal channel with side slopes 1;1 and a value of 2.67 in a triangular channel. When the coefficients a and b are fixed, plotting Q against $(h - h_0)$ produces a straight line on double logarithmic paper, as can be seen from the transformed equation of the rating curve:

$$\log(Q) = \log(a) + b \cdot \log(h - h_0) \tag{2.28}$$

In reality a cross section of a river bed is a composite of sections. Consequently, a rating curve on double logarithmic paper also can be a composite of several straight lines, each with its own values for a and b. Often one distinguishes between conditions under normal flow and bank full flow. In rivers with movable beds, h_0 fluctuates considerably and therefore should be updated regularly, preferably after each significant flood.

Applying linear regression is a suitable way to obtain the coefficients a and b. Then the logarithmic values of the discharge and the corresponding observed water depth above zero flow ($h - h_0$) are plotted. A first estimate of the stage at zero flow, h_0, can be obtained by plotting non-transformed values of discharge Q against stage h. It is possible to try to get a better fit by slightly changing h_0. The best fit will have the highest correlation coefficient.

Measuring of stage

The stage of a river is the height of the water surface above an established reference plane. Stages are relatively easy measurable, even though the requirements for the accuracy of the measurement can differ for each different purpose. The highest accuracy is required when the slope of the water surface for a relatively short reach of the river needs to be determined. Stages are frequently used to determine stage-discharge curves or to estimate a discharge at a particular moment with the help of a discharge-curve. In the Netherlands, the ISO-norm 4373-1979 (see ISO, 1983, pp. 426-443) deals with water level measurements. (See also Jansen (1979, p. 174 and further))

Selection of Location

Before starting with the actual measurements a suitable location has to be selected. The choice of the location depends on:

- Purpose of the measurement
- Hydrological and hydraulic properties of the location
- Costs of installation and sustenance
- Accessibility and distance to public road
- Availability of mains voltage

Besides these points of decision the watercourse, at the point of measurement, has to fulfill certain requirements for the usefulness of the measurement. The requirements for the watercourse are:

- The watercourse has to be straight for at least 100m upstream and downstream from the point of measurement for a regular velocity distribution in the water
- No inundation (flooding) when extreme high water occurs (1 time per 50 years)
- No backwater
- Close to the point of measurement, measurements of velocity have to be possible so stage-discharge curves can be compiled
- No big changes in resistance (think of plant growth or lack of such in wintertime, this for instance occurs with discharge measurement stations in the Swiss Rhine).

Stage equipment

With measurements of stage one should differentiate between direct and indirect measuring systems. Direct systems measure the stage directly, while indirect systems measure not the stage, but another parameter (e.g., water pressure) which is being converted into a stage.

Available systems are:

Figure 2.23: a) vertical staff gauge; b) inclined staff gauge

- Staff gauges
- Ultrasonic water level gauges
- Electronic pressure transducers

Staff gauges:

The easiest way to measure a stage is with a staff gauge, normally either a vertical staff gauge (see Figure 2.23a) or an inclined staff gauge (see Figure 2.23b). The latter is less vulnerable for ships and floating ice. The zero line of the scale has to be fixed to a reference plane by means of a leveling instrument. The smallest subdivision of the scale is 10 mm. The reading accuracy is 3 mm.

Advantages of staff gauges:

- Easily installed
- Direct reading

Disadvantages of staff gauges:

- Difficult reading when waves occur
- Not suitable for automatic data processing
- Chance of inaccurate reading. Staff gauges usually show only intervals of decimetres and centimetres, this has often showed to be a recipe for reading the improper meter indication.

Ultrasonic water level gauges:

The ultrasonic water level gauge has a sensor located at either a fixed point above the water surface or a fixed point below the water surface (see Figure 2.24). This sensor sends, with high frequency, sonic waves to the water surface, the sonic waves are reflected by the water surface and received again by the sensor. The time interval between transmission and reception is measured, from knowledge of the velocity of sound in air or water, as the case may be, this can be converted into a stage.

The main advantage of the ultrasonic water level gauge with a sensor located at a fixed point

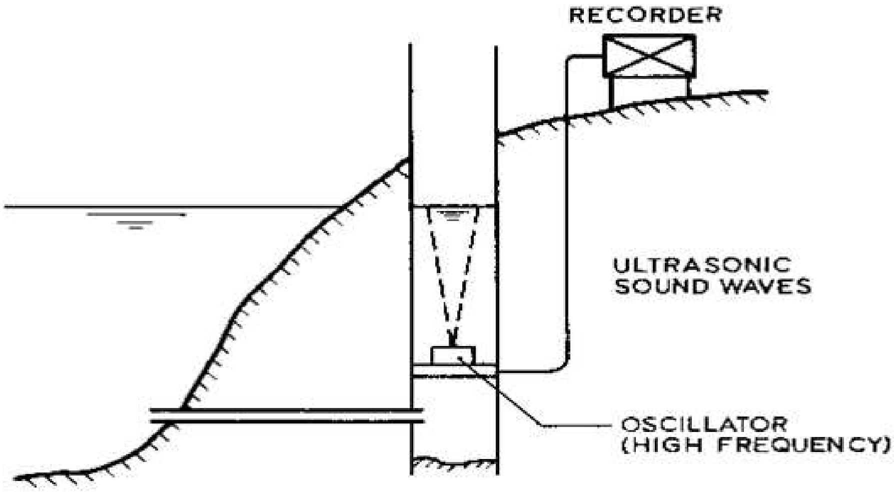


Figure 2.24: Ultrasonic water level gauge with sensor below the water surface

Figure 2.25: Sealed pressure transducer (Van Essen Diver)

above the water surface is that the sensor does not have contact with the water. This is especially useful for measurements in polluted water; the ultrasonic water level gauge is for this reason often used by purification plants.

The velocity of sound in air or water changes with density. Hence temperature and humidity affect the accuracy of the method. Most systems correct their measurements automatically for temperature and humidity, but this is only a partial correction. The accuracy of this sensor is about the same as the accuracy of the diaphragm-type pneumatic gauge. The ultrasonic water level gauge is used mainly within a stilling well (for stilling well see next section).

Electronic pressure transducers:
An electrical pressure transducer has two main components: the force summing device, which responds to the pressure (caused by the water head above the device), and the sensor which converts the output of the force summing device into an electrical signal.

The transducer may contain signal processing electronics, which change the low-level sensor output into a form suitable for transmission over long distances. In all transducers, water pressure + atmospheric pressure are measured with respect to a reference. This reference gives the transducer a classification:

1. Vented transducer - reference to atmospheric pressure
2. Sealed transducer - reference to a fixed known pressure

Since all open channel flows are subject to atmospheric pressure, the vented gauge type is the most suitable for application to measurements of stage. The vented gauge incorporates a vent pipe. Disadvantage is that for instance condensation water in the vent pipe, can cause the transducer to become inaccurate. Sealed transducers need additional measurement of atmospheric pressure. In Section 4.1 further details on the pressure sensor are given.

For all measuring systems a regular check of equipment on the measuring site is necessary. Suspicions about correctness of the data are never superfluous. Regular calibration (in the field) is highly recommend.

Stilling well and intake
Measuring systems can be used within a stilling well. A stilling well is used to damp out natural oscillations in the water surface, waves and turbulence among other things. The function of the stilling well is to provide within the well an accurate representation of the mean water level. The stilling well is in contact with the stream via the intake. The function of the intake is to allow water to enter or leave the stilling well so that the water in the well is maintained at the same elevation as that in the stream under all conditions of flow.

The system has to be designed in such a way that significant variations in stage also occur within the stilling well.

Structures

On small rivers it is often convenient to measure flow by means of a weir or flume. Such structures have the advantage that they are less sensitive to the downstream conditions, the channel roughness and the influence of backwater than the velocity-area method for example.

The philosophy of the method is founded on the premise that the relation of discharge to water level is found empirically or is based on physical principles.

The water level, or head, is measured at a prescribed distance upstream of the structure. For the simple, and usual case, where the downstream water level is below some limiting condition and where it does not affect the upstream head, there is a unique relation between head and discharge. This condition is termed the free-flow or modular condition. If, however, the tail water level affects the flow, the weir is said to be drowned, or submerged, and operates in the non-modular condition. For this condition an additional downstream measurement of head is required and a reduction factor must be applied to the modular or free-flow discharge equation. When the flow in the non-modular condition increases until the weir is almost or wholly submerged, the structure no longer performs as a measuring device.

There is a significant degradation in the accuracy of discharge measurements in the non-modular flow condition. This degradation is due mainly to the difficulty in measuring the downstream head because of turbulence and the uncertainty in the coefficient of discharge in the non-modular range.

In order to increase the range of a structure in the free-flow condition, it is sometimes convenient to raise the design height; otherwise for a weir the device may be used as a section control in the non-modular condition, and rated by a current meter in this range.

The installation of a measuring structure in a river requires major civil engineering work and each device needs to be designed as such with due attention to foundations. Ground conditions may require sheet piling and cut-offs to prevent the seepage of water under or around the structure or risks to stability. If provision for energy dissipation downstream is not included in the construction, for example, there may be a tendency to scour depending on the ground conditions. If this becomes excessive the stability of the structure will be in danger. If the structure, however, is founded on rock, for example, such problems may not be severe but nevertheless due attention may still be required with regard to energy dissipation. The energy downstream of a structure is normally dissipated by a hydraulic jump and generally design and construction considerations involve the installation of a stilling basin and a suitable elevation for the floor of the basin.

The capital cost of installing a measuring structure is much greater than the capital cost of installing a velocity-area station but the operational costs of a structure are very much lower. This is due to the calibration costs in manpower in establishing the stage-discharge relation for a velocity-area station. The advantage, in this respect, of a standard measuring structure is that records of discharge are obtained immediately on completion of the installation.

A weir or a flume is basically a section control designed and built to specific criteria. The rating equation for a stage discharge relation and measuring structure, therefore, take the same form, which is:

$$Q = Ch^n \tag{2.29}$$

Ideally the flow conditions upstream of a measuring structure are governed by the geometry of the structure and the approach channel, and by the physical properties of the water. They are not affected by the flow conditions in the channel downstream or by the roughness and geometry of the channel upstream.

At the control section critical flow conditions occur, where for a given discharge the depth is such that the total head is a minimum. The total head is the total energy of the flow per unit weight of water and, by Bernoulli's theorem; this is the sum of the potential head, the pressure head and the velocity head. It is generally referred to the crest of the structure as a datum.

In presenting the theoretical analysis for the establishment of the discharge equation the broad-crested weir is used (see Figure 2.26). Taking the critical section to occur where the flow is parallel near the end of the crest of the weir, then the Bernoulli equation gives:

$$H = d + \frac{u^2}{2g} \tag{2.30}$$

With:

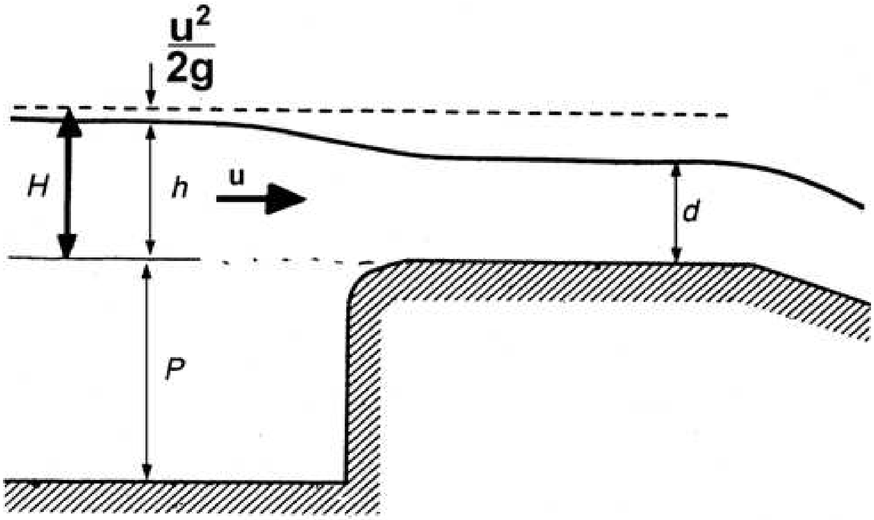


Figure 2.26: Schematic illustration of principle and theory of flow over broad crested weir

H	the total head
g	the acceleration due to gravity
u	the velocity on the streamline
d	depth of flow over the crest
h	depth of flow in front of the weir

If the approach channel is deep, the total head H is the same for all streamlines and it follows from equation 2.30 that the velocity at the critical section is constant with depth and equal to $\overline{u}$, the mean velocity. The discharge over a width b is:

$$Q = bd\overline{u} \tag{2.31}$$

and substituting from equation 2.30

$$Q = bd\sqrt{2g\,(H-d)} \tag{2.32}$$

Differentiating equation 2.32, treating H as a constant and putting dq/dd=0 gives the following equation for critical flow conditions

$$d_c = \frac{2}{3}H \tag{2.33}$$

or

$$Q = \left(\frac{2}{3}\right)^{2/3} b\sqrt{g}H^{3/2} \tag{2.34}$$

Equation 2.34 is derived from theoretical concepts and calibration of a structure imposes on the equation a coefficient of discharge. The equation than becomes:

$$Q = \left(\frac{2}{3}\right)^{2/3} C_d b \sqrt{g} H^{3/2} \tag{2.35}$$

Since the total head H cannot be measured in practice, an iterative procedure is necessary to compute the discharge from equation 2.35. In order to avoid this cumbersome procedure the discharge equation can be presented as

$$Q = \left(\frac{2}{3}\right)^{2/3} C_d C_v b \sqrt{g} H^{3/2} \tag{2.36}$$

This is the basic equation for measuring structures with:
C_v the dimensionless coefficient for the velocity of approach
C_d the coefficient of discharge

From an inspection of equations 2.35 and 2.36 it can be seen that:

$$C_v = \left(\frac{H}{h}\right)^{2/3} \tag{2.37}$$

The Bernouilli equation relates H and h as follows:

$$H = h + \frac{\overline{u}^2}{2g} = h + \frac{Q^2}{2gA^2} \tag{2.38}$$

The energy head H and measured water level in front of the structure will be practically equal if $\overline{u}$ is negligible. For a large range of discharges this will be the case if A is large. This is the reason that a stilling basin of water upstream of the structures often is part of installation requirements.

With each of the discharge equations for measuring structures, limitations are imposed as to their use in practice. Such limitations concern in range in head under which a laboratory calibration was made, the distance upstream to the head measurement section, the limits on certain ratios such as h/P, where P is the height of the weir, and others which will be presented later. The values and uncertainties of the coefficients of discharge and velocity are dependent on maintaining these limitations in the field installation. If the limitations are relaxed for any reason the measuring structure may still produce a record of streamflow but the uncertainty of the measurement will be impaired. In most, if not all, of these cases no data are available to give values of uncertainties in the coefficients, and therefore in the discharge measurements where the specified limitations have been relaxed. The only resort in these circumstances is to field calibration.

Chapter 3

Rivers and Deltas

Handouts will be provided during the lectures and are available on Blackboard.

Chapter 4

Water in the city

4.1 Urban drainage

by J.A.E. ten Veldhuis

4.1.1 Introduction

What is urban drainage? Drainage systems are needed in developed urban areas because of the interaction between human activity and the natural water cycle. Human activity leads to the production of wastewater that needs to be drained properly to prevent health risks and environmental pollution. Concentration of human activity in urban areas also leads to the covering of land with impermeable surfaces. Rainwater that falls one these surfaces needs to be drained to prevent flooding, damage and inconvenience. Thus, urban drainage systems handle two types of water: wastewater and stormwater. The systems have two major interfaces: on the input side with water use by citizens and characteristics of runoff areas and on the output side with the environment, especially surface water and groundwater (Figure 4.1).

Urban drainage systems consist of a network of gutters, underground pipes interconnected by manholes, open channels, pumping stations and special structures like overflow weirs. This section focuses on the underground system of pipes and manholes and what special conditions apply for measurement in the underground system. This system will be referred to as the sewer system; measurements at overflow weirs apply to combined sewer systems, while measurements at other locations in sewer systems may apply to combined systems, separate storm water systems and separate wastewater systems.

Figure 4.1: Interfaces between urban drainage system and urban environment

Good management of urban drainage systems (including sewer systems) is founded on reliable information on expected inflows into urban drainage systems and on outflows: occurrence of flooding, overflow frequencies, overflow volumes, loads of pollutants and infiltration flows.

The focus of this section is on water quantity measurements: measurement of water levels and discharge at overflows to surface water and in sewer pipes.

Information on urban drainage systems: models and measuring programs Information to design urban drainage system and study their behaviour can be obtained in two ways: by modeling and measurement. Traditionally, urban drainage systems have been studied by developing hydrodynamic models that simulate system behaviour. The introduction of measurements in the field of urban drainage is relatively recent: since the 1990's research programs have been established at pilot scale for research purposes and only in the last decade have measurement programs found their way to wide-spread application in practice. As a result, many existing programs suffer from child's diseases and will have to be improved as practical experience is rapidly growing.

Objectives of measuring in urban drainage systems In general, the following objectives apply to measuring in urban drainage systems:

- Check compliance with regulations, e.g. maximum yearly overflow volume
- Check achievement of objectives set in policy documents, e.g. reduction of the number of flooded locations
- Check observed deficiencies in system behaviour, e.g. odour complaints, illicit connections
- Calibrate and verify hydrodynamic models
- Increase understanding of flow processes and system behaviour

Measurement parameters and locations The objective of a measurement programs determines the parameters that need to be measured. These can include:

- Water levels: to quantify overflow frequencies, flooding frequencies, to quantify overflow volumes at overflow weirs, to calibrate hydrodynamic models
- Discharge: to quantify overflow volumes, to calibrate models
- Temperature: to identify illicit connections
- Turbidity: to study erosion and sedimentation processes, to study relations between turbidity and pollution parameters
- Oxygen content: to study odour complaints, to study effects of organic pollution loads
- Suspended solids: to quantify overflow pollution loads, to study sedimentation processes
- Other water quality parameters: to quantify overflow pollution loads, in-sewer chemical and biological processes

In this course we only focuss on water level and discharge measurements. More details on the other measurements are given in CE4440 'Hydrological Measurements'.

The selection of measurement locations for underground systems is limited by difficulties of accessibility. Locations with relatively easy access are overflow locations, manholes and pumping stations and most measurement programs are confined to a selection of these locations .

4.1.2 Water level measurements in sewer systems

The following sections provide a description of available techniques for water level (stage) measurements: ultrasonic water level gauges and electronic pressure transducers.

Pressure sensors (see also Section 2.5.3)

Pressure sensors are based on the principle of hydrostatic pressure: this allows direct quantification of the water depth based on a pressure measurement (eq. 4.1).

$$h = \frac{\Delta p}{g\left(\rho_w - \rho_a\right)} - \frac{\left(\Delta p\right)^2}{2gK\left(\rho_w - \rho_a\right)} \tag{4.1}$$

With:

h	water depth (m)
g	gravitation acceleration (m/s^2)
K	Compression modulus (N/m^2)
ρ_w	density of water (kg/m^3)
ρ_a	density of air (kg/m^3)
Δp	difference between water pressure and atmospheric pressure (N/m^2)

The pressure that is measured by the sensor is the total pressure that consists of hydrostatic pressure and air pressure. Air pressure may vary between approximately 960 hPa and 1040 hPa, equivalent to 80 cm of hydrostatic pressure. Therefore, both air pressure and hydrostatic pressure need to be measured to calculate the correct pressure difference and derive the water level h. Most sensors have a small vent tube for this purpose.

The second term of equation 1 compensates for compressibility of water; this term is negligible for measurements at relatively shallow water depths (1-10 m), as in urban drainage systems. The density of wastewater varies between approximately 1000 and 1005 kg/m^3, depending on suspended solids content and dissolved gases. When selecting a measurement location, it is important to verify that no air entrainment (e.g. at internal weirs) takes place in the vicinity of the sensor.

Pressure sensor is a robust type of sensor that can reach an accuracy of 0.1% to 0.6% under practical conditions (Clemens [2001]). The accuracy of the pressure sensor is mainly determined by the quality of the membrane. Possible membrane materials are ceramic, metal and synthetic materials; ceramic membranes are most robust than others. The main disadvantage of application of pressure sensors in wastewater systems is that they need to be submerged: both the sensor and the vent tube are therefore susceptible to pollution and damage by floating obstacles. Pollution and moisture condensation in the vent tube can cause zero point drift, which leads to deviation of the measurement signal. Ageing of the membrane can also cause signal deviation with time. Pressure sensors should be maintained at regular intervals to prevent pollution and to check and correct zero point drift.

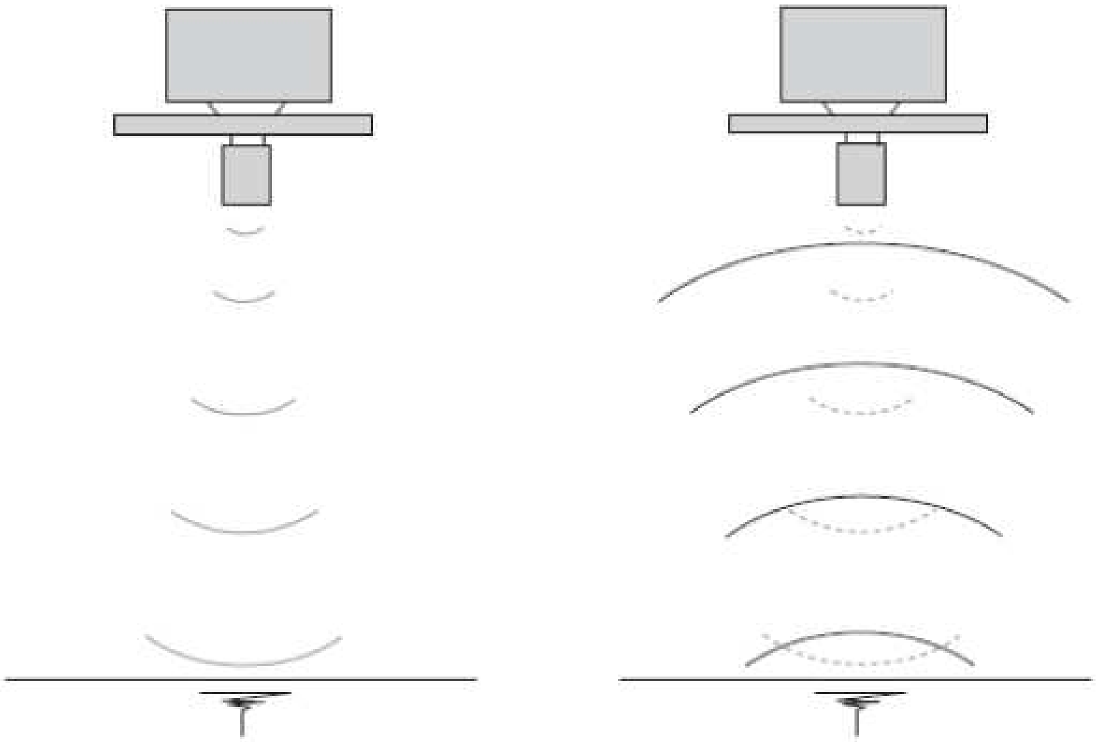


Figure 4.2: Principe of ultrasonic measurement: signal sending (left) and receiving (right) (RIONED [2009])

Ultrasonic sensors (see also Section 2.5.3)

Ultrasonic sensors are based on the principle of sound reflection. The accuracy of this measurement is determined by the accuracy of the time of measurement and the value for the speed of sound.

Ultrasonic sensors can be applied above or below the water surface. Application above Installation of the sensors above the water surface has the advantage that avoids contact of the sensors with the (waste)water flow is avoided . Application under submerged conditions is not suitable for wastewater flows, because air bubbles and sound absorbents weaken the signal. Ultrasonic sensors require a minimum distance of 25 cm to 50 cm free of obstacles between the water surface and the sensor. Combined sewers and stormwater sewers get surcharged (pipes are fully filled) during heavy rainfall and the pressure level may even rise to ground level. This implies that sensors located in manholes may submerge and no longer provide a reliable measurement.

Under practical conditions, especially in sewer systems, the sound reflection principle is not as accurate as the pressure measurement:

- Turbulence, air bubbles, floating objects, layers of foam or fat can disturb the measurement by absorption of the sound pulse or by influencing the reflection angle.
- Sound reflection may be influenced by the geometry of the measurement location, normally a sewer manhole, which leads to inaccuracy.

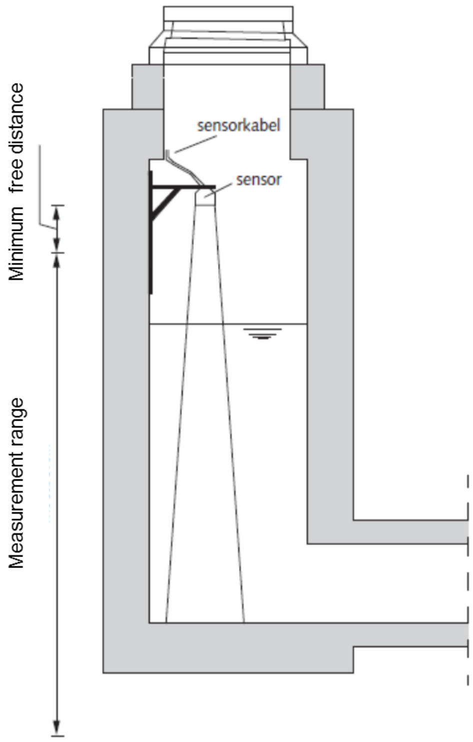


Figure 4.3: Installation of ultrasonic sensor in a sewer manhole. The measurement range covers only about 80% of the total possible water level variation. (RIONED [2009])

Applications of water level measurements in sewer systems

Water levels are measured in sewer systems to quantify overflow frequencies, flooding frequencies, to quantify overflow volumes at overflow weirs and to calibrate hydrodynamic models. Overflow frequencies were measured in the past to check compliance with environmental regulations that were expressed in terms of a maximum yearly overflow frequencies. Overflow frequencies need to be measured as close as possible to the overflow weir crest, to detect every single overflow event . If sensors are located further upstream, e.g. in the next manhole 100 meters or more upstream, the water level measurement is no longer representative of the water level near the overflow crest and cannot be used to detect overflow events. Measurements were usually performed in a small measurement range, near the weir crest. The disadvantage of this type of measurement is that it provides no information on overflow circumstances: the processes of system filling (only the last 10-20cm of filling are detected) and whether the surface water level at the other side of the weir influenced the overflow (submerged weir). Present application of water level measurements at overflow weirs usually aim to measure a larger range of water level variations for better understanding of flow processes near the overflow weir and to be able to quantify overflow volumes. A change in environmental regulations has been an important incentive for this modification in measurement purpose: regulations are now based on maximum yearly overflow volumes and not just on overflow frequencies. This implies that a sensor needs to measure not only if an overflow event occurs, but also how the water level varies during the event so the overflow volume can be derived from the measurement. For application of water level measurements for quantifica-

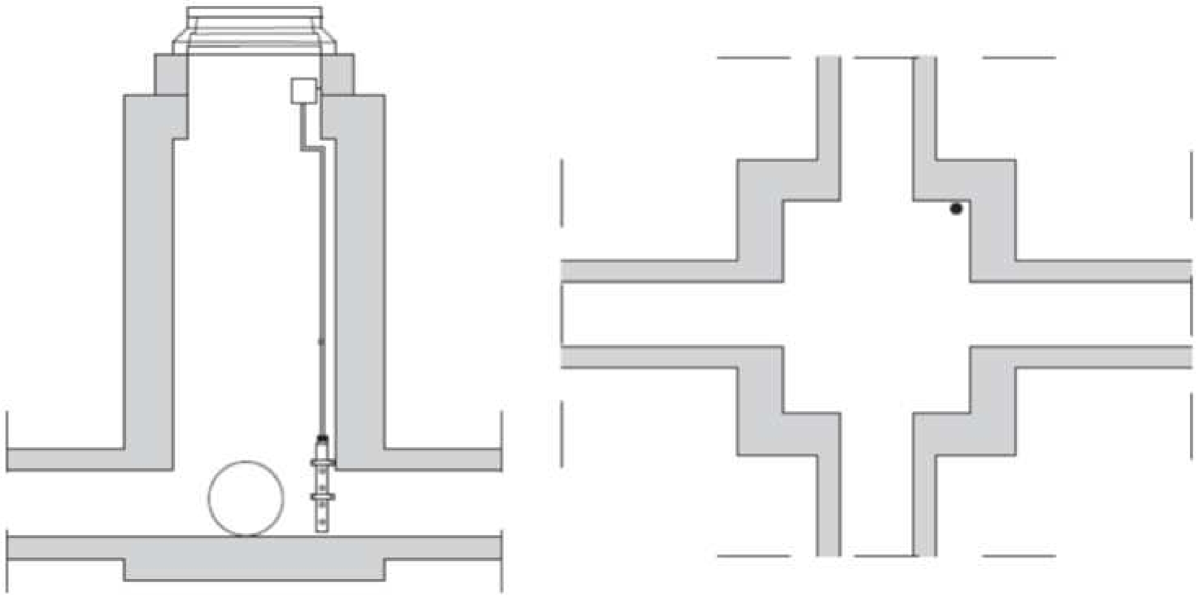


Figure 4.4: Position of pressure sensor in sewer manhole

tion of overflow volumes, see the chapter on discharge measurements in sewer systems. Water level measurements at overflow locations can help to obtain rudimentary understanding of flow processes and system behaviour. Additional measurements are applied if better understanding of processes is needed, for instance to explain unexpected flooding or odour complaints and for calibration and verification of hydrodynamic models. In those cases, water level measurements are applied at several critical locations in the sewer system; the selection of those locations is based on expert knowledge of system functioning and on practical considerations such as location accessibility. In addition to water level measurements, measurements at pumping stations and discharge measurements can be applied.

Practical considerations for installation of water level sensors

Water level sensors are usually installed in sewer manholes, because these are relatively accessible compared to other locations in sewer systems. Pressure sensors are installed near the bottom of a manhole (figure 4.4), ultrasonic sensors near at some distance below the manhole cover (figure 4.3.

Sensors that are applied in combined sewer systems and wastewater systems need to be robust: they need to be able to endure aggressive environmental conditions (wastewater may produce sulphuric and other gases), moisture, pollution by sediments and fat and large variations in flow velocity (in combined systems).

4.1.3 Discharge measurements in sewer systems

In sewer systems, flow and discharge measurements are applied at two types of locations:

- Over special constructions, especially overflow weirs and internal weirs.

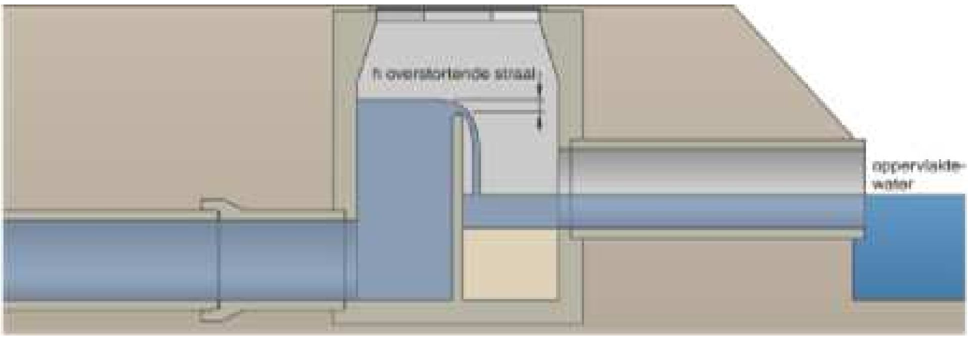


Figure 4.5: Principle of overflow weir constructed inside a sewer manhole. Water flows from the sewer system (left) over the weir towards surface water (right). (Courtesy of VPB)

- Inside conduits (= pipes); either full pipes or partially filled pipes.

Flow measurement at weirs

The principle of stage-discharge measurement is explained in chapter 2.5.3. The discharge is derived from a water level measurement above a weir, based on equation 4.2, which is repeated here:

$$Q = Ch^n \tag{4.2}$$

With:

With:	Q	Discharge (m^3/s)
	C	empirical coefficient (m^2)
	h	water depth upstream of weir (where velocity is near zero) (m)
	n	empirical coefficient (-)

Weirs in sewer systems usually take the form of a brick wall of a metal plate that is constructed inside a manhole or at the end of an outflow pipe and can be seen as sharp-crested weirs. Weirs in sewer systems are seldom designed and constructed especially for discharge measurement and flow conditions can be far from those required for accurate measurement at sharp-crested weirs.

The discharge coefficient C is well-defined for special, standardised weir structures. As figures 4.7 and 4.8 show, the geometry of weirs under practical conditions deviates from standardised forms in all sorts of ways. For weir structures in sewers, local calibration is required to find the appropriate value of the weir coefficient. The accuracy of discharge measurements at weirs depends significantly on the conditions of the weir installation, the condition of the weir crest and the accuracy of the water level measurement. The water level sensor must be installed at some distance (at least three times the water depth at weir) upstream of the weir to measure the correct water depth: at smaller distance from the weir the velocity increases and so does the velocity head. The water level in that case deviates strongly from the energy head that should be used in the weir formula.

Additionally, flow over a weir may become influenced by the downstream water level. This may occur in flat areas where differences between ground level and surface water levels are small,

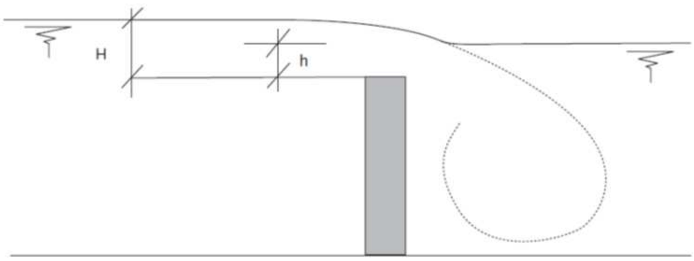


Figure 4.6: Flow over a drowned weir: the discharge over the weir is influenced by the downstream water level, which negatively affects the accuracy of discharge measurement over the weir.

Figure 4.7: Overflow weir in sewer manhole during construction (Courtesy of Nederbetuwe)

such as polder areas. Especially in small catchments, where surface water levels and sewer water levels are influenced by the same local hydrological conditions, simultaneous rise of both water levels is likely to occur. The weir may become drowned, or submerged (figure 4.6). For this condition an additional downstream measurement of head is required and a reduction factor must be applied to the free-flow discharge equation. The accuracy of discharge measurements for a submerged weir is low, because of the uncertainty in the coefficient of discharge under this condition.

Discharge measurement in full pipes

Electromagnetic flow meter The most commonly used technique for discharge measurements in full pipes is the electromagnetic in-line flowmeter. The principle of electromagnetic induction is based on Faraday's law. According to this law, a conductive liquid that moves through a magnetic field create a voltage. Two field coils at opposite sides of the pipe create a magnetic field (figure 4.9). As water flows through the pipe, an electric voltage is formed that is measured by two electrodes in the pipe wall. The voltage is directly proportional to the flow velocity. The discharge can easily be derived by multiplication with the pipe cross-section.

This measurement principle only works in full pipes; in sewer systems it is mostly applied in

Figure 4.8: Outflow pipe of sewer overflow structure (weir is inside a manhole located at some distance inside the sewer system). The grid is applied to prevent playing children from entering the sewer system; the polluted fence leads to severe obstruction of the sewer outflow, which is likely to influence the overflow weir further upstream.

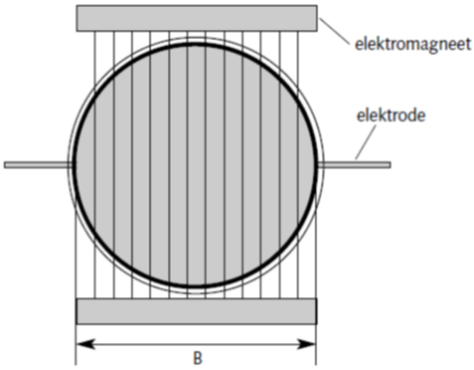


Figure 4.9: Principle of electromagnetic induction in a flow meter

pressured pipelines. A measurement error of no more than 0.5% to 1% can be reached with this type of flowmeters.

For a short explanation of the electromagnetic flowmeter principle, see:
`http://youtu.be/aWzeEMW16xU`
`http://youtu.be/f949gpKdCI4`

Ultrasonic flowmeters Discharge in full pipes can also be measured by ultrasonic flowmeters. This method is based on the relation between the speed of sound in liquid and the flow velocity of the liquid. Pairs of sensors are fitted across from each other in the pipe; the sensors can send alternately send and receive sound signals. The transit time of sound is measured by the sensors. As the liquid flows, the sound signals are accelerated in the direction of the flow. The difference in transit time in the direction of the flow and against the direction of the flow is proportional to the flow velocity. To compensate for possible deviations from a uniform velocity pattern, several sensors are fitted along the pipe wall. The discharge is derived by multiplication with the pipe cross-section.

For a short explanation of the ultrasonic flowmeter principle, see:
`http://youtu.be/Bx2RnrfLkQg`

The advantage of ultrasonic flow meters is that they can be installed by a clamp-on system on the outside of the pipe. The measurement can be disturbed by gas bubbles or sound absorbing particles in the flow. The expected measurement error of this method is 3% to 5%.

Doppler flow meters Doppler flow meters are based on the Doppler effect: the wave length of sound changes as the sound source moves towards or away from the observer. The change in wave length can be related to the velocity of the source. Ultrasonic Doppler flow meters are clamped on the outside of a pipe. A sound transmitter continuously transmits high frequency sound that travels through the pipe wall and through the flowing liquid. The sound is reflected by particles or bubbles in the liquid. Because the liquid and floating particles are in motion, the frequency of the reflected sound signal is altered; the change in frequency is proportional to the flow velocity. The Doppler flow meter continuously measures the frequency shift to calculate the flow velocity. This method only works if they the liquid contains sufficient suspended particles to reflect the signal; in (diluted) wastewater this is usually the case.

For a short explanation of the Doppler flowmeter principle, see:
`http://youtu.be/8VkxbrD6Uik`

The advantage of Doppler flow meters is that they are clamped on the outside of the pipe and have no contact with the flow. The signal can be disturbed by sediment materials on the pipe wall and high concentrations of suspended particles. This method is less accurate than other flow measurements for full pipes applied in sewer systems.

Discharge measurement in part-full pipes

The flow in part-full pipes depends not only on the flow velocity, but also on the filling rate of the pipe. Therefore, discharge measurements in part-full pipes consist of a velocity measurement and water level measurement to determine the filling rate. The flow is derived from these two measurements. Three types of flow measurement are applied in part-full sewer pipes: Doppler radar, ultrasonic Doppler and electromagnetic measurement.

Doppler radar flow meters Doppler radar flow measurements are based on the same principle as ultrasonic Doppler meters. Doppler radar flow meter transmits an electromagnetic signal instead of a sound signal. The electomagnetic signal is reflected by the water surface; the frequency shift caused by the moving water surface is translated into a flow velocity. This method can only be applied in part-full pipes, because a minimum distance between sensor and water surface is required. Doppler flow meters for part-full pipes measure the velocity at the water surface and translate this velocity into an average flow velocity over the entire wet cross-section of the pipe. An independent water level sensor, either a pressure sensor or an ultrasonic sensor, measures the water depth, thus the filling rate of the pipe.

Ultrasonic Doppler Ultrasonic Doppler meters are also applied in part-full pipes. Unlike in full pipes, the ultrasonic flow sensor for part-full pipes is placed on the pipe bottom. The flow velocity is measured in the same way as in a full pipe: an ultrasonic signal is continuously injected into the water. The high frequency sound is reflected back to the sensor from particles or bubbles suspended in the liquid. If the fluid is in motion, the echoes return at an altered frequency proportionate to the flow velocity. The water level in the pipe is measured by an ultrasonic sensor installed at the bottom of the pipe. To measure water level the sensor transmits ultrasonic pulses that travel through the water and reflect off the liquid surface. The instrument precisely measures the time it takes for echoes to return to the sensor. With this technique the instrument can measure flow velocity with an accuracy of $\pm$ 2%, under favourable conditions.

Electromagnetic flow meter Electromagnetic flow meters for part-full pipes differ from those in full-pipes in that the electrodes are situated at the 10% filling level of the pipe in order to be able to measure flow at low filling rates. The flow velocity is derived from the voltage induced by the flowing liquid. The capacitive plates measure are situated along the pipe wall and measure the height of the water level.

Flow measurements in part-full pipes are less accurate than those in full pipes. The main cause of errors is the translation of the measured flow velocity, at the water surface or near the bottom of the pipe, to an average cross-sectional flow velocity. Under practical conditions, distortions of the velocity profile by sedimentation at the pipe bottom, obstacles, pipe bends and other pipe profile changes take place. As a result, the local velocity measurement is no longer representative and the assumed velocity profile that is used to translate the local velocity to cross-sectional

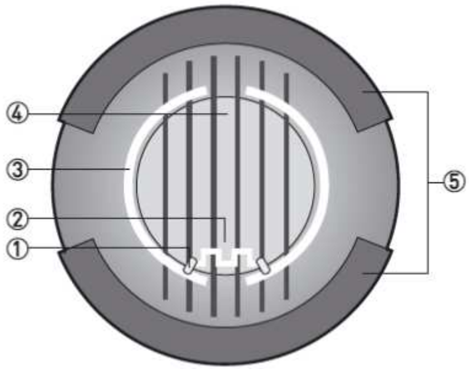


Figure 4.10: Principle of electromagnetic flow meter for part-full pipe: 1- Electrodes, 2- Induced voltage (proportional to flow velocity), 3-Capacitive plates in liner for height measurement, 4-Magnetic field, 5-Field coils

average velocity is no longer valid. The accuracy of discharge measurements in part-full pipe varies between 90% down to only 50%.

4.2 Measuring in drinking water production and distribution

by S.G.J. Heijman

4.2.1 Introduction

Measuring different parameters is very important in drinking water production. Different groups of water quality parameters can be distinguished:

Health related parameters

These parameters are of course the first priority to deliver good drinking water to the customers. **Pathogens** like bacteria and viruses can be the source of water born diseases like diarrhoea, cholera and typhoid. In figure 4.11 the occurrence of typhoid in the Netherlands in the past century is shown it is clear form the graph that wars have a large impact on the occurrence as well as the drinking water distribution. Typhoid is overcome as all people have access to clean and safe drinking water.

Another health related parameter is the number of **micropollutants** in the drinking water. Pollutants like pesticides, fire retardant, hormones, pharmaceuticals can be present in surface water or groundwater. In the Netherlands different treatment technologies are used to remove them form the source. Because of the improved analytical techniques they can be sometimes be detected in drinking water in concentrations of nanogram per liter. Micropollutants are not regarded as an acute hazard (like pathogens are). They can contribute to the occurrence of cancer and of course they do not belong in drinking water.

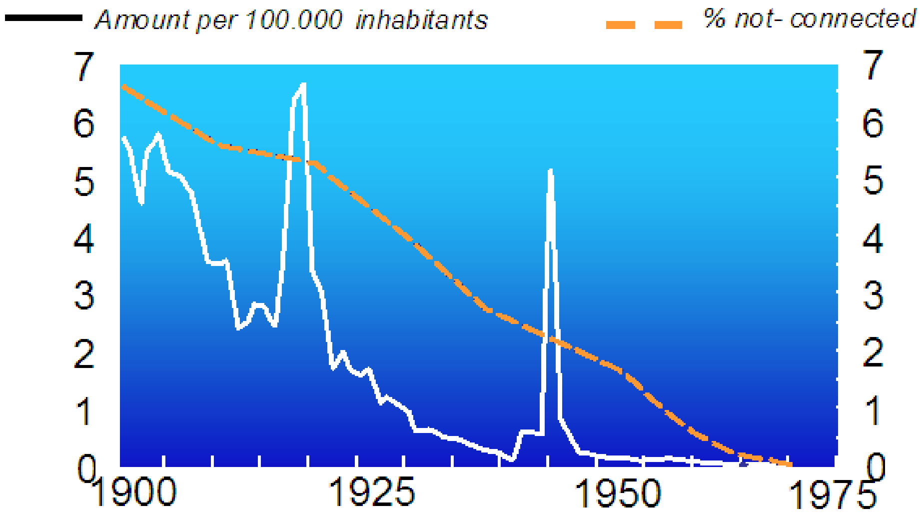


Figure 4.11: Occurence of typhoid in the 20th century in the Netherlands

Figure 4.12: Two sources of micropollutants

	Individual compounds	Concentration
EDC	Oestron	5 ng/l
	Bisphenol A	< 10 µg/l
	Phtalates	3 µg/l
	Alkylphenolethoxylates	2 µg/l
Pharmaceuticals	Ibuprofen	20 ng/l
	Sulfamethoxazole	40 ng/l
	Carbamazapine	90 ng/l
	Iopamidol	70 ng/l
	Clofibric acid	30 ng/l
	Amidotrizoic acid	80 ng/l
Others	NDMA	2 ng/l
	MTBE	3 µg/l

Table 4.1: Unwanted micropollutants in drinking water (maximum concentration occasionally found in the Netherlands)

Inorganic health related parameters are most of the time ions. In too high concentration these ions have a negative effect on human beings. It depends on the ion at what concentration this effect is starting. For normal salt (e.g. NaCl) levels of more than 1000 mg/l can cause kidney problems. In some arid areas people are depending on brackish groundwater with these concentrations of salt. Normal drinking water always contains less than 500 mg/l. Some inorganic contaminates are already toxic at very low levels. An example is arsenic. The guideline of the WHO is 0.01mg/l for arsenic. In India, Bangladesh and Cambodia millions of people are exposed to too high concentrations of arsenic from shallow wells. After drinking this water for more than ten years the skin cancer can develop (see figure 4.13).

Process parameters

These parameters are measured in order to control the drinking water production process and the transport in the distribution network. In water production flow, pressure, temperature, UV-transmission, turbidity, electric conductivity are measured online. In figure 4.14 an impression is given of a modern control room of a big drinking water production plant. The measurements in the process are the eyes and ears of the operators.

Aesthetic parameters

Most of the water quality that effect health cannot be observed by the consumers. Some parameters can be easily observed and are nevertheless important for customers although there is no direct health danger. These parameters are effecting:

Figure 4.13: Arsenic poisoning in Bangladesh

Figure 4.14: Control room of a modern water production plant

- Taste and odour of the water. Components like Geosmin and methylisoborneol can be detected by human beings in very low concentrations and give the water a bad taste or smell. Consumers do not appreciate this water.
- Turbidity. Turbid or cloudy water is associated as not clean water.
- Colour. Too many humic acids in the water can give a yellow colour. Especially in groundwater with from peaty soil the colour can be too high. (figure 4.15)
- Iron. Iron in the water can give a bad taste and can cause staining of laundry.

Figure 4.15: Water with low and water with high concentration of humic acids

4.2.2 Measuring water quality from source to tap

From source to tap is a very important concept in drinking water production. It is important to protect the drinking water sources and prevent contamination with for instance pathogens. It is important to remove the remaining pathogens during water treatment and it is important to have a tight distribution network which is pressurized 24 hours a day to prevent contamination from distribution network. Only if all the chains are monitored sufficient quality can be guarantied.

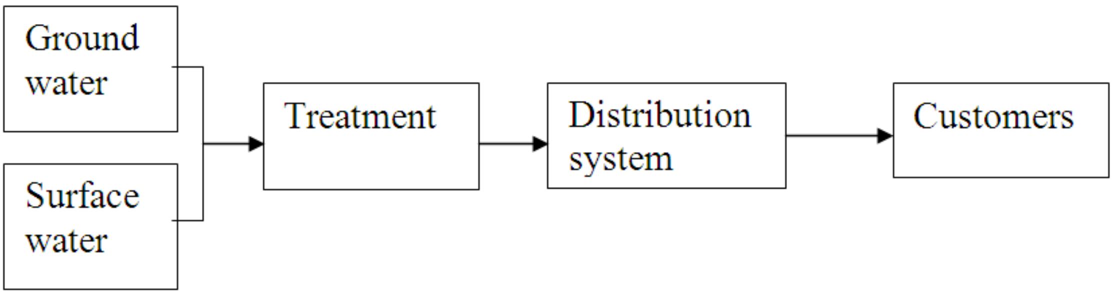


Figure 4.16: Water chain.

Source

Monitoring groundwater is completely different from monitoring surface water. Groundwater is very constant in quality and changes in concentrations of micropollutants develop over years.

Most of the groundwater is free of pathogens because pathogens do not survive in the subsurface for more than a month. The quality of surface water can change within one hour. Contamination can come from continues discharges of industry or agriculture or accidental discharge for instance caused by a fire in a chemical factory (see figure 4.17). In the Netherlands and Germany we organised as a result of this disaster a so called Rhine alarm: The water companies downstream of the river Rhine are warned in an early stage so they can anticipate on the pollution expected in the days to come.

In surface water there is also difference in water quality in summer and winter. Not only the temperature is different also pesticide concentrations are higher in summer because they are sprayed in summer more than in winter.

Figure 4.17: Fire in the Sandoz factory in Basel polluted the river Rhine in 1986

Treatment

In the treatment process parameters are measured to control the different processes. Most filtration processes are monitored by measuring the flow and/or the pressure during filtration. This is the case for rapid sand filtration but also with membrane filtration. If clogging occurs the flow rate is decreasing or the pressure is increasing and the filter can be cleaned in different ways to improve the performance again. In the past the operator was most of the time monitoring the processes and deciding when the filters should be cleaned. Now this is in almost all the drinking water production plants an automated procedure.

Human pathogens are most of the time too low in concentration to measure online. To guard the disinfecting of a certain treatment step sometime surrogate parameters are measured. An example of this is the integrity test of ultrafiltration and microfiltration membranes. The membranes are often used as the main disinfecting step in the treatment process because they are able to remove 99.99% of the bacteria (4 logunits removal). In those cases it is very important to guard the integrity, but pathogens in the permeate are hard to measure because of the low concentration. With ‘particle counters’ the concentration of particles of for instance 2 micron are measured before and after the membranes. Because bacteria are about of the same size, removal of particles is corresponding with removal of bacteria.

At the end of the treatment plant the finished water is sampled frequently and send to a laboratory for analyses. These samples are also necessary to meet the guidelines of the Dutch Drinking Water Directives.

Distribution system

In the Netherlands most of the drinking water is distributed without chlorine (=disinfectant) in it. Despite of the absence of chlorine the number of contamination's of the network is very rare. This is achieved by having the network pressurised for 24 hours a day all days of the year. Also the maintenance of the network is very good and the leakage is very low. A reported incident of E. coli contamination a few years ago was caused by a leakage of the roof of a clean water storage. Rain water contaminated with E.coli of birds could enter the storage. In order to detect such incidents the network is monitored by taking samples in the distribution network. In most samples no E.coli is detected.

A problem in the distribution network is the accumulation of particles. With changing water flow in the pipes these settled particles can be re-suspended and can cause brown or coloured water problems at the water taps. In order to investigate discoloured water in the distribution network monitoring systems (figure 4.18) are installed along pipelines. With turbidity measurements the faith of particles is investigated and the effect of cleaning the pipes is checked.

Figure 4.18: Monitoring system for the distribution network

Customers

Within the customers house at the moment only the water meter is present. This water meter is important because the water meter gives information about the use of water by each household and it gives an indication about water leakage in the distribution system. In the future more monitoring in the houses are expected because there is a fast development in so called 'lab-on-a-chip technology'. With this technology parameters are measured with a cheap (mass production) and small chip. This technology will be available in ten to twenty years.

4.2.3 Some examples of online measurements

Measuring in nanofiltration or reverse osmosis systems

In figure 4.19 a test rig (only one membrane element) is shown. In figure 4.20 the scheme of the same system is shown. The monitoring of a full scale membrane system with 100 or even 1000 elements has about the same setup.

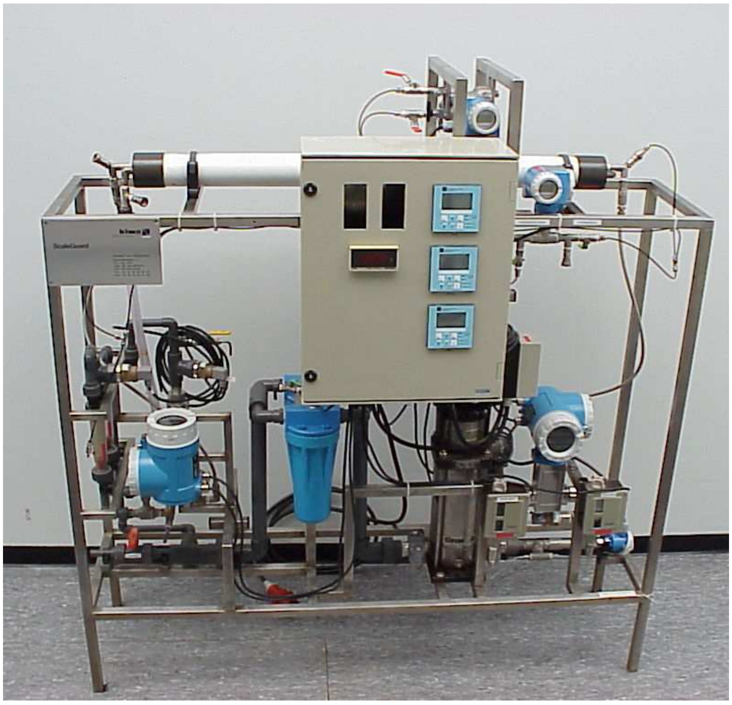


Figure 4.19: Test rig for nanofiltration or reverse osmosis

Measuring/adjusting pH

Often online measurements are used to regulate a dosing in figure 4.21 a pH-measurement and dosing system is shown. Most of the time a feed-back regulation system is used: the pH is measured in the pipeline after a static mixer the dosing is done before the static mixer. If the measured pH is not corresponding with the set point (so the wanted pH) the dosing is adapted until the measured pH is the same as the set point.

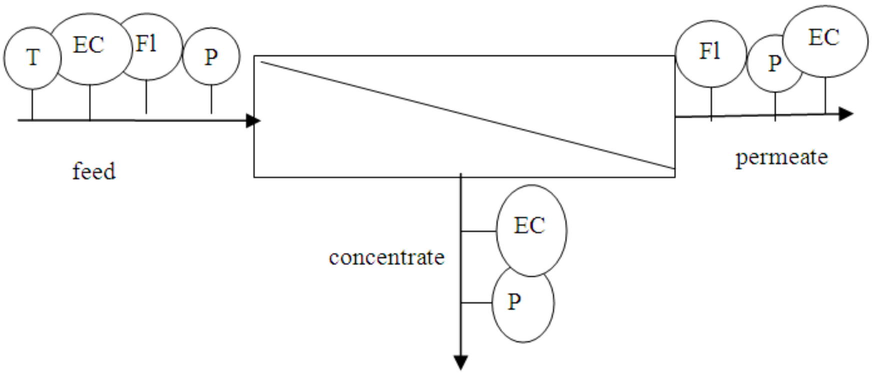


Figure 4.20: Monitoring equipment for the test rig

Figure 4.21: Simple pH-regulation system

Measuring Electric Conductivity

Probably the most simple measurement in water treatment (see figure 4.22). The more ions the lower the resistance of the water and the higher the conductivity.

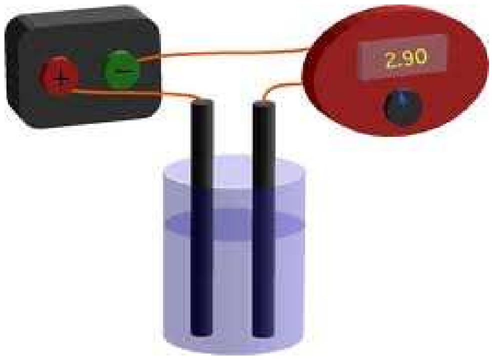
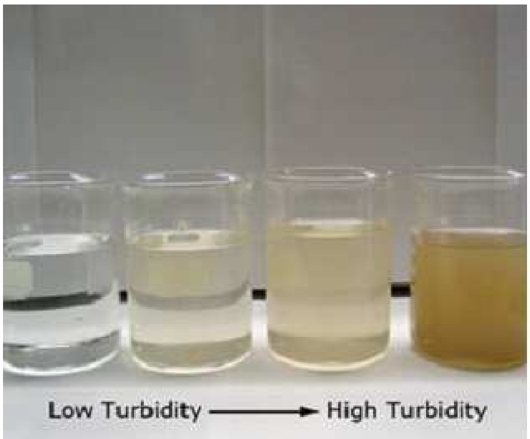


Figure 4.22: Electric conductivity (EC) measurement

Turbidity measurement

Particles scatter light in all direction (see figure 4.23). For the measurement used in drinking water the scattered light is measured at an angle of 90 degrees. The unit is NTU or FNU



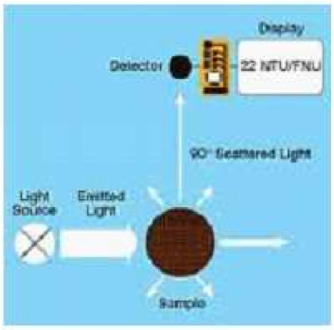


Figure 4.23: Principle of turbidity measurement

Chapter 5

Control Theory

Handouts will be provided during the lectures and are available on Blackboard.

A good reference book is:
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