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Understanding the water balance and its estimation methods

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Abstract

In every ecosystem, the calculation of the water balance in the hydrological cycle requires accurate estimations of the different processes and water components/fluxes. To address this need, this chapter first discusses the concept of the hydrological cycle and water balance at the different spatial scales (from plot to global/continental)). Then, the components/fluxes of the hydrological cycle are analyzed, with a specific focus on atmospheric, surface and sub-surface waters. Finally, an overview of the different approaches to calculate the water balance are presented, including the in-situ measurement methods as well as the hydrological modeling.

Keywords: hydrological cycle; modeling; in-situ measurements; evapotranspiration; infiltration; surface runoff; sub-surface runoff; atmospheric water.

Sub-chapters

1- Introduction to the hydrological cycle

A system can be described as an aggregation of interrelated components subject to regular interactions. The operation of a system is to generate output from input or interrelate input and output. Therefore, a hydrologic system could be explained as a *hydrological system* (hereinafter *hydro-system*), including components of a landscape that store water in its natural state (solid, liquid, gas) that interact regularly according to the physical laws that govern the state, movement and storage of water (Dooge, 1968).

The natural and continuous processes of water movement near or below the earth's surface form the so-called *hydrological cycle* (Figure 1), where water moves either from one location to another or is being transformed from a state (i.e., liquid, solid, gas) to another. This cycle encompasses the three main terrestrial components: water bodies (including oceans), atmosphere, and land (including vegetation). The hydrological cycle starts with the evaporation from the ocean, due to the radiant (heat) energy from the sun (solar radiation). Convection lifts the water evaporated from the ocean to the atmosphere, where, under suitable conditions, the vapor turns into precipitation (water, snow or ice). Precipitation can:

- Be intercepted by the vegetation, and then it directly evaporates back into the atmosphere (*interception*);
- Infiltrate (entering the soil) and evaporates from the soil surface or transpires through vegetation (*evapotranspiration*);
- Turn into an overland water stream (*surface runoff*);
- Migrate in the deeper layers of soil (*infiltration*).

All these *hydrological processes*, which originate from the interactions among the precipitation (on its turn considered as a hydrological process) on one side and land-vegetation, water bodies and atmosphere on the other side (interception, evapotranspiration, infiltration, runoff), are components of the hydrological cycle and vary significantly in both, time and space.

The hydrological processes generate the *water fluxes* (or *water flows*, depending on the object of focus), which are usually defined as:

- *Surface water flux*, which is the volumetric flow of water passing through the land surface;
- *Sub-surface water flux*, which is the volumetric flow per unit of the cross-sectional area of the porous medium (soil) constituting the groundwater storage in the saturated zone;
- *Atmospheric water flux*, which is either in the form of precipitation reaching the land surface from the atmosphere or returned to the atmosphere through the evapotranspiration process.

While the dimension for water flow is [L^3], generally, the water flux is related to the surface unit of the hydro-system, and thus the dimension is [L].

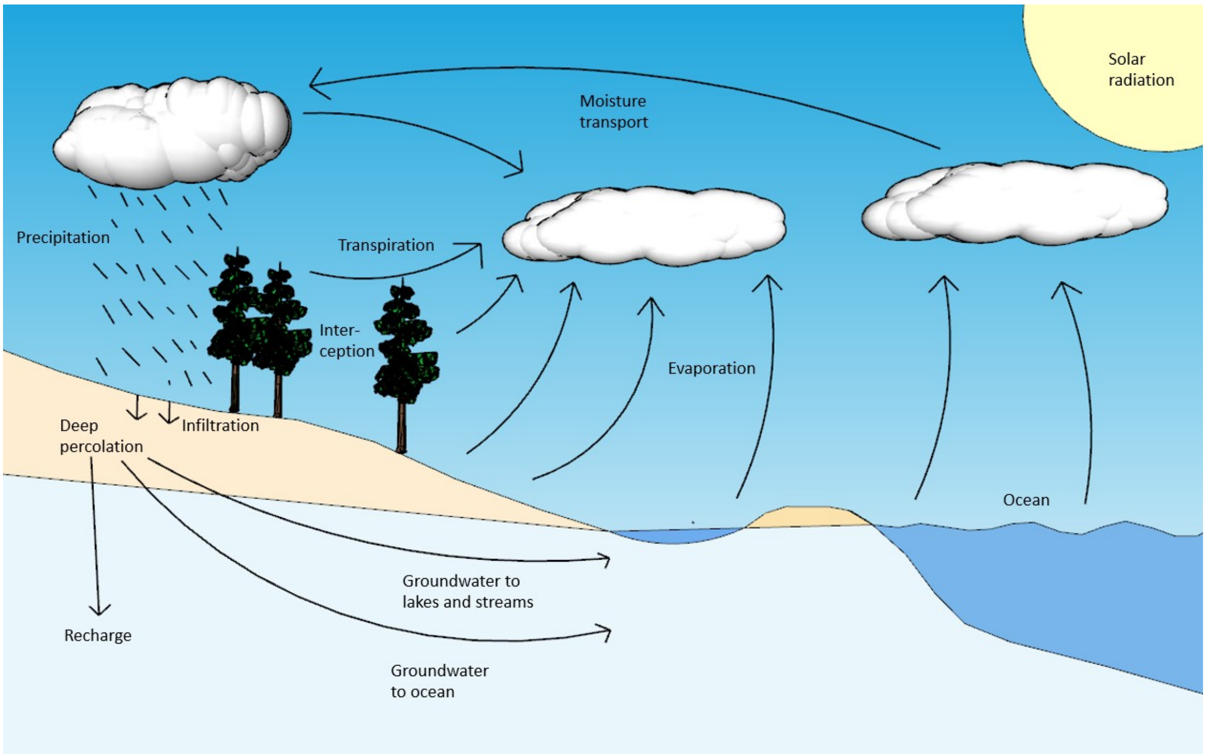


Figure 1. The different processes of the hydrological cycle.

2- The water balance

According to the *law of conservation of mass*, the rate of change in water storage within a hydro-system over any specific period of time must be equal to the difference between the rates of inflow and outflow of water across its boundaries (Byeon, 2014). To describe water flow into and out of a hydro-system (i.e., through the various hydrological processes) and quantify the rate of change in the water being stored in the hydro-system (i.e., in the land, atmosphere, and water bodies), a *water balance* equation is used. The spatial scale of the hydrological system for which the water balance is calculated may range from a small sample of soil (*plot scale*) to an entire catchment (*catchment scale*) or the *global/continental scale*.

To estimate the water balance, it is first necessary to define the *spatial boundaries* of the considered hydro-system (called *control volume*) and the *reference period* of time. The control volume (e.g., a catchment or a soil column) is a volume in space in which fluxes of water, energy and other mass are stored internally or transported across its boundaries. For instance, with regard to a catchment to which the water balance must be applied, the control volume is the space within (i) the ground surface; (ii) the horizontal layer (roof) over the tallest vegetation; and (iii) the vertical lines extruded from the perimeter of the ground/impervious surface and the roof of vegetation. The reference period is the temporal scale (event, monthly, seasonal, annual, decadal or longer) when the changes in the water storage and fluxes of the control volume are estimated. Thus, the total amount of water that is stored in a control volume is the *water storage*, whose dimension is $[L^3]$.

A water balance accounts for the horizontal flow of water through the landscape in watercourses, for the vertical water fluxes among the atmosphere, ground surface, and groundwater and for the changes in water storage within the control volumes. In its general form, the water balance can be expressed using the following equation (Sutcliffe, 2004):

$$P + I = ET + Q + \Delta S + \Delta G + \Delta W \quad (1)$$

where P is precipitation, I the inflow (the water flow entering the control volume), ET the evapotranspiration, Q the outflow (the water flow leaving the control volume, and ΔS, ΔG and ΔW the changes in the water content of soil (in the unsaturated zone), groundwater storage (in the saturated zone), and water amount stored into surface water bodies, respectively. In the following, we discuss the water balance at different spatial scales.

2.1- Plot scale

The water balance at the plot scale (Figure 2) is usually applied for agricultural purposes. It considers the root zone per unit area as the control volume. The difference between the water fluxes entering and leaving the control volume must be equal to the changes in the water content, throughout the reference period. In a nutshell, the water content of the soil volume increases when water is added due to infiltration or capillary rise and decreases when water is lost by evapotranspiration or deep percolation. The water balance equation at plot scale is usually represented as follows (Zhang et al., 2002), where all the values are expressed as water fluxes or equivalent water depth throughout the reference time period:

$$\Delta S = P - I - E - T - Q - DP + CR \tag{2}$$

where ΔS is the change in water content of the root zone, P is the precipitation, I is the interception, E is the direct evaporation from the soil surface, T is the transpiration through vegetation, Q is the runoff (surface runoff and interflow, see section 3.2 and 3.3), DP is the deep percolation towards the groundwater storage, and CR is the capillary rise.

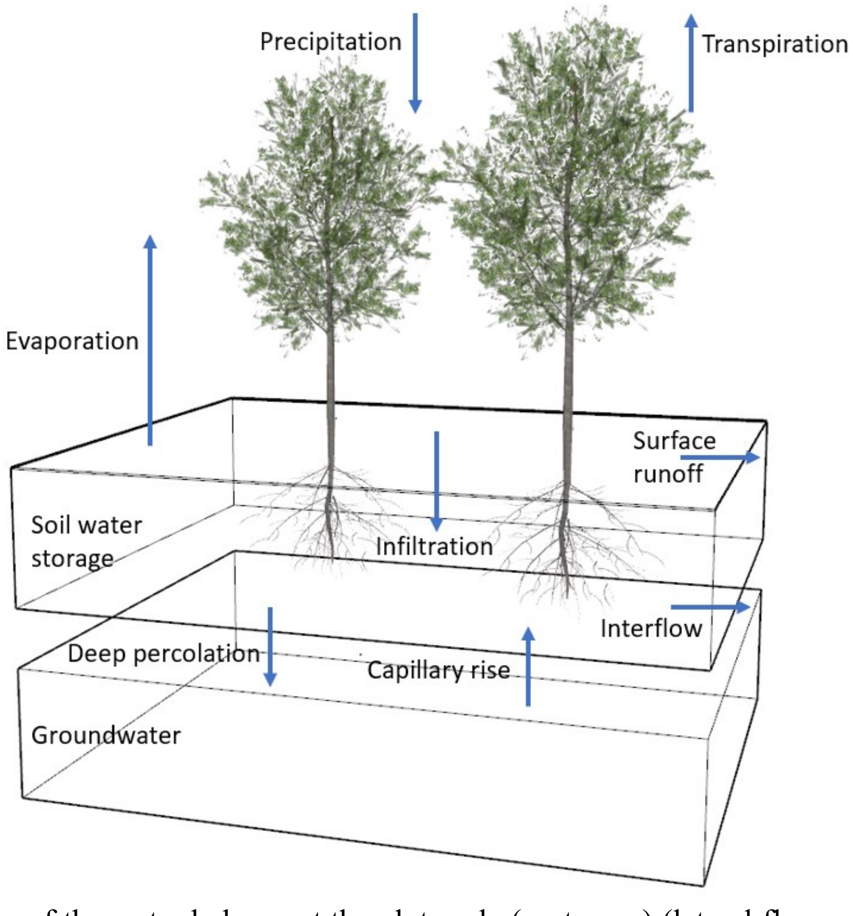


Figure 2. Scheme of the water balance at the plot scale (root zone) (lateral flow = interflow + groundwater flow).

2.2- Catchment scale

A catchment, divided by watershed from the adjacent system, is the geographical unit of interest for terrestrial hydrology to apply the water balance and part of the extensive water cycle. Precipitation is caught up by catchment, and the drainage network of the catchment collects and conveys part of this water to a common outlet. The catchment outlet can be the mouth of the main watercourse into the sea, the confluence into another stream, or the section where it flows into a lake, reservoir or wetland (Figure 3). By definition, when applying the water balance at the catchment scale (i.e., when the catchment is the control volume), the streamflow at the catchment outlet represents the integrated response to all hydrological processes within the catchment (Kirchner, 2009; Singh and Woolhiser, 2002). At the catchment scale, the water balance equation can be expressed as follows (Zhang et al., 2001), where all water fluxes are estimated as the catchment-scale spatially average:

$$\Delta S + \Delta G + \Delta W = P - ET - Q \qquad (3)$$

where ΔS, ΔG and ΔW are the changes in water content of the soil, groundwater and surface water bodies, P is the precipitation, ET is the evapotranspiration, Q is the runoff (streamflow).

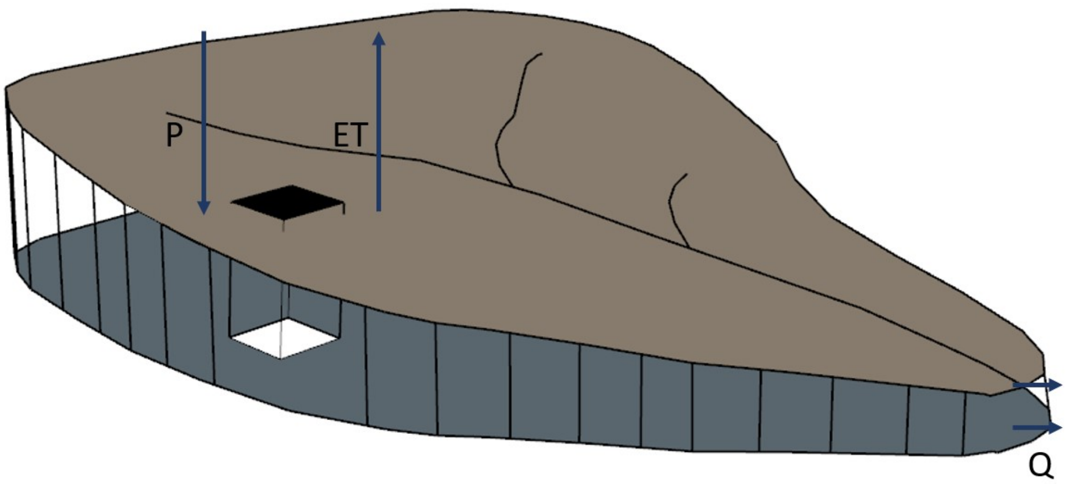


Figure 3. Scheme of a catchment (control volume); *ET* = evapotranspiration; *P* = precipitation; *Q* = runoff; the brown area is the soil surface, while the gray area is the sub-soil; the vertical box is the soil column.

2.3- Global/continental scale

At global or continental scales, the water balance is generally accounted for the fundamental components, i.e., precipitation on the land surface is balanced out by streamflow, evapotranspiration, and the change in water storage (Güntner, 2008). Since the water quantity of the oceans is considered to be constant for long periods of time, the streamflow (quantity of water returned from continents to the sea) and the water of oceans loss by evaporation must be equal (Marcinek, 2007). To satisfy the water balance equation at the global or continental scales, the observations of all the components of the hydrological cycle with a global perspective are required. Particularly, precipitation needs continuous monitoring, as it is the major component of the cycle.

3- Water balance components/fluxes

<u>3.1- Atmospheric water</u>

Atmospheric water consists of precipitation and evapotranspiration. Precipitation accounts for the major contribution to the water balance of a terrestrial control volume and consists of water that drops from the atmosphere in either liquid or solid-state. Precipitation is generated by the condensation of moisture in the atmosphere, because of the cooling of an air portion. Formation of precipitation is driven by the origin of the lifting motion that triggers it, which eventually leads to various temporal and spatial rainfall regimens, which are typical of the climatic type of an area. According to the *Köppen-Geiger Climate Classification System* (Köppen, 1936), five major climatic types are recognized based on the annual and monthly averages of temperature and precipitation:
A - Tropical Moist Climates: all months have average temperatures above 18° C, which cause the daily intense convective precipitation;
B - Dry Climates with deficient precipitation during most of the year, which lead to convective precipitation events (while in general precipitation rate is low, but occurs in form of extreme events with high intensity);
C - Moist Mid-latitude Climates with Mild Winters, with mid-latitude cyclones causing the winter storms;
D - Moist Mid-Latitude Climates with Cold Winters, similar to the C category, but with precipitation mostly in form of snowfall in winter;
E - Polar Climates: with extremely cold winters and summers, which cause dry conditions with a low amount of precipitation and mostly in the form of snow.
Rainfall is the liquid form of precipitation while reaching the earth. Other forms of precipitation are snowfall, which is frozen water in a crystalline state; hail that is frozen water in a massive state; sleet, which is melted snow, regarded as a mixture of snowfall and rainfall.
Precipitation is characterized by high spatial and temporal variability, which can be analyzed by its main attributes:

- *Depth* (volume of precipitation accumulated on a horizontal surface area in a certain time, if precipitation can not drain, evaporate or percolate from this surface, dimensions $[L^3/L^2]$);
- *Duration* (the time from the start to the end of precipitation, [T]);
- *Intensity* (time rate of rainfall depth, equal to the ratio of the precipitation depth by its duration, [L/T]).

Usually, precipitation depths and intensities of a storm are graphically reported in charts that are called *pluviographs* and *hyetographs*.
Other characteristics determining precipitation variability are:

- *Frequency* (the number of times, during a certain period, that precipitation of a specific magnitude or greater occurs);
- *Pattern* (shape of the temporal diagram of a precipitation event);
- *Areal extent*;
- *Movement*;
- *Location*.

Rainfall frequency provides the information on how often precipitation with a given characteristic is likely to happen, which will consequently determine the frequency of occurrence (or return period) of the resulting runoff (in particular, the frequency of the peak flow). Precipitation pattern, areal extent, and movement determine how a portion of the drainage area contributes over time to the runoff, usually caused by the type of storm (rainfall event, which is caused by the original climate conditions). For instance, precipitation associated with cold fronts (thunderstorms) tends to be more regional, faster moving, and of shorter duration, while warm fronts tend to generate slowly moving storms of broader areal extent and longer durations. Moreover, the location where a regional storm occurs in the catchment influences the temporal distribution of the runoff. To give an example, a storm falling near the catchment outlet will result in a very quick occurrence of peak flow, as well as a rapid passage of the flood. Precipitation movement affects the runoff rate, depending on the catchment shape (particularly in elongated catchments).

Duration, intensity and frequency of precipitation are often considered in combination, to constitute the intensity-duration-frequency (IDF) curve. This is the diagram of intensity versus duration that is provided for each frequency (or return period) of precipitation. The IDF curves are specific for a given location. The storm (the so-called *critical event*) used for predicting or estimating the runoff *hydrograph* (see sub-chapter 3.5) with a specific return period (or a certain frequency) is considered as "*design storm*", which can be derived from the IDF curves or the statistical analysis of observed rainfall. Since the amount and timing of runoff depend on the magnitude, the spatial and temporal distribution of rainfall, the hyetographs of both actual and design storms are fundamental elements for projects requiring hydrological information and for hydrological modeling (McCuen, 1982; Chow et al. 1988; Haan et al., 1994).

The other component of atmospheric water is evapotranspiration (ET) that includes the processes in which water is transferred into gas flux (vapor) to the atmosphere. ET processes are generally referred to as (Labedzki, 2011):

- *Evaporation,* the water transfer from the surface of a water body or from bare soil to the atmosphere;
- *Transpiration,* the water absorbed by vegetation roots from the soil and routed through the leaves (across the canopy stomata) to the atmosphere;
- *Evaporation of intercepted water,* the share of precipitation which falls onto the vegetation canopy surface and is directly returned into the atmosphere.

We can distinguish potential and current ET. Potential ET can be defined as the water loss from a surface with no water limitation. It can be expressed as a function of the physical variables of the atmosphere (i.e., depends on the energy that is available to convert liquid water to vapor from climatic driving forces, like solar net radiation), where the resulting water vapor can freely move away from the surface. The water content of soil and ET are known to be highly correlated. To calculate the actual ET, potential ET is reduced based on real soil water content (Beven, 2011). Allen et al. (1998) identified the following variables influencing ET:

- *Meteorological parameters*, such as solar radiation and air temperature, humidity, and wind speed;
- *Crop parameters,* such as species, growth stage, height, ground cover, water stress, and rooting characteristics;
- *Soil parameters*, such as roughness, salinity, fertility and albedo; and
- *Management parameters*, such as the application of fertilizers and soil cultivation practices.

The estimation of ET from vegetated areas is a basic tool to compute the water balance for estimating the water requirements of irrigated crops and for planning water management. Since direct measurements of ET are difficult, in many cases it is easier to estimate ET fluxes as a residual of the water balance equation, or by application of models using meteorological and other data as input. Generally, the difference between precipitation and ET largely controls the amount of water surplus in a hydro-system (except at the event scale). Sub-chapters 4.1.1 and 4.2 provide an overview of methods and models widely applied to measure and/or estimate ET.

3.2- Surface water

Surface water is the hydrological response of soil to a precipitation event. During a storm, the share of atmospheric water that is not intercepted by vegetation does not infiltrate or percolate through soil (see section 1), flows by gravity over the soil along hillslopes and then in stream channels (*surface runoff*) (Figure 4). Of the *total runoff*, the share directly generated by rainfall takes a rapid route to the stream channels (*direct runoff* or *quickflow*), while the other part of precipitation that infiltrates into the soil takes a much slower route (*base flow* or *delayed flow*) (Ward and Robinson, 1976). Generally, while interception, evapotranspiration and infiltration run out with precipitation, runoff may also continue in dry periods, fed by the base flow. Runoff is usually expressed as water depth per time and space units (mm, that is m^3 per m^2 of the area and per hour or day), which makes the comparison with precipitation (measured by the same unit) simpler.
The *total runoff* or *streamflow* is the result of several water flow paths (Ward and Robinson, 1976) (Figure 4):
(i) *Direct precipitation* over the water surface;
(ii) *Overland flow*;
(iii) *Interflow* (shallow sub-surface water);
(iv) *Groundwater flow* (deep sub-surface water).

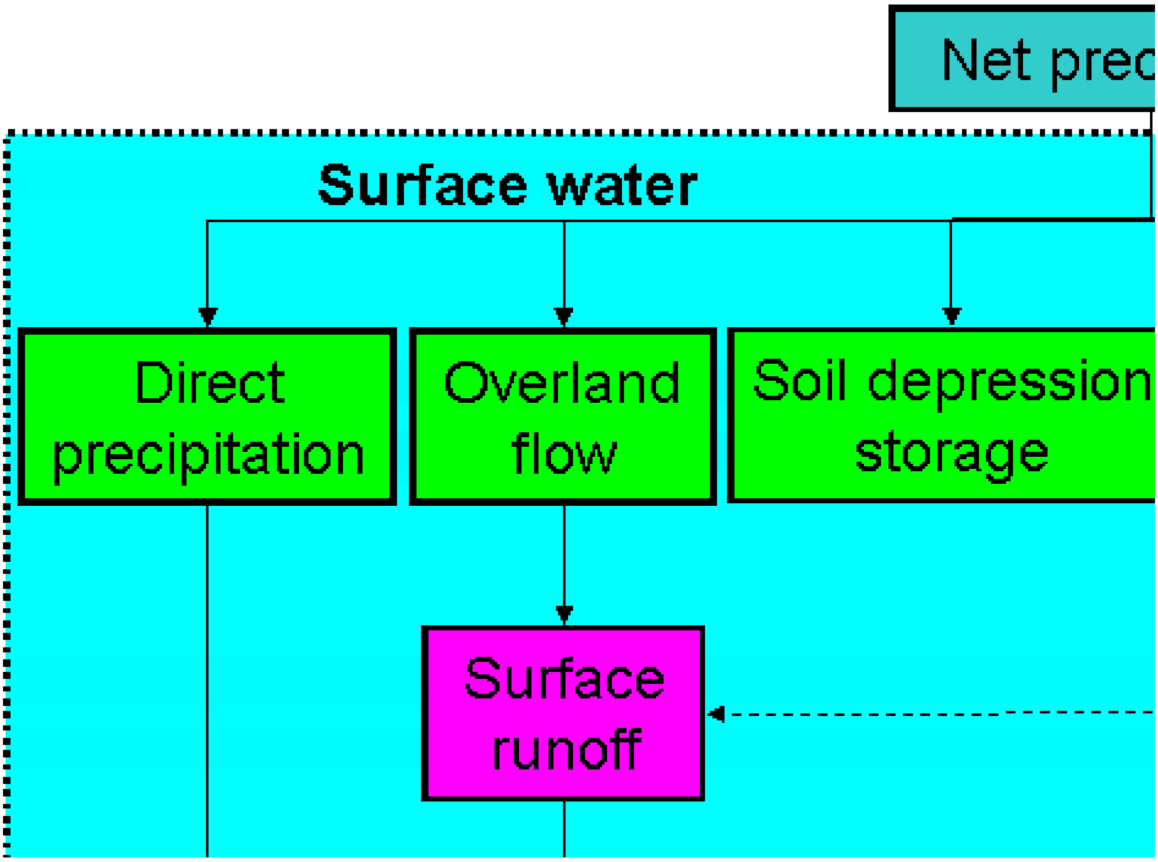


Figure 4 - Schematization of total runoff components (from Ward and Robinson, 1976, modified).

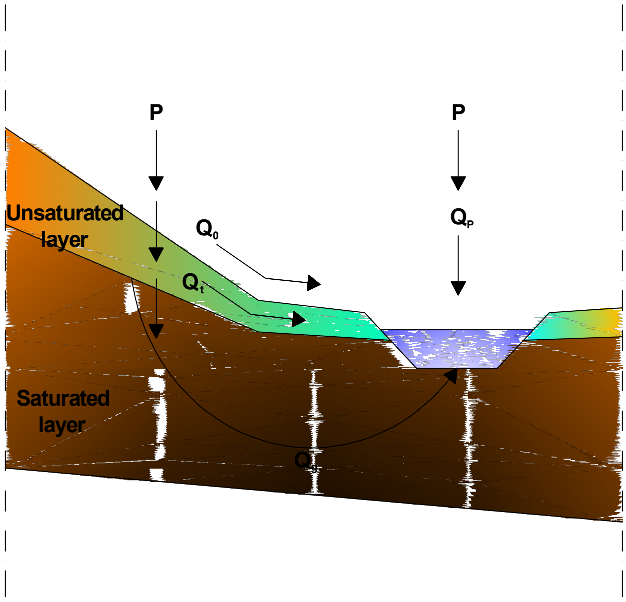


Figure 5 - The different components of runoff along a hillslope (P = precipitation; Q_p = direct precipitation over water surfaces; Q_o = overland flow; Q_t = interflow; Q_g = groundwater flow).

The direct precipitation is the share of the atmospheric water input that directly falls over the water surfaces (lakes, artificial reservoirs, stream channels of the hydrographic network). This share is usually limited since these surfaces cover a very small area of the catchment system.
The overland flow includes the sheet flow (the laminar water stream flowing downslope) and the concentrated flow into rills and gullies. The overland flow generates surface runoff, which is the faster component of the quick flow (Figure 4).

3.3- Sub-surface water

The interflow is the share of the water infiltrating through the soil surface that flows laterally (calculated according to Darcy's law, regulating the filtration process into porous media) into the upper soil layer to the channel. This flow occurs in the unsaturated zone or the capillary edge over the aquifer when the horizontal hydraulic conductivity is much higher compared to the vertical conductivity into the soil profile. Moreover, the horizontal hydraulic conductivity decreases with soil layer depth (in absence of artificial disturbance, such as tillage and soil compaction), which makes the interflow of the upper soil layers much faster than the delayed interflow and groundwater flow in the deeper layers. Quick interflow and part of the delayed interflow feed the sub-surface runoff (Figure 4).
The water that moves vertically is known as deep percolation, which is the water flux below the root zone. The groundwater flow is fed by the percolation of the infiltrated rainfall into the deep soil layers, reaching water tables, and runs towards the stream channel through the saturated zone. Sometimes, the infiltrated rainfall can directly generate the groundwater flow. The groundwater flow is delayed by days or even months with respect to precipitation, surface runoff or interflow, due to the very low hydraulic conductivity, but does not fluctuate rapidly (Figure 5).
Groundwater flow and delayed interflow, beside a share of the sub-surface runoff, are the components of the base flow (or delayed flow) (Figure 4).

The different time rates and amounts of the runoff components of the streamflow determine the nature and magnitude of the hydrological response of a territorial unit (e.g., plot, hillslope, and catchment) to precipitation. The current comprehension of this hydrological response can be attributed to two different runoff generation mechanisms, conceptualized by two famous hydrologists (Horton, 1933; Hewlett, 1961).

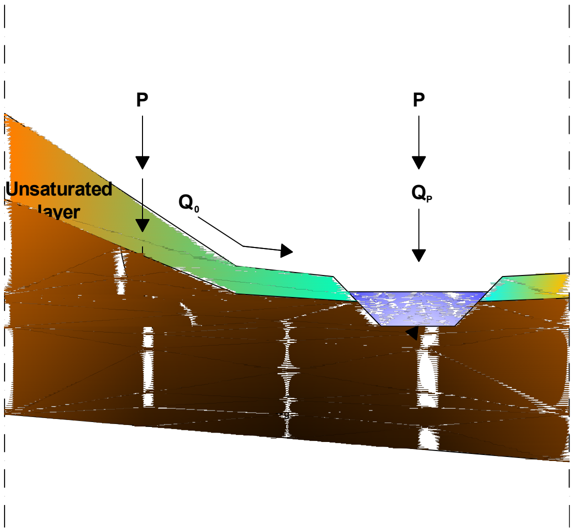
3.4- Runoff generation mechanisms

3.4.1- Infiltration-excess (or Hortonian) mechanism

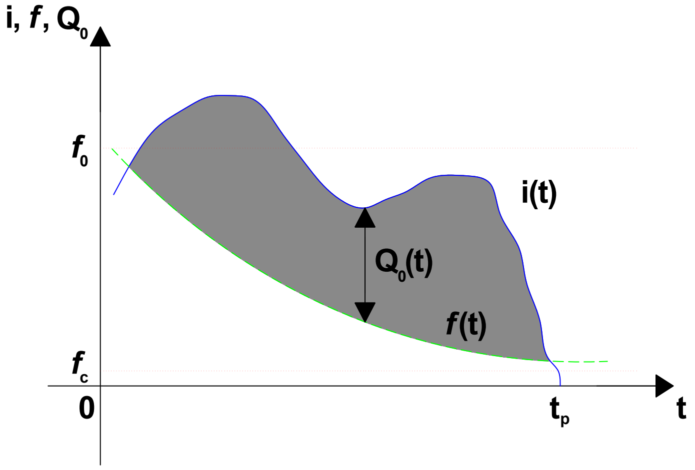
In 1933, R.E. Horton hypothesized that, at the soil surface, the shares of net precipitation infiltrating or moving over the soil as overland flow strictly depends on the *soil infiltration capacity* (f). This is the maximum rate at which rainfall infiltrates into the soil when water is continuously and sufficiently available over its surface (Hillel, 1998). Once the storm starts, f gradually decreases with time until a steady value (f_c), due to the progressive soil saturation during precipitation; the decrease rate of f theoretically follows an exponential law from an initial value (f_0) until the asymptotic f_c. After the storm event, the initial f_0 is recovered. If f is higher than the rainfall intensity (i), all net precipitation infiltrates, feeding the sub-surface flows, and no runoff is observed (Figure 6a). By contrast, if $f < i$, the excess precipitation, equal to difference $i - f$, is the overland flow (Figure 6b), accordingly defined as *infiltration-excess (or Hortonian) overland flow*. The runoff generation mechanism conceptualized by Horton is typical of the arid or semi-arid areas, where the overland flow is produced by infrequent but heavy rainfalls and is influenced by soil surface processes, such as sealing, cracking, and freezing.

3.4.2- Saturation-excess (or Hewlett's) mechanism

Later, in 1961, J.D. Hewlett hypothesized that, during intense and prolonged events, the precipitation infiltrating into the soil feeds the sub-surface water. Due to the progressive saturation of the soil profile, the water table rises up, primarily starting on the lower slopes of the catchment and along valley bottoms adjacent to stream channels (Figure 7a). Throughout the storm, progressively larger areas, on which the soil infiltration capacity decreases to zero, saturated in the catchment, and the excess precipitation feeds the surface water, accordingly defined *saturation-excess overland flow* (Figure 7b); only the saturated areas contribute to surface runoff. The runoff generation mechanism theorized by Hewlett is typical of the humid and sub-humid areas, where the morphology and other properties of soils let the water table rise easier during the precipitation events.

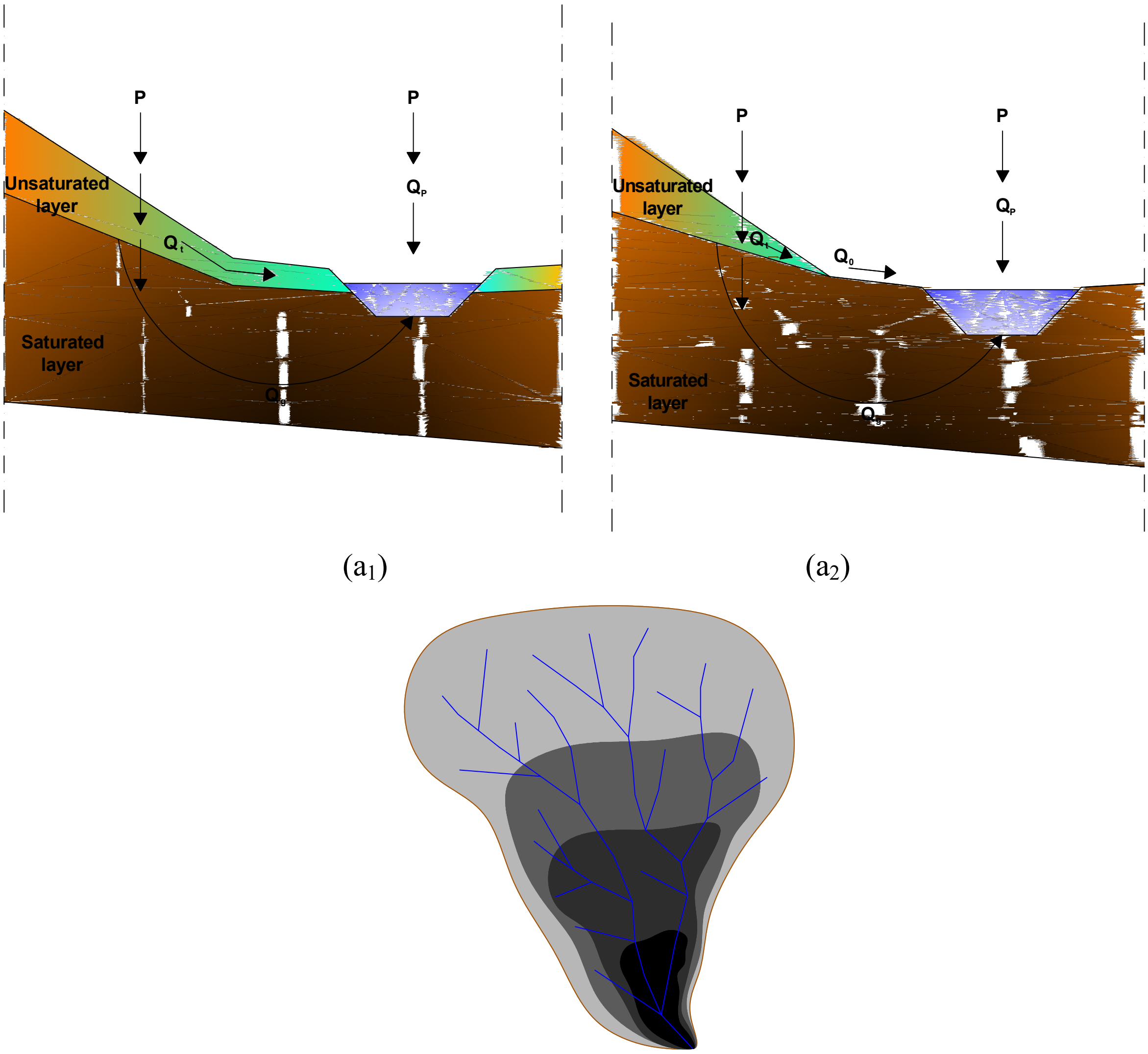


(a)



(b)

Figure 6 - The runoff generation mechanism by infiltration excess. (a) Flow paths in the hillslope profile (P = precipitation; Q_p = direct precipitation over water surfaces; Q_o =overland flow); (b) the theoretical diagram (f = infiltration rate; f_c = steady infiltration rate; f_0 = initial infiltration rate; t_p = precipitation duration; i = precipitation intensity; Q_0 = overland flow).



(a$_1$) (a$_2$)

(b)

Figure 7 - The runoff generation mechanism by saturation-excess. (a) Flow paths in the hillslope profile at the start (a$_1$) and during (a$_2$) precipitation (P = precipitation; Q_p = direct precipitation over water surfaces; Q_o = overland flow; Q_t = interflow; Q_g = groundwater flow); (b) the progressive saturation mechanisms of the contributing areas (saturation degree increases with the grey intensity).

3.5- Temporal evolution of surface and sub-surface water

In humid and sub-humid climates, *perennial* watercourses have permanent water flow. In arid or semi-arid areas (such as in the Mediterranean environment), watercourses are *intermittent*. Here, the water flows in the channels for some months during the year. In *ephemeral* watercourses, the water flows only for hours or days following a storm (Wohl, 2017).

Usually, a diagram (hydrograph) graphically represents the time evolution of runoff, from the precipitation start until the flood depletion. The hydrograph is commonly reported together with the hyetograph (that is, the time evolution of the rainfall intensity) of the generating precipitation event. Schematically, a hydrograph consists of three components:

- A rapidly increasing curve (*concentration or rising limb*), which starts from the time when the hydrological response of the stream channel to the precipitation begins by an increase in discharge (see point S of Figure 8) until to the *flood peak* (the maximum value of the surface runoff, point P of Figure 8);

- The peak discharge - that is the highest point on the hydrograph, - which occurs when different parts of the catchment simultaneously contribute to the runoff at the outlet.
- A slowly decreasing curve (*recession or falling limb*), which represents the runoff depletion after the storm until the time when the stream flow regime returns to the regular value of the dry period (baseflow).

The duration of the concentration limb is much lower compared to the recession limb. The time between the precipitation start and the flood peak is called *concentration-time* (Figure 8), while the time between the flood start and end is called *flood duration*.

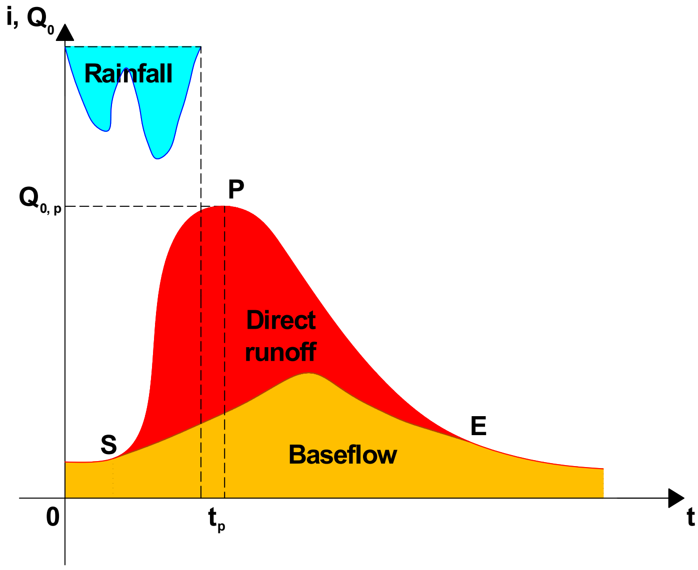


Figure 8 - Schematic flood hydrograph with a possible separation method (t_p = rainfall duration; $Q_{0,p}$ = peak runoff; i = rainfall; Q_0 = runoff; S = flood start; P = hydrograph peak; E = flood end).

The hydrograph shape is influenced by the amount and velocity of each runoff component. It is practically impossible to individually identify these components (except for very small catchments with a simple hydrographic network, such as the headwaters). Conversely, the separation of the *quick flow* and *delayed flow* components is generally feasible, using a set of arbitrary graphical techniques, which allows the identification of the direct runoff and the base flow hydrographs. Each separation technique requires the detection of the start and end of the flood. While the flood start is easily recognizable (it is simply the time when the hydrograph suddenly increases after the precipitation start), the identification of the flood end is much more difficult. The literature proposes hydrograph separation techniques of different complexity. The most common techniques are based on drawing a line passing across the hydrograph from the point of flood start:

(i) A horizontal separation line;

(ii) A line of constant slope (0.000546 m^3 s^{-1} km^{-2});

(iii) A straight line to a selected point on the recession limb (point *E* of Figure 8), which can alternatively be: (a) the point of greatest curvature close to the lower end of the recession limb; (b) the point when the recession limb starts to decrease according to an exponential law:(c) the point corresponding to a given time interval from peak flow.

(iv) the broken line consisting of (1) the line prolonged from the pre-storm hydrograph below the concentration limb until the vertical under the peak flow, and (2) the line whose extremes

are the intersection of the previous line with the vertical under the peak flow and the point over the recession limb selected as above at the point (iii).
The technique (i) is simple but can lead to unrealistic results (that is, a constant hydrograph of sub-surface water, which thus does not reflect the effects of the precipitation). By contrast, the techniques (ii) to (iv), although being complex, seem to be more realistic and appropriate to reproduce the actual hydrological effects of floods.

4- Methods for components/fluxes estimation in the water balance

4.1- Field measurements

4.1.1- Atmospheric water

4.1.1.1- Precipitation

The amount of precipitation falling on the ground and retained over the soil without any water losses (evaporation, infiltration, depression storage) or runoff is measured (in m^3 or mm) at regular time intervals (e.g., daily; hourly; every 15 min etc.) using rain gauges. These devices must be installed in the open air on a horizontal surface close to the ground (to reduce wind effects), but avoiding rain splash or submersion by floods, and far from obstacles (to prevent rainfall interception).

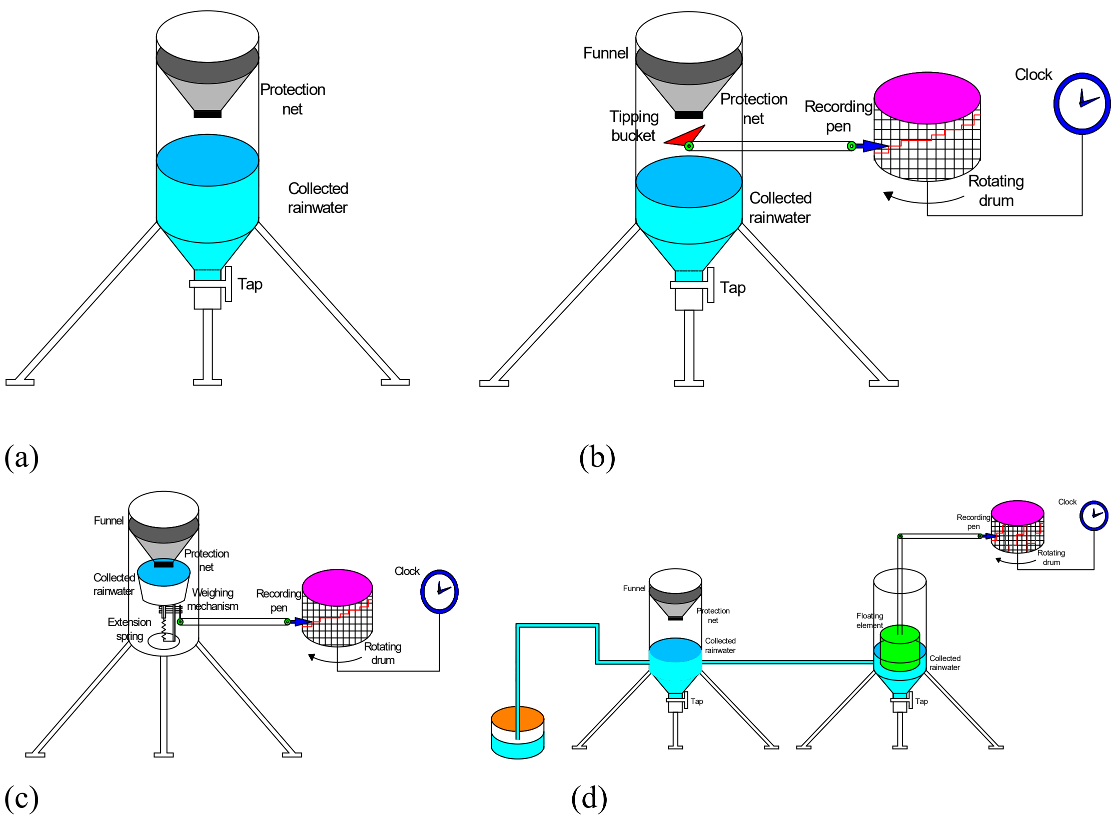


Figure 9 - Sketches of some rain gauge types (a, pluviometer; b, tipping bucket pluviograph; c, weighing pluviograph; d - natural syphon pluviograph).

Rain gauges generally consist of open cylindrical vessels collecting rainfall with or without recording equipment. Some rain gauges (pluviometers) do not record the precipitation depth. A pluviometer is simply a small metal tank (diameter of about 0.10 m) placed on a horizontal plane at a height of about 0.30 m above the ground level. The tank contains an upper funnel receiving precipitation, which is collected into a lower bottle or the tank bottom; the upper net protects the pluviometer from dirt (Figure 9a). Once a day, the collected rainwater is manually measured, and this record is the total rainfall of that day.
Other rain gauges (pluviographs) continuously and automatically record the precipitation depth. A writing pen activated by the surveying equipment traces the record on a graduated paper mounted on clockwork driven drum or mass memory. These records, which also report the duration of the precipitation event, allow the rainfall intensity calculation. The types of pluviographs are:

- Tipping bucket pluviograph (Figure 9b), based on a pair of small buckets under the funnel, which alternatively tip, when a given rain volume falls into the buckets and actuates the writing pen;
- Weighing pluviograph (Figure 9c), where a weighing mechanism under the tank receiving the rainwater is connected to the writing pen;
- Natural siphon pluviograph (Figure 9d), where the collected rainwater is poured into a float chamber, causing the float to rise and actuate the writing pen. When full, a syphon pipe automatically empties the float chamber and the writing pen is reset to zero for the next record.
- Recently, the rainfall measurements of rain gauges are progressively replaced by estimations by radar. The latter devices use microwaves with a wavelength from 0.03 to 0.10 cm and operate at several hundreds of kilometers from rainstorms. Beside rainfall amounts and intensity, radar allows the simultaneous measurement of areal extent, location, and movement of the storm and even the velocity and distribution of raindrops.

4.1.1.2- Evaporation and evapotranspiration

Evaporation from water bodies (e.g., lakes, reservoirs, river channels) and evapotranspiration from vegetated surfaces can be directly measured or estimated using indirect methods.
Evaporation from water bodies is directly measured by evaluating the reduction of level for a sample of open water in an *evaporation pan* (with standardized size) over time; generally, the related atmospheric variables, such as the precipitation, temperature, wind speed and humidity of air are simultaneously measured using a weather station. Daily, water evaporation from the pan (expressed in mm/day) is the difference of the water levels in two consecutive days, depurated from rainfall over the pan and other water losses, such as bird or animal consumption. The water evaporation of the specific pan must be corrected through a coefficient (K_p), which depends on the type of pan, environment and operations.
Evapotranspiration from vegetated surfaces is directly measured using *lysimeters*. A lysimeter is a hydraulically isolated tank filled with soil cores (with or without vegetation). In this tank, evapotranspiration is evaluated weighing the lysimeter over time and simultaneously measuring the precipitation over the tank and the drainage from the sample as components of the water balance equation.
The indirect methods for estimating evaporation and evapotranspiration are:

- *Bowen ratio method*, which estimates evapotranspiration as a function of the Bowen ratio (sensible to latent heat), on its turn calculated from measures of the atmospheric temperature and humidity gradients close to vegetation;
- *Eddy correlation* (or *covariance*) *method*, which estimates evapotranspiration as the temporal average from the correlation coefficient between variations in vertical wind speed and atmospheric humidity measured above the vegetation.

Recently, other methods based on remote sensing systems (such as drones and satellites) have been proposed. These systems measure atmospheric variables (such as the water vapour concentration, temperature, sensible heat, aerodynamic exchange resistance) surrounding the measurement area, and calculate the evapotranspiration fluxes from these derived measurements using mathematical algorithms and/or energy balance equations.

4.1.2- Surface water

Surface runoff can be measured using methods that can be classified as follows (Dobriyal et al., 2017, Figure 10):

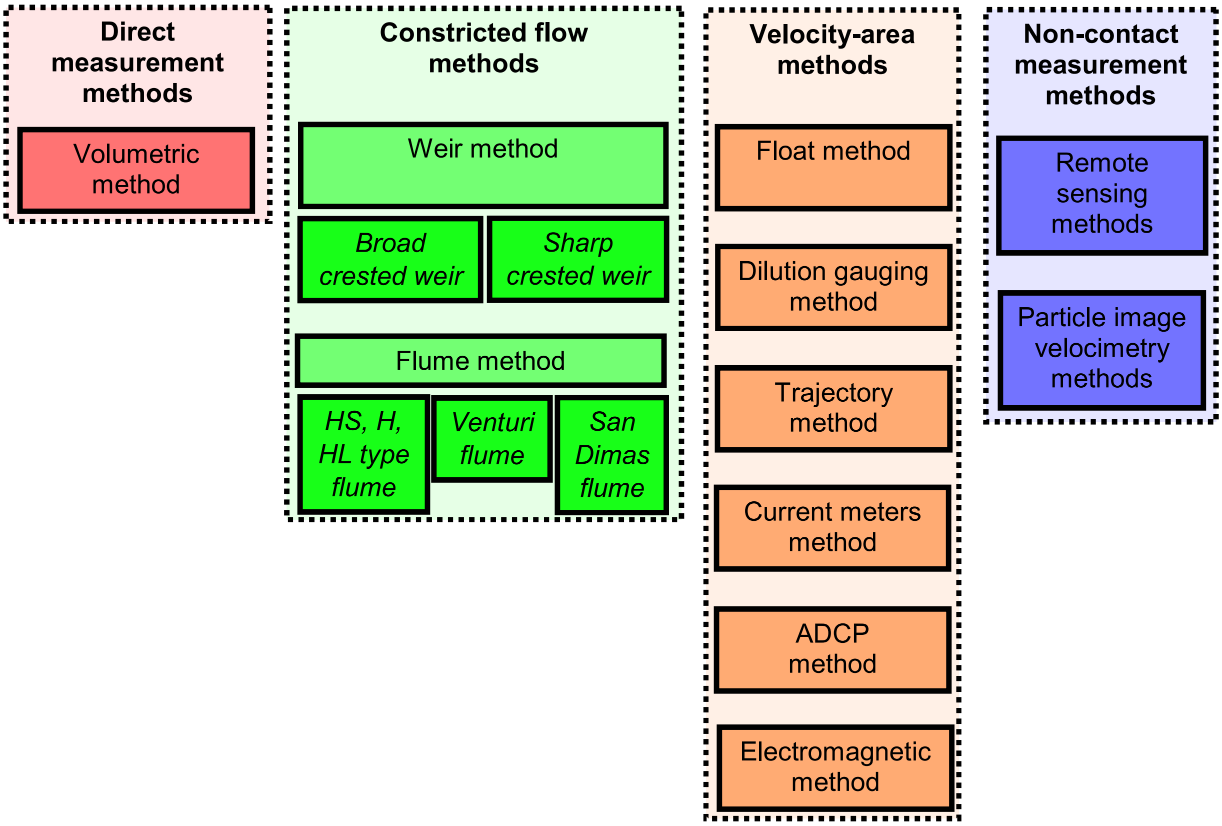


Figure 10 - Classification of methods for surface runoff measurement (source: Dobriyal et al., 2017, modified).

- *Direct method* (it is a *volumetric method*, based on filling a tank of known volume during a given time);
- *Velocity-area methods*, based on the integration of simultaneous measures of local low velocities by cell areas of a channel cross-section, including:
  - The *float method*, where the flow velocity is the ratio between the floating distance of an object of low density and the related travel time;
  - The *dilution gauging method*, which measures flow velocity as a function of the diffusion rate of a tracer, e.g., chemical or radioisotopes;
  - The *trajectory method,* where the flow velocity is measured using the hydraulic equations at the outlet jet of pipeline in which all the streamflow is diverted;
  - The *current meter method*, assuming that the flow velocity is proportional to the rotation speed of a mechanical rotor (Figure 11a);
  - The *acoustic Doppler current profiler method*, where the flow velocity is estimated by the difference in the frequency of the sound transmitted by the device into the water and echoes received from suspended particles (Figure 11b);

- The *electromagnetic method*, measuring by electromagnetic probes placed on each side of the stream the electromotive force induced in the water by a generated earth's magnetic field, which is directly proportional to the flow velocity.

(a) (b)

Figure 11 - A current meter (Valeport Inc., UK, a) and an acoustic Doppler profiler (SonTek Inc., USA, b).

- *constricted flow methods*, based on forcing the water stream passing over a broad crested or sharp-crested weir (*weir method*, Figure 12a) or an HS, H and HL type, Venturi, Parshall and San Dimas flumes (*flume method*, Figure 12b) of known geometry, for which the application of the hydraulic equations to the measured water depth gives the surface runoff;

(a) (b)

Figure 12 - A Parshall flume (Badger Meter GMBH, Germany, a) and a triangular weir (IEI Inc., USA, b).

- *Non-contact methods*, consisting of the *remote sensing methods* (using passive or active sensors, which provide:
    - Direct measures of water surface levels from radar altimeters/ high-resolution satellite imagery, or
    - Correlations of remotely-sensed water surface areas with ground measurements (water depths or discharges), and the *particle image velocimetry method* (i.e., determining the water velocity recording the laser light scattered by liquid or solid particles on a photo-camera).

Overall, the direct and constricted flow methods are quite accurate but are advised for surface runoff measures in very small channels. The other methods are more suitable for streamflow measurements in medium to large rivers. In particular, the *velocity-area methods* can be used in easily accessible watercourses for instantaneous measurements and construction of water depth-discharge equations. The non-contact methods, although being quite expensive, do not require the presence of surveyors close to the channel and are better suitable for continuous and real-time flow measurements also in the case of floods.

4.1.3- Sub-surface water

Measuring infiltration is of fundamental importance since the related process governs both the surface and sub-surface water. The infiltration measurements are based on field evaluation of the hydraulic conductivity, which is not constant, but varies in time (with soil saturation during a storm) and space (from point to point, depending on several soil properties, such as texture and aggregate stability). The methods for soil hydraulic conductivity measurement use *infiltrometers* or *permeameters* that can be classified as follows (Angulo-Jaramillo et al., 2000):

- *One-dimensional pressure ring-infiltrometers* (e.g., *one-ring and double-ring infiltrometers, mini-disk infiltrometers,* Figure 13a), in which the water, supplied to the soil surface at a positive pressure head, infiltrates vertically in a ring (or two coaxial rings) pressed into the soil. This process carries on until a constant infiltration rate is observed, which is assumed to be the soil infiltration capacity;
- *Unconfined three-dimensional tension disc infiltrometers* (Figure 13b), which supply the soil with water at negative pressure at its surface (to prevent wetting up of soil larger pores with possible short circuits for flow), thus allowing the evaluation of soil-water properties of the soil matrix without being dominated by flows in the larger pores (Youngs, 1991).

(a) (b)

Figure 13 – Minidisk infiltrometer (Decagon Inc., USA, a) and double-ring infiltrometer (b).

Infiltration can be also indirectly measured using *rainfall simulators* (Figure 14). These devices generate an artificial rainfall with controlled depth, intensity and drop size; the infiltrated flow is the difference between precipitation and surface runoff, and the ratio to the infiltration time gives the hydraulic conductivity.

Figure 14 - A rainfall simulator (Eijkelkamp, Nederland).

4.2- Modeling

Experiments about hydrology and specifically the estimation of the water balance components are usually costly and time-consuming, and sometimes not even applicable on large spatial scales and long-term periods. A possible alternative solution is the use of models in replacement of hydrological experiments and to the calculation of the water balance. In order to develop appropriate and sustainable strategies for management of water resources ì, the modeling approach, combined with field observations and laboratory experiments, allows a better understanding of the hydrological cycle and provides scientifically sound information about the hydrological processes and fluxes. According to the description of hydrological models provided by Kirkby (1996), "Models are thought experiments which help refine our understanding of the dominant processes, testing whether we have a sufficient and consistent theoretical explanation of physical processes". Water balance models generally simulate all components of the terrestrial hydrological cycle and the interaction of surface and sub-surface processes holistically, maintaining a continuous water balance for the area of interest (Wagener et al., 2004; Beven, 2011). Some of these components are interrelated, and therefore require iterative calculations; to model some hydrological processes, the estimation of a certain number of input parameters is required. Water balance modeling is based on equations (1) to (3), depending on the control volume. In principle, modeling the water balance sounds simple; however, in practice, it is difficult to measure or estimate every single component of the hydrological cycle, particularly at larger spatial scales (i.e., from hillslope to continental scales). Such models can incorporate the spatial and temporal variability of the primary driving forces, such as precipitation and solar radiation, and land-surface heterogeneity (e.g., soil, vegetation). A mathematical description based on laws of physics (e.g., mass and momentum conservation applied to soil and water) is the first step in the formulation of a model that will produce quantitative predictions of the hydrological processes and fluxes. The general structure of all water balance models is similar. To set up a model, it is necessary to write equations that relate the rates of change in water storage in the control volume to the hydrological fluxes across its external surface throughout the reference period. To summarize, although many equations differ in their structure, complexity and input parameters, the most common equations related to the estimation of hydrological processes and/or fluxes in water balance models are the following:

- For sub-surface flow in:
  - Saturated zone: Darcy's law (Darcy, 1856), which assumes a linear relationship between the flow velocity and hydraulic gradient through a coefficient of proportionality (*hydraulic conductivity*);
  - Unsaturated zone: Richards' equation (Richards, 1931), which is the combination of Darcy's law with the continuity or mass balance equation in a non-linear partial differential equation;
- For surface flow: (i) Saint-Venant equations (1797–1886), which assume that the flow can be expressed in terms of average cross-sectional velocities and depths, and are based on the balances of both flow mass and momentum; (ii) Diffusion wave and Kinematic wave equations, which are simplifications of Saint-Venant equations, where some terms are neglected (Lighthill & Whitham, 1955);
- For atmospheric water flow (evapotranspiration): (i) Penman-Monteith's equation, which is the most widely used and recommended method to directly estimate the potential ET, and indirectly actual ET by reducing the potential ET according to the actual soil water content (Monteith et al., 1965); (ii) Haude (Haude, 1955); (iii) Hamon (Federer and Lash, 1978); (iv) Hargreaves-Samani (Hargreaves and Samani, 1982); (v) Thornthwaite (Thornthwaite, 1948), and other models; however, the low availability of meteorological data and field measurements may be a limiting factor in applying some of the more demanding methods/equations and particularly Penman-Monteith equation.

When it is necessary to calculate the infiltration rate at the soil surface, Horton (Horton, 1933; 1940) and Green–Ampt (Green and Ampt, 1911) models can be used.

In general, the water balance models allow the estimation or prediction of surface and sub-surface flows and are commonly known as *hydrological models*. The models mostly differ in how ET and soil water content are conceptually considered and mathematically simulated. There are several classifications of hydrological models, based on structure, spatial scale, time scale, a time step of computation, interpretation of the catchment processes, etc. Table 1 and 2 illustrate two possible classifications of hydrological models. These classifications are based on *model structure* and *spatial processes*, respectively. The model structure, roughly varying from *simple model* to *complex model* based on the governing equations and number of modeled variables/input parameters, identifies how water balance components are calculated. Simple models need relatively few variables, while the most complex models (such as the physically-based models, Figure 15) require a large number of interconnected variables to simulate the hydrological processes and fluxes in the water balance of the hydro-system (Sitterson, et al., 2017).

Table 1. Classification of hydrological models according to the structure (Pechlivanidis, et al., 2011; Sitterson, et al., 2017).

Characteristics	**Hydrological model**		
	Empirical	*Conceptual*	*Physically-based (process-based or mechanistic)*
Methodological approach	Non-linear statistical relationship between inputs and outputs; observation-oriented; black-box concept	Simplified water balance equations representing hydrological components in the catchment	Physical laws formulated as partial differential hydrodynamic and porous media flow equations and resolved by numerical techniques
Advantages	Small number of input parameters; fast computational time	Simple model structure; easy to calibrate	Very accurate; the connection between model parameters and physical catchment characteristics
Limitations	Lack of physical significance between model parameters and catchment properties; input data falsification	Spatial variability within catchment not entirely addressed; lack of physical meaning in governing equations and parameters	A large number of data and parameters needed for running and calibration; catchment-specific
Most suitable applications	Ungauged catchments; runoff only desired output; rough estimation of output	Limited computational time; low detail of catchment characteristics	Availability of large and accurate input data; fine spatial and temporal scales
Examples	SCS-Curve Number; Artificial Neural Networks	TOPMODEL; HSPF; HBV; Stanford	MIKE-SHE; KINEROS; VIC; WaSiM-ETH

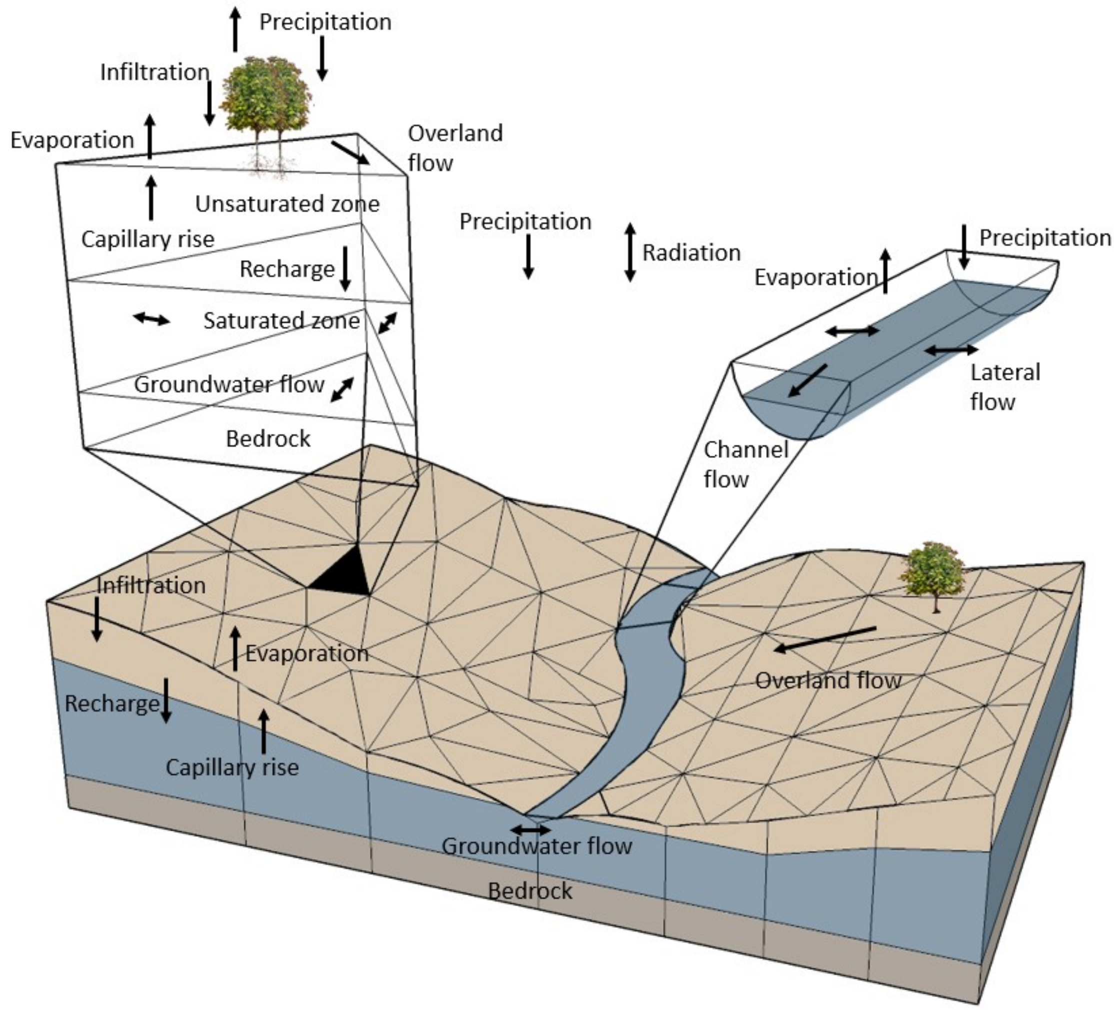


Figure 15. Schematization of the hydrological processes and control volume (in soil layers and Triangular Irregular Network) of a physically-based model.

Spatial variability in geology, topography, vegetation, and soil influence the hydrological processes (and in particular the rainfall-runoff transformation) within a catchment, and, thus, should be carefully considered in modeling (Beven, 2012). According to the spatial structure classification, the hydrological models can be classified as *lumped* (average weather and geomorphological conditions are assumed for the modeled catchment), *semi-distributed* (the catchment is discretized in sub-catchments or hydrologically homogenous response units), and *fully-distributed* (a catchment is discretized in grid cells) (Figure 16 and Table 2).

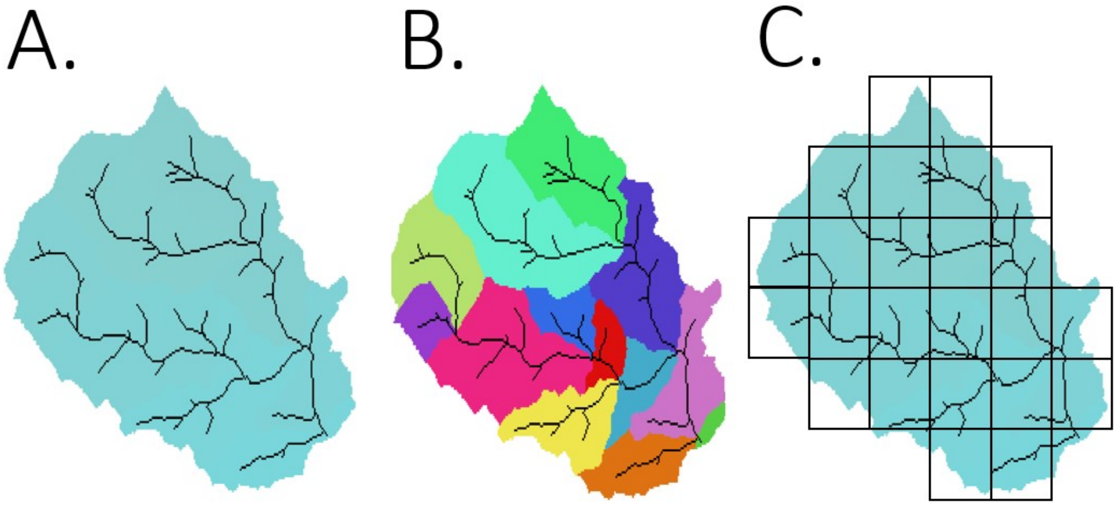


Figure 16. Spatial structure classification of hydrological models. A: Lumped model, B: Semi-distributed model, C: Distributed model.

With regard to the model time scale, a water balance is generally set up for an adequately long period, such as wet/dry season, calendar or hydrological year, decade, etc. Depending on variability of the captured hydrological flux or process of the water balance (e.g., surface water, rainfall, groundwater, or overall water availability), different temporal output data are provided, such as daily, monthly, seasonal, annual, or multi-year runoff volumes or sediment flows (averages or totals). Some hydrological models work at the event scale, that is, they estimate or predict the simulated hydrological variable as the product of a storm or a precipitation event. In addition to the time scale of models, the time step of computation (e.g., monthly, daily, hourly, etc.) is another important characteristic of the water balance models, since it influences the accuracy of the output variable and the computational time of the simulation procedure (i.e., the finer the time step, the longer the model computation).

A key issue for the practical use of the hydrological models is the reliability of their outputs for the modeler's purpose. The reliability of a model prediction or estimation is evaluated through a comparison of the modeled hydrological variable with a field measurement corresponding to the same meteorological input in an instrumented catchment. For example, the accuracy of a modeled peak flow for a flood after a storm is checked by its comparison with the maximum flow observed for the same storm and measured at the outlet of a catchment equipped by a flume (see sub-chapter 4.2). *Sensitivity analysis* and *calibration/validation* are useful procedures to facilitate model evaluation and application. Sensitivity analysis allows the identification of the model parameter(s) to which a model is most sensitive. Calibration adjusts one parameter or a set of parameters to make the simulated variable as close as possible to the corresponding observation. Model validation is the process to run a model by using model parameters determined during the calibration process (Moriasi, et al., 2007).

Table 2. Classification of hydrological models according to the spatial structure (Beven, 2012; Sitterson, et al., 2017).

Characteristics	Hydrological model		
	Lumped	**Semi-distributed**	**Distributed**
Methodological approach	Spatial variability is not considered; entire catchment is modeled as one unit; calculation of one runoff value for the entire catchment at the outlet; all data are constant over space and time	Reflect some spatial variability; dividing the catchment into smaller sub-catchments (Hydrological Response Units), with different parameters for each; calculate runoff at the pour point for each sub-catchment, but do not calculate runoff at every grid cell	Account for detailed spatial heterogeneity in inputs and parameters by grid cells (small elements); calculate distinct hydrological response for each cell separately
Input data	All data averaged for the entire catchment	Separated within the catchment but homogenous within the sub-catchments	All specific data at grid cell: DEM; land use; precipitation; soil properties; topography; and catchment characteristics
Advantages	Fast computational time; ideal for simulating average conditions	Represents important features in the catchment; fast computational time; fewer data and parameters needed than a distributed model	Physically related to hydrological processes
Limitations	Loss of spatial variability; not representative for large areas; over-or under-parameterization	Data into sub-catchments are averaged, and manipulation of input data is possible; loss of spatial resolution	Data intense; long computational time
Most suitable applications	Regulatory purposes that look at long-term conditions	-	For management practices by providing detailed data for small elements
Examples	Empirical and conceptual models; machine learning	Conceptual and some physical models; TOPMODEL; SWAT	Physically distributed models, MIKE SHE; VELMA; WASiM-ETH

From the notion of the *global water system* and the interdependency of earth components, which needs the integration of those systems in integrated models, *Global Hydrological Models* have been developed, in which the global water flow is connected to other hydrological systems through physical relationships (Alcamo et al., 2003). They are similar to catchment models but differ in processes description, parameter estimation approaches, the temporal and spatial resolution of input data and outputs. Global models provide useful spatial and temporal estimates of global water resources and, hence, the analysis of possible changes is attainable, particularly, under the explosion of global data availability from satellites in the last two decades (Sood and Smakhtin, 2015).

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