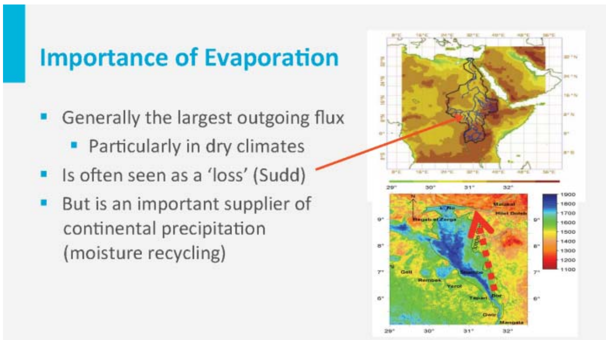
CTB3300WCx – Introduction to Water and Climate

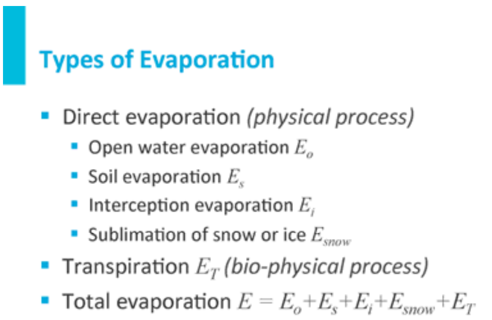
GWC 2 – Evaporation

Hubert Savenije

Welcome!
My name is Hubert Savenije and I am a Hydrologist.
Today we are going to talk about evaporation.



Of all moisture fluxes over the Earth's surface, evaporation is the largest outgoing flux. Of course precipitation is larger, being the source of all terrestrial moisture. But in general, evaporation is larger than river discharge, particularly in dry climates. Many professionals, particularly engineers, consider evaporation a 'loss': precious water that we 'loose' to the atmosphere. This thinking has lead to ambitious projects, such as the proposed draining of the Sudd wetland in South Sudan, through the Jonglei canal. But evaporation is not a loss. Evaporation is responsible for all biomass production through transpiration; And moreover much of the evaporated moisture returns to the Earth as recycled moisture, as we have seen earlier. If we don't understand evaporation, then we also can't understand river runoff, because they are closely intertwined



There are five distinct types of evaporation:
Split in direct and indirect evaporation
Direct evaporation consists of:
Open water evaporation: the evaporation from a free water surface
Soil evaporation: the evaporation of soil moisture reaching the surface through capillary rise
Interception evaporation: the evaporation of a wet surface after a rainfall event
Sublimation: the evaporation of snow or ice directly into the gas phase
The indirect evaporation is called Transpiration: the exchange of moisture through the stomata of vegetation
The total evaporation from the land surface is the sum of all these processes.
The first 4 are purely physical processes that transform moisture form the solid or liquid phase into the gas phase

Transpiration of moisture is a rest product of photosynthesis in vegetation.

Evaporation or 'evapotranspiration'

Avoid to use the term 'Evapotranspiration'

- 'Evapotranspiration' is opaque jargon for **bulk evaporation,** masking that we do not know its composition
  → *Use (total) evaporation instead*

Some people call the Total Evaporation "Evapotranspiration". I advise strongly against it. The term is opaque jargon that nobody understands outside an incrowd community. It hides the fact that evaporation consists of many different processes. Famous experts on evaporation, such as Brutsaerts and Shuttleworth, avoid the term.

Potential Evaporation

Potential evaporation, E_p

- Would occur if there is no shortage of water,
- Or other factors that may limit transpiration *(temperature, solar radiation, humidity).*

Actual evaporation, E

- occurs if these stress factors are accounted for

Evaporation is generally constrained by the supply of moisture. Logical: If there is no moisture available, then there is nothing to evaporate. We use the term Potential Evaporation for evaporation where there is no shortage of water, or where there are no other limiting factors besides the availability of energy. The actual evaporation is what you get if you take all limiting factors into account.

Average (annual) evaporation

$$\overline{E} = \overline{P} - \overline{Q}$$

$\overline{Q}$ is the mean annual runoff [mm/a]
$\overline{P}$ is the mean annual precipitation [mm/a]
$\overline{E}$ is the mean annual evaporation [mm/a]

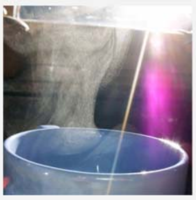
$$\frac{\overline{E}}{\overline{P}} = 1 - \frac{\overline{Q}}{\overline{P}} = 1 - C_R$$

In the previous module we looked at the water balance.
Here is the water balance:
The change in storage equals to the precipitation, minus the evaporation, minus the runoff. The water balance closely connects the precipitation, the evaporation and the runoff.
If we now look at averages, if we take the average over the year, over many years, then this term becomes very small, and we may often disregard it, if the average is taken over several years. The total amount of rainfall that has fallen over, for instance, 10 years is, of course, much larger than the difference in storage between the beginning and the end of the decade. The same can be said about the other fluxes.
So for long-term averages, indicated by a bar above the variables, a very simple relation applies:
Whereby the evaporation equals the difference between the precipitation and the discharge (all expressed per unit surface area). Dividing by the precipitation we see that the proportion of evaporated precipitation equals 1 minus the runoff coefficient.

Actual evaporation

$$E \leq E_p$$ Energy constraint

$$\overline{E} \leq \overline{P}$$ Moisture constraint



So there are two conditions that apply to evaporation:
The actual evaporation E is always less than the average annual potential evaporation. And the average annual evaporation should be less than the precipitation.
The first condition is an energy constraint. The second is a moisture constraint. These conditions form the asymptotes of the Budyko curve.

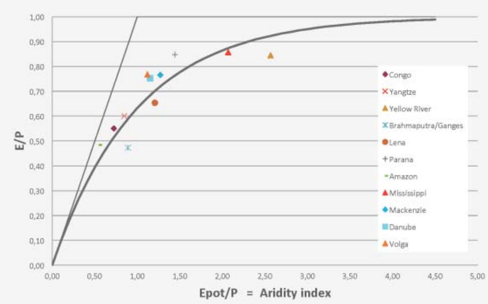
Budyko Curve

$$\frac{\overline{E}}{\overline{P}} = \left(1 - \exp\left(-\frac{\overline{E}_p}{\overline{P}}\right)\right) = 1 - C_R$$

- If P $\rightarrow 0$, $E = P$
- If P $\rightarrow \infty$, $E = E_p$

Budyko (1920-2001)

Budyko (1920-2001) developed a simple mathematical form to describe the relation between average annual evaporation, precipitation, and potential evaporation. There are many mathematical forms, but this is the simplest. For the Budyko curve we have a relation between the [...], which is defined by the potential evaporation divided by the rainfall, and these are average annual values, and this [...] is 1 in this point, if you are on the left of 1, you are in a wet climate, on the right you are in a dry climate. On the vertical axis we have the relative evaporation, defined by:
the average annual evaporation divided by the average annual precipitation. And of course this is bound by a maximum 1, which you could say is the moisture constraint. You can never have more evaporation than precipitation. And on the other hand, we have what we call the energy constraint, that means that the evaporation can never be more than the potential. And the Budyko curve nicely runs between them. And the distance between the moisture constraint and the curve, this is the runoff coefficient.



Here we plotted the water balance of some of the larger river basins of the world within the Budyko diagram.
The wet catchments (Amazon, Parana, Ganges and Yangtze) have a less than 1 aridity and have a large runoff coefficient (more than 40%). The others lie in dryer territories and have a smaller runoff coefficient. The plots don't lie exactly on the curve. There may be many reasons for that, among them errors of observation or calculation. But there are also physical reasons, related to landscape, geology, climate and vegetation. But this goes a bit beyond the scope of this lecture. If you are interested, there is a lot of literature that you can refer to.

Meteorological factors affecting evaporation

- Energy balance
- Radiation
- Humidity
- Aerodynamic resistance

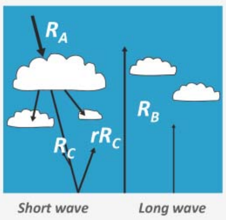
There are 4 meteorological factors that affect evaporation:
The energy balance, driven by incoming solar radiation.
The humidity of the air and the aerodynamic resistance.

Radiation Balance

$$R_N = (1 - r)R_C - R_B$$

R_N : Net short wave radiation [W/m²]
R_C : Incoming short wave radiation
R_B : Outgoing long wave radiation
r : Albedo or whiteness

Surface	Albedo (r)
Open water	0.06
Grass	0.24
Bare soil	0.10 – 0.30
Fresh snow	0.90



We start with the radiation balance. The Net incoming short wave radiation is the amount of short wave radiation entering the top of the atmosphere: RA. Minus the amount that is reflected or absorbed by clouds: Rc. Minus the part that is reflected by the Earth surface (r Rc). Minus the part of the energy that is lost by long wave radiation: RB. The reflectivity is called the albedo or the whiteness. A completely black surface has an albedo of zero. It absorbs all radiation. Snow has an albedo close to 1, reflecting almost all incoming radiation.

Radiometer

We can measure the incoming radiation Rc and the outgoing radiation fluxes r Rc and RB with a Radiometer.

Sunshine Recorder

- R_C can be determined empirically by the theoretical sun hours and n/N
- n/N is the ratio of recorded sun hours to the theoretical (potential) sun hours

Netherlands	$R_C = (0.20 + 0.48\, n/N)R_A$
Average	$R_C = (0.25 + 0.50\, n/N)R_A$
New Delhi	$R_C = (0.31 + 0.60\, n/N)R_A$
Singapore	$R_C = (0.21 + 0.48\, n/N)R_A$

We can also use a more classical instrument to determine the number of hours of sunshine. The Campbell-Stoke sunshine recorder. Depending on where you are on Earth there are different empirical formulas to calculate Rc as a function of the number of sunshine hours and short wave radiation RA.

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

Maximum amount of sun hours per day N

	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												

Short wave radiation expressed in terms of evaporation
RA/λ in kg m-2day-1

The potential number of hours depends on the time of the year and the position on Earth. This is tabulated and so is the short wave radiation occurring above the clouds.

Outgoing Long wave radiation

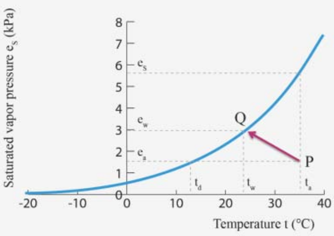
- R_B is calculated through an empirical equation

$$R_B = \underbrace{\sigma(273+t_a)^4}_{\text{left}}\underbrace{\left(0.47-0.21\sqrt{e_a}\right)}_{\text{middle}}\underbrace{\left(0.2+0.8\frac{n}{N}\right)}_{\text{right}}\ [\mathrm{Jd^{-1}m^{-2}}]$$

The outgoing long wave radiation RB is also calculated with a semi-empirical equation, based on the long wave radiation of a warm body. The left part of the equations contains the Stefan Bolzmann coefficient sigma and t_a is the actual temperature near the surface. The right part contains the effect of clouds, which reduce the outgoing radiation like a blanket.
The middle part contains the air humidity (indicate e_a), water vapour being a strong greenhouse gas. So if the air is dry, there is more outgoing long wave radiation. Other greenhouse gasses are not included in this equation. But humidity is, because it is a strong and highly variable greenhouse gas.

Humidity

$$e_s = 0.61\exp\left(\frac{19.9t_a}{273+t_a}\right)$$



$$s = \frac{\mathrm{d}e_s}{\mathrm{d}t} = \frac{5430e_s}{(273+t_a)^2}$$

$$e_a(t_a) = e_s(t_w) - \gamma(t_a - t_w)$$

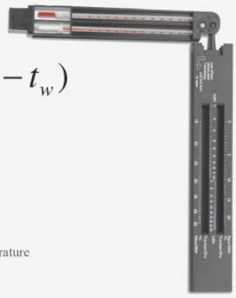
This brings us to the humidity, which not only affects the long wave radiation, but also directly influences evaporation. Let’s first look at the relation for Saturated Vapour Pressure. On the horizontal axis is the temperature and on the vertical axis the pressure at which water vapour is saturated in the air and reaches the point of condensation. This is called the saturation vapour pressure, expressed in kPa or kN/m2. The equation can be derived theoretically from Entropy considerations. The slope of the curve is s. How do we get onto the curve. Imagine we are in the point P with temperature ta and saturation pressure ea, reflecting the number of moisture molecules per unit volume. If we now move left in the figure until we hit the curve, we have reached the dew point td, where the molecules start to condense. We can also move up at the same temperature, but now we are increasing the number of molecules until the pressure is so high that they condense. There also is a spontaneous way to reach the curve, if we have a wet surface, like a wet sponge. If you blow wind over a wet sponge, the fast molecules will leave, the sponge gets cooler, until it reaches the dew point, but at a higher vapour pressure because it has absorbed energy from the surrounding (warmer) air.

Psychrometer

$$e_a(t_a) = e_s(t_w) - \gamma(t_a - t_w)$$

$$h = \frac{e_a(t)}{e_s(t)}$$

t_a is the dry bulb temperature
t_w is the wet bulb temperature
$e_s(t_w)$ is the saturation pressure at the wet bulb temperature
γ is the psychrometer constant (0.066 kPa/°C
h is the relative humidity



This equation shows how it happens. It can be demonstrated by the psychrometer, which has a wet and dry bulb thermometer. We need the psychrometric constant gamma to be able to calculate the actual vapour pressure and hence the relative humidity.

Energy balance

$$\frac{\Delta S_E}{\Delta t} = R_N - H - A - \rho\lambda E \quad [\mathrm{Wm}^{-2}]$$

Assume: $\Delta S/\Delta t = 0,\ A = 0$

On daily basis !!

$$E = \frac{(R_N - H)}{\rho\lambda} = \frac{(1-r)R_C - R_B - H}{\rho\lambda} \quad [\mathrm{m/d}]$$

The energy balance states that the change of energy stored in the ground equals the difference between the net incoming radiation, the sensible heat flux H, the advected heat A and the latent heat of evaporation rho-lambda-E. All energy terms are in W/m2, or J/s/m2. We may assume that average over a day the change of energy stored is negligible and we generally also neglect advection. This results in an expression for the evaporation depending on the balance between net incoming radiation and the sensible heat flux. This is the basis for the Penman equation.

Penman (1948)

- Open water evaporation based on the energy balance,
- but making use of empirical relations
- 4 standard meteorological variables:
  - *air temperature*
  - *relative humidity*
  - *wind velocity*
  - *net radiation*

Besides the energy balance, it makes use of empirical relations. It uses 4 standard meteorological variables: air temperature, relative humidity, wind velocity and net radiation.

Penman Formula

$$E_o = \frac{\left(\frac{sR_N}{\rho\lambda} + \frac{c_p\rho_a}{\rho\lambda}\frac{e_s - e_a}{r_a}\right)}{s+\gamma} \quad [\mathrm{m/d}] \qquad r_a = \frac{245}{(0.54u_2 + 0.5)}\frac{1}{86400}[\mathrm{d/m}]$$

R_N	net radiation at the Earth surface	[J day^{-1} m^{-2}]
λ	heat of evaporation (λ = 2.45 MJ/kg)	[J kg^{-1}]
s	slope of the saturation pressure curve	[kPa K^{-1}]
c_p	specific heat of air (1004 J kg^{-1} K^{-1})	[J kg^{-1} K^{-1}]
ρ_a	density of air (1.205 kg/m3)	[kg m^{-3}]
ρ	density of water (1000 kg/m3)	[kg m^{-3}]
e_a	actual vapour pressure of the air at 2 m elevation	[kPa]
e_s	saturation vapour pressure for the temp. at 2 m elevation	[kPa]
γ	psychrometer constant (γ = 0.066 kPa/°C)	[kPa K^{-1}]
r_a	aerodynamic resistance	[day m^{-1}]

Penman assumed that the sensible heat flux H is correlated to the latent heat flux (evaporation). He could then rework the equation to obtain the equation for open water evaporation Eo. We recognise the symbols that we have already seen: the slope of the saturation pressure curve s. The net incoming radiation RN, the density of water rho the energy required to evaporate a kg of water lambda the specific heat of air, the difference between the saturation and actual vapour pressure and the aerodynamic resistance, which is the driver of the turbulence that exchanges air with air layers higher up.
It exchanges heat and moisture with the atmosphere through diffusive transport.

Penman-Monteith

$$E_a = \frac{\left(\frac{sR_N}{\rho\lambda} + \frac{c_p \rho_a}{\rho\lambda} \frac{e_s - e_a}{r_a} \right)}{s + \gamma\left(1 + \frac{r_c}{r_a}\right)} \quad \text{[m/d]}$$

Monteith expanded the open water equation of Penman to a vegetated area. He introduced the crop resistance rc, which provides a brake on the transpiration of vegetation as a result of environmental constraints.

Crop resistance r_c

- Provides a constraint on the transpiration of vegetation
- Depends on the opening of stomata in leaves, as a function of:
  - *Soil moisture availability*
  - *Relative humidity*
  - *Sunlight*
  - *Temperature*

This crop resistance depends on the opening of stomata in the leaves, which reacts to the availability of moisture in the soil the relative humidity the sunlight for photosynthesis and the temperature. Plants don’t like it when it is too hot or too cold. The details of this go beyond this lecture, but it is not a bad idea to read the paper by Lan Wang who explains this in detail.

Evaporation of the World

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Contrasting roles of interception and transpiration in the hydrological cycle – Part 1: Simple Terrestrial Evaporation to Atmosphere Model

Review Status: This discussion paper is under review for the journal Earth System Dynamics (ESD).

L. Wang-Erlandsson[1,2], R. J. van der Ent[1], L. J. Gordon[2], and H. H. G. Savenije[1]
[1]Department of Water Management, Faculty of Civil Engineering and Geosciences, Delft University of Technology, Delft, the Netherlands
[2]Stockholm Resilience Centre, Stockholm University, Stockholm, Sweden

Abstract. Terrestrial evaporation consists of biophysical (i.e., transpiration) and physical fluxes (i.e., interception, soil moisture, and open water). The partitioning between them depends on both climate and the land surface, and determines the time scale of evaporation. However, few land-surface models have analysed and evaluated evaporative partitioning based on land use, and no studies have examined their subsequent paths in the atmosphere. This paper constitutes the first of two companion papers that investigate the contrasting effects of interception and transpiration in the hydrological cycle. Here, we present STEAM (Simple Terrestrial Evaporation to Atmosphere Model) used to produce partitioned evaporation and analyse the characteristics of different evaporation fluxes on land. STEAM represents 19 land-use types (including irrigated land) at sub-grid level with a limited set of parameters, and includes phenology and stress functions to respond to changes in climate conditions. Using ERA-Interim reanalysis forcing for the years 1999–2008, STEAM estimates a mean global terrestrial evaporation of 73 800 km^3 $year^{-1}$, with a transpiration ratio of 59%. We show that the terrestrial residence time scale of transpiration (days to months) has larger inter-seasonal variation and is substantially longer than that of interception (hours). Furthermore, results from an offline land-use change experiment illustrate that land-use change may lead to significant changes in evaporative partitioning even when total evaporation remains similar. In agreement with previous research, our simulations suggest that the vegetation's ability to transpire by retaining and accessing soil moisture at greater depth is critical for sustained evaporation during the dry season. Despite a relatively simple model structure, validation shows that STEAM produces realistic evaporative partitioning and hydrological fluxes that compare well with other global estimates over different locations, seasons and land-use types. We conclude that the simulated evaporation partitioning by STEAM is useful for understanding the links between land use and water resources, and can with benefit be employed for atmospheric moisture tracking.

Here you see his paper, it's open access.

Direct measurement of evaporation

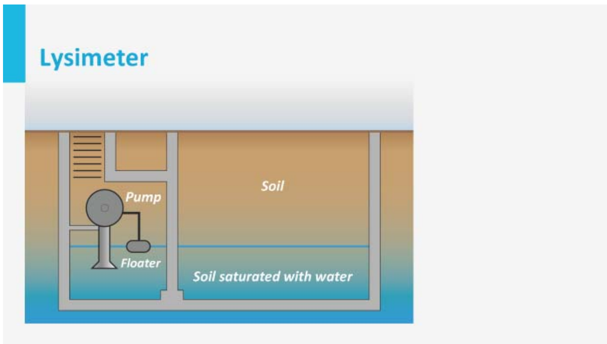
- Water balance: $E = P - \frac{Q}{A} - \frac{dS}{dt}$ [L/T]
- Evaporation pan
- Lysimeter
- Shallow Lysimeter

There are different ways in which components of evaporation can be measured directly: At catchment scale, it can be done on basis of the water balance: E=P-Q- dS/dt. But there are also instruments, such as the evaporation Pan, the Lysimeter or the shallow lysimeter.

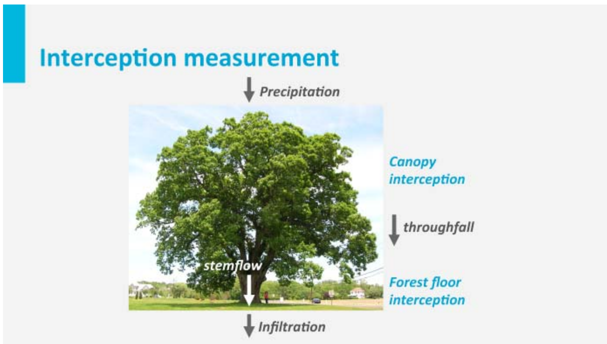
Pan evaporation

$$E_0 = k_{pan} E_{pan} \quad \text{[mm/d]}$$

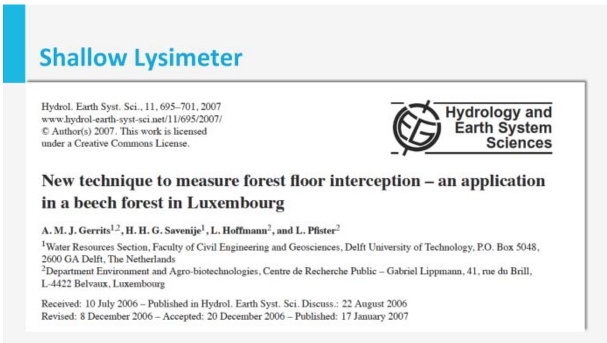
The pan is the most simple way of measuring the open water evaporation as a function of the atmospheric conditions. Through an empirical equation the open water evaporation can be derived form the observed pan evaporation. But open water evaporation is not the same as the evaporation of a vegetated surface.



The lysimeter (these can be very large things ranging from 5 to 100 m2 or even more) is a way of calculating the total evaporation of a vegetated surface by weighing the storage and of course tracking all the water that goes in, or is removed.



For the measuring of interception we have developed our own instruments. One of our students developed a devise to weigh a tree.



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Hydrology and Earth System Sciences

New technique to measure forest floor interception – an application in a beech forest in Luxembourg

A. M. J. Gerrits[1,2], H. H. G. Savenije[1], L. Hoffmann[2], and L. Pfister[2]

[1]Water Resources Section, Faculty of Civil Engineering and Geosciences, Delft University of Technology, P.O. Box 5048, 2600 GA Delft, The Netherlands
[2]Department Environment and Agro-biotechnologies, Centre de Recherche Public – Gabriel Lippmann, 41, rue du Brill, L-4422 Belvaux, Luxembourg

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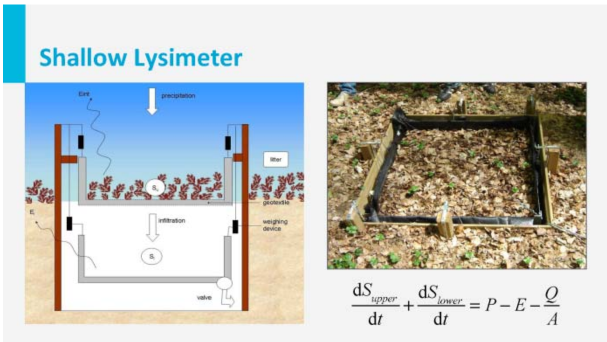
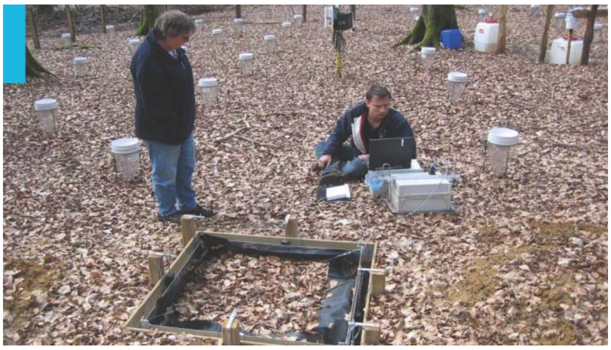
And Dr Coenders developed a shallow lysimeter to determine the ground interception, which is published in this open access journal.



It has two containers on top of each other which are continuously weighed. The top one is permeable the lower one has a tap.



From the water balance we can compute the evaporation.

The evaporation tower

Hydrol. Earth Syst. Sci., 18, 2021–2032, 2014
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Hydrology and Earth System Sciences

A new method to measure Bowen ratios using high-resolution vertical dry and wet bulb temperature profiles

T. Euser[1], W. M. J. Luxemburg[1], C. S. Everson[2], M. G. Mengistu[2], A. D. Clulow[2], and W. G. M. Bastiaanssen[1]

[1]Delft University of Technology, Water Resources Section, Stevinweg 1, 2628 CN Delft, the Netherlands
[2]Centre for Water Resources Research, School of Agriculture Earth and Environmental Sciences, University of KwaZulu-Natal, Private Bag X01, Scottsville, 3209 Pietermaritzburg, South Africa

Correspondence to: T. Euser (t.euser@tudelft.nl)

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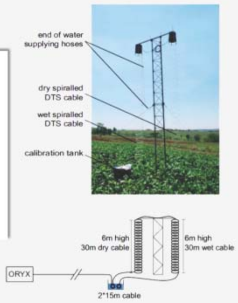

More advanced observations of evaporation are done by eddy covariance on a tower, or by an instrument we developed ourselves: The DTS based wet and dry cable approach, which is also published in open access.

Further reading

Mohamed, Y. A., van den Hurk, B. J. J. M., Savenije, H. H. G., and Bastiaanssen, W. G. M., 2005. Hydroclimatology of the Nile: Results from a regional climate model. *Hydrol. and Earth Syst. Sc.*, 9: 263-27. http://www.hydrol-earth-syst-sci.net/9/263/2005/hess-9-263-2005.html

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Euser, T., W. M. J. Luxemburg, C. S. Everson, M. G. Mengistu, A. D. Clulow, and W. G. M. Bastiaanssen, 2014. A new method to measure Bowen ratios using high-resolution vertical dry and wet bulb temperature profiles, *Hydrol. Earth Syst. Sci.*, 18, 2021-2032. http://www.hydrol-earth-syst-sci.net/18/2021/2014/hess-18-2021-2014.html

You can see that evaporation has many different aspects and that there still is a lot to investigate before we fully understand. All the intricate ways in which water changes phase (and face) in the hydrological cycle. But being difficult to grasp does not mean that it is unimportant. It still is the largest flux resulting from terrestrial precipitation. Which reminds me: we have not yet discussed precipitation much. I'll see you in the next module, about precipitation.

GWC 2: Evaporation

CTB3300WCx: Introduction to Water and Climate

Prof.dr.ir. Hubert H.G. Savenije

TU Delft

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