

# Groundwater Recharge

Irina Engelhardt

## Glossary

Recharge: Water which percolates down through the unsaturated zone and enters the dynamic ground-water flow system, ***thereby contributing to replenishment of the groundwater reservoir.***

Recharge area: Area receiving water from infiltration and percolation. Flow of water is directed **downward away from the water table**. The recharge area encompasses most of the upper free surface of the groundwater reservoir.

Discharge area: The area of a groundwater basin in which the groundwater flow is **directed toward the water table**.

## Groundwater Flow Net with Recharge and Discharge Area

Topographically controlled flow pattern:

- Water table is higher under topographically high areas and lower under topographically low areas
- Water table relief is less marked than the relief of the ground surface
  - Works best in areas with plenty of recharge
  - In arid areas and strongly heterogeneous areas, water table does often not resemble the ground surface

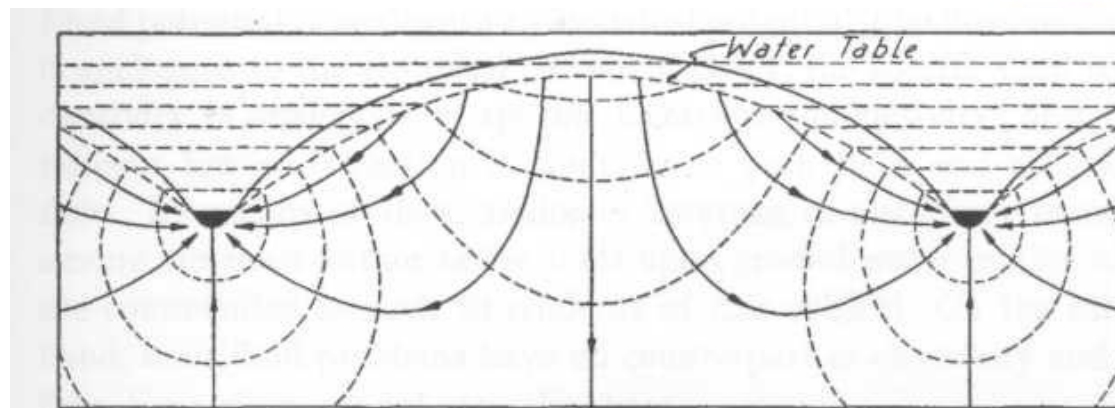
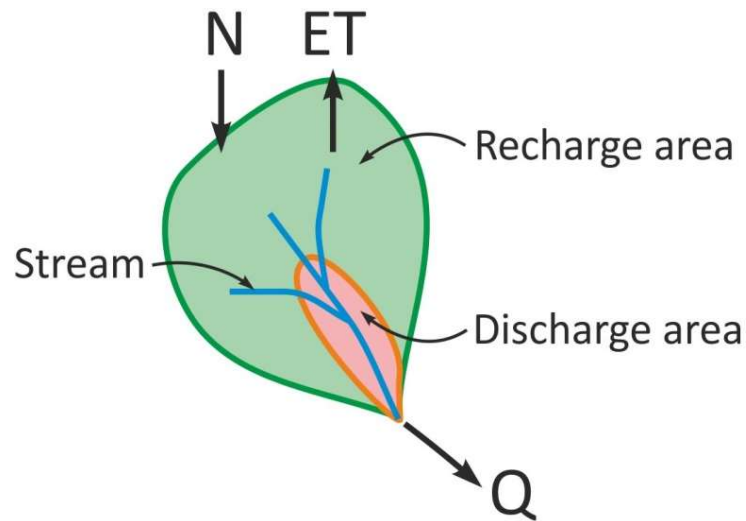


FIG. 45.—Approximate flow pattern in uniformly permeable material between the sources distributed over the air-water interface and the valley sinks.

(Hubbert 1940)

# Recharge as Component of the Water Balance



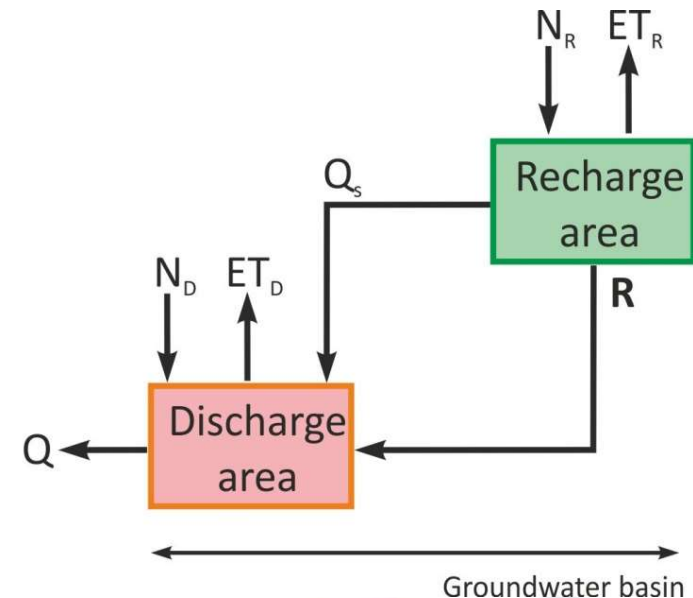
$$N = ET + Q \pm \Delta S$$

N = precipitation

ET = evapotranspiration

Q = discharge

$\Delta S$  = change in storage



$$R = N_R - ET_R - Q_s$$

R = recharge to ground water which equals D (groundwater discharge)

Q<sub>s</sub> = surface runoff

ET = evapotranspiration

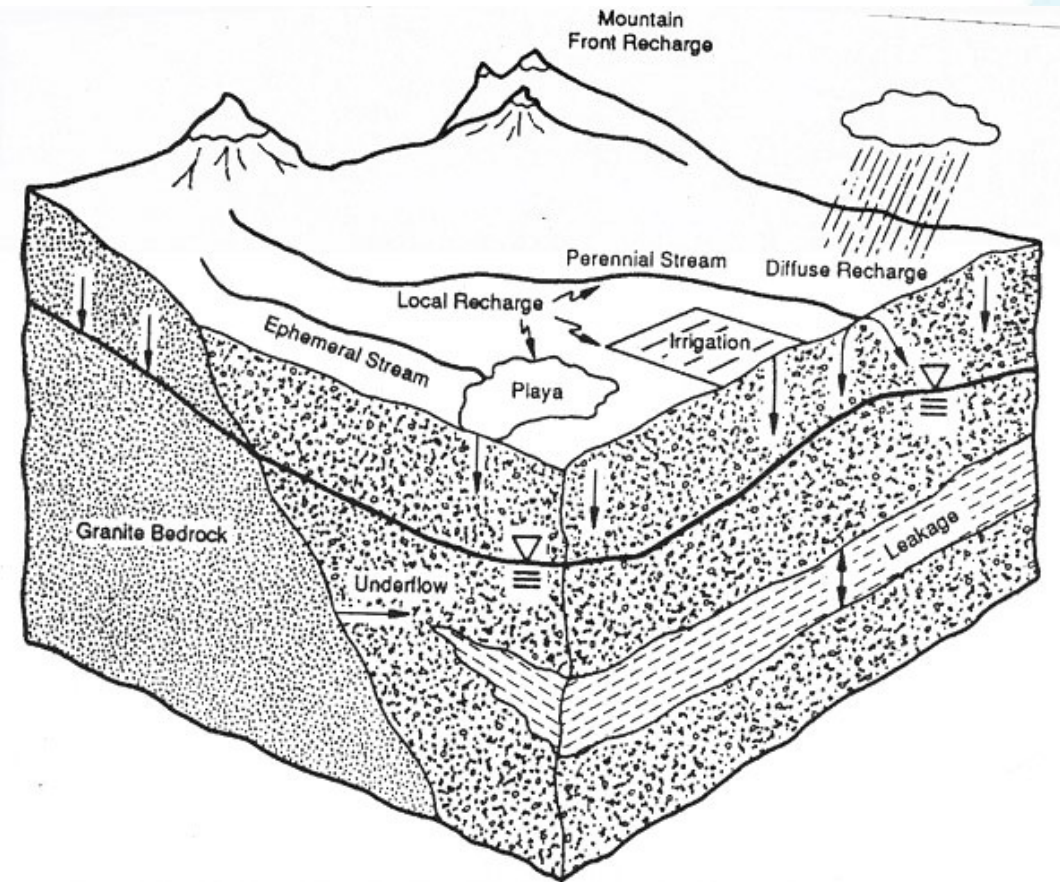
(Modified from Freeze & Cherry 1979)

# Concepts of groundwater recharge

Recharge of groundwater may occur from: precipitation, rivers, canals, lakes, irrigation, urbanization

**Direct recharge:** water added to the groundwater reservoir in excess of soil moisture and evapotranspiration deficits, by **direct vertical percolation** of water.

**Indirect recharge:** results from **percolation** to the water table **following runoff** and localization in joints, as **ponding** in low lying areas and lakes, or **through beds** of surface water courses.



(Lerner et al. 1990)

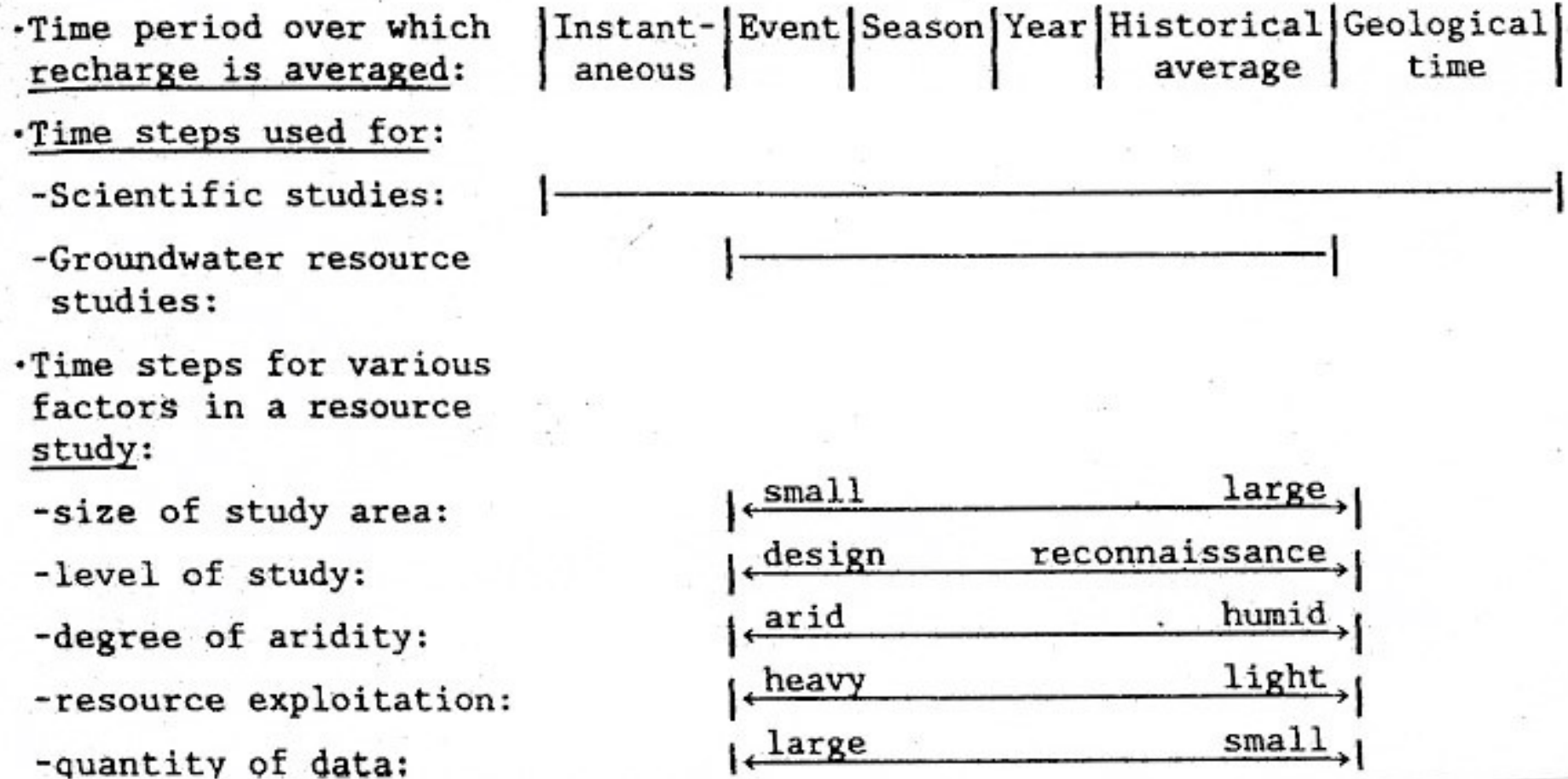
(Stephens 1996)

# Groundwater Recharge

In which regions do you expect local recharge to be a major component to refresh the groundwater resources?



# Groundwater Recharge – Time Scales



(Lerner et al. 1990)

## Guidelines to Determine Groundwater Recharge

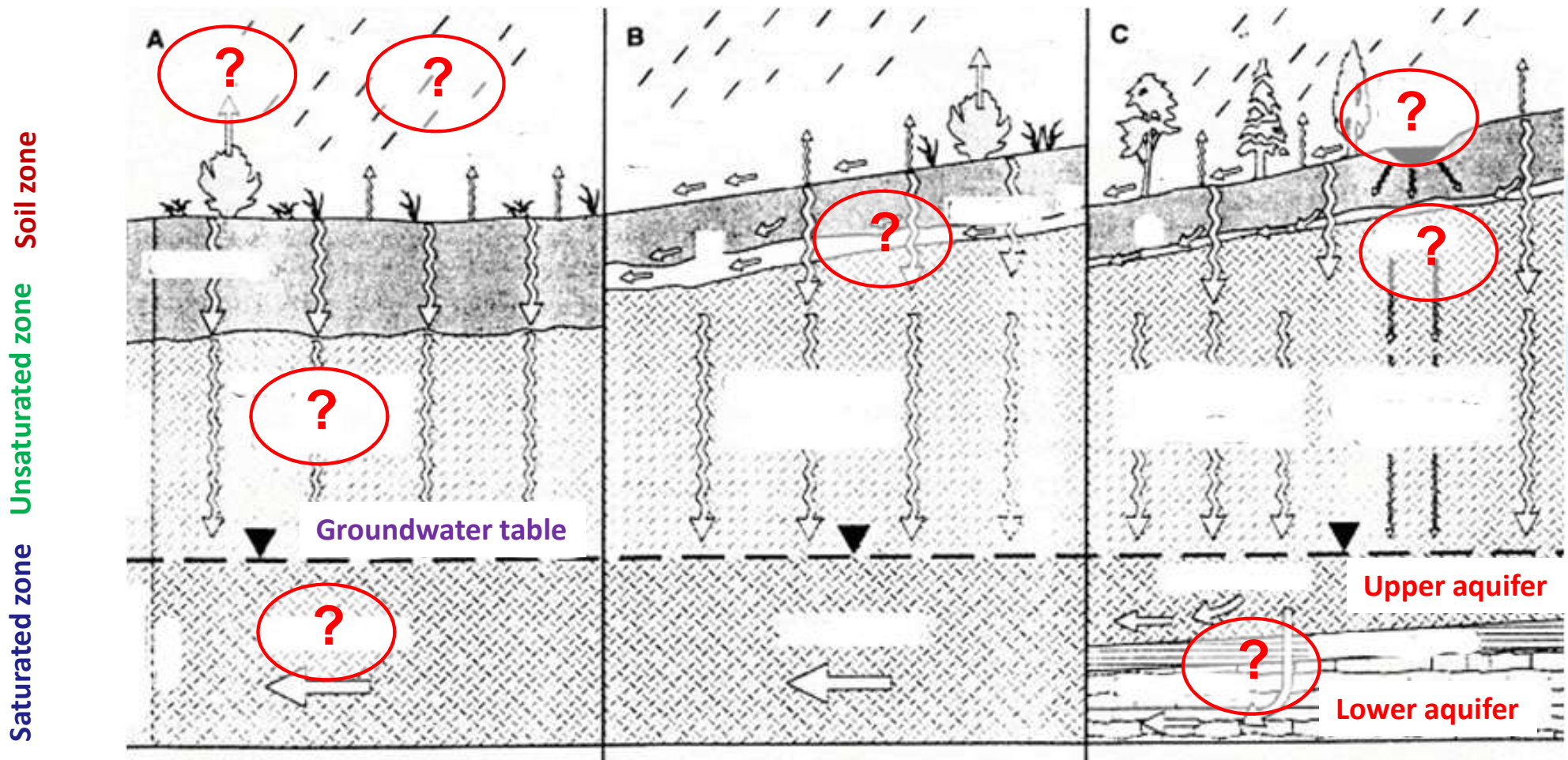
- **How much groundwater** can an aquifer **store**? (a "full " aquifer cannot take up more water).
- **How much water percolates** through the unsaturated zone per time? (potentially high groundwater recharge may never happen because the hydraulic conductivities of the geological layers are too small, e.g. loess)
- Where does **the excess water flow** to in cases of potentially high groundwater recharge? (e.g. interflow )
- Relationship between **potential and actual** groundwater **recharge**?
- What are the results from **other methods to determine groundwater recharge**? (If possible, apply more than one method)



## Errors and Inaccuracies Effecting Groundwater Recharge Calculations

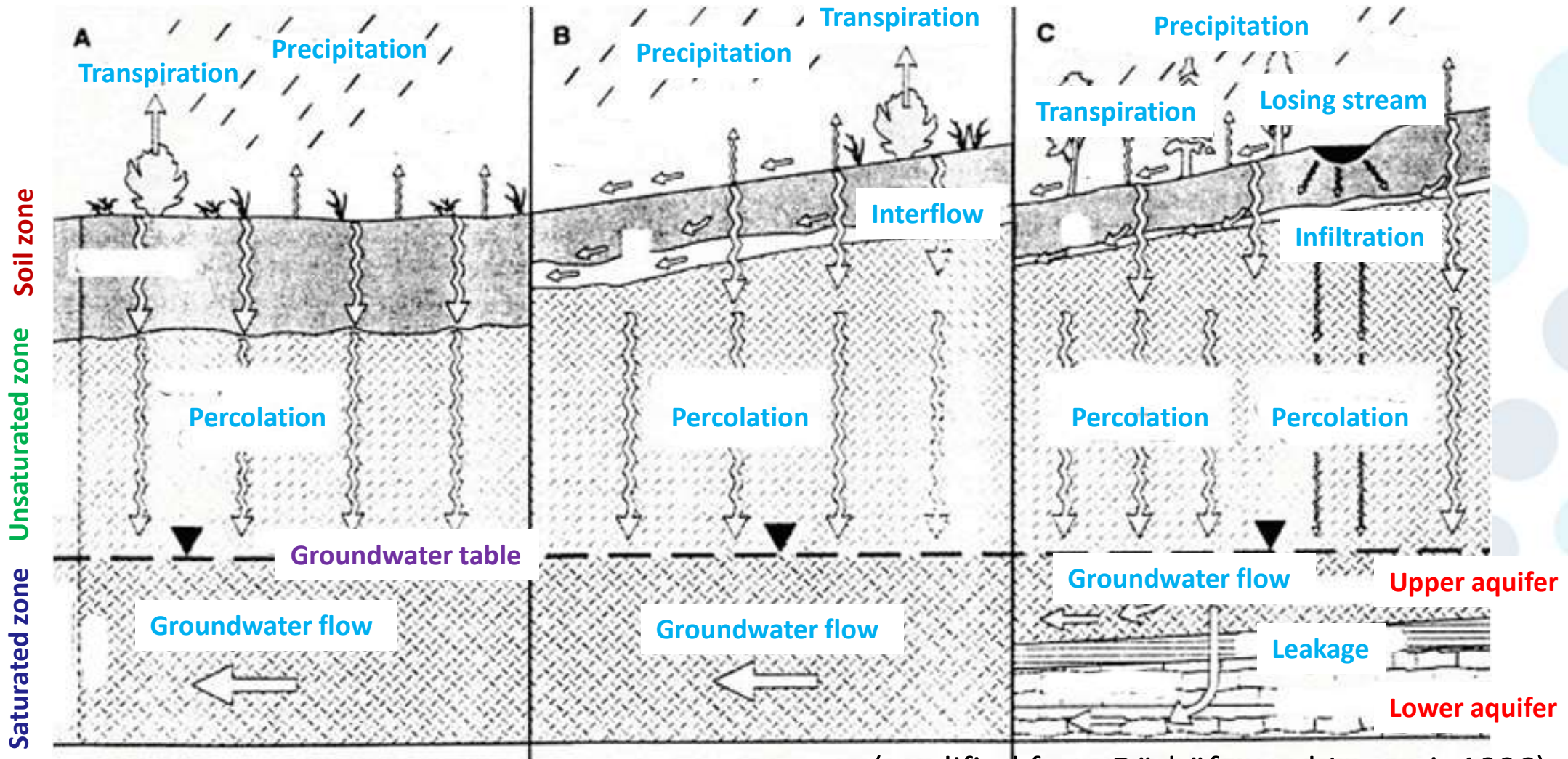
- **Conceptual model:** A wrong **conceptual model** is the most important error. It is the result of **misconceptualisation** of the groundwater recharge process, or the process has been oversimplified.
- **Temporal and spatial interpretation:** Most of the groundwater recharge **processes** are **non-linear with respect to time**. Errors are generated when monthly, annual, or long-term data series are used, **to quantify shorter time periods**.
- **Errors associated with measurement:** Those **errors are created by the measuring device itself**.
- **Calculation errors.**

# Conceptual Models of Groundwater Recharge in Temperate Humid Climate



(modified from Dörhöfer and Josopait 1996)

# Conceptual Models of Groundwater Recharge in Temperate Humid Climate



(modified from Dörhöfer and Josopait 1996)

# Methods for Estimating Groundwater Recharge

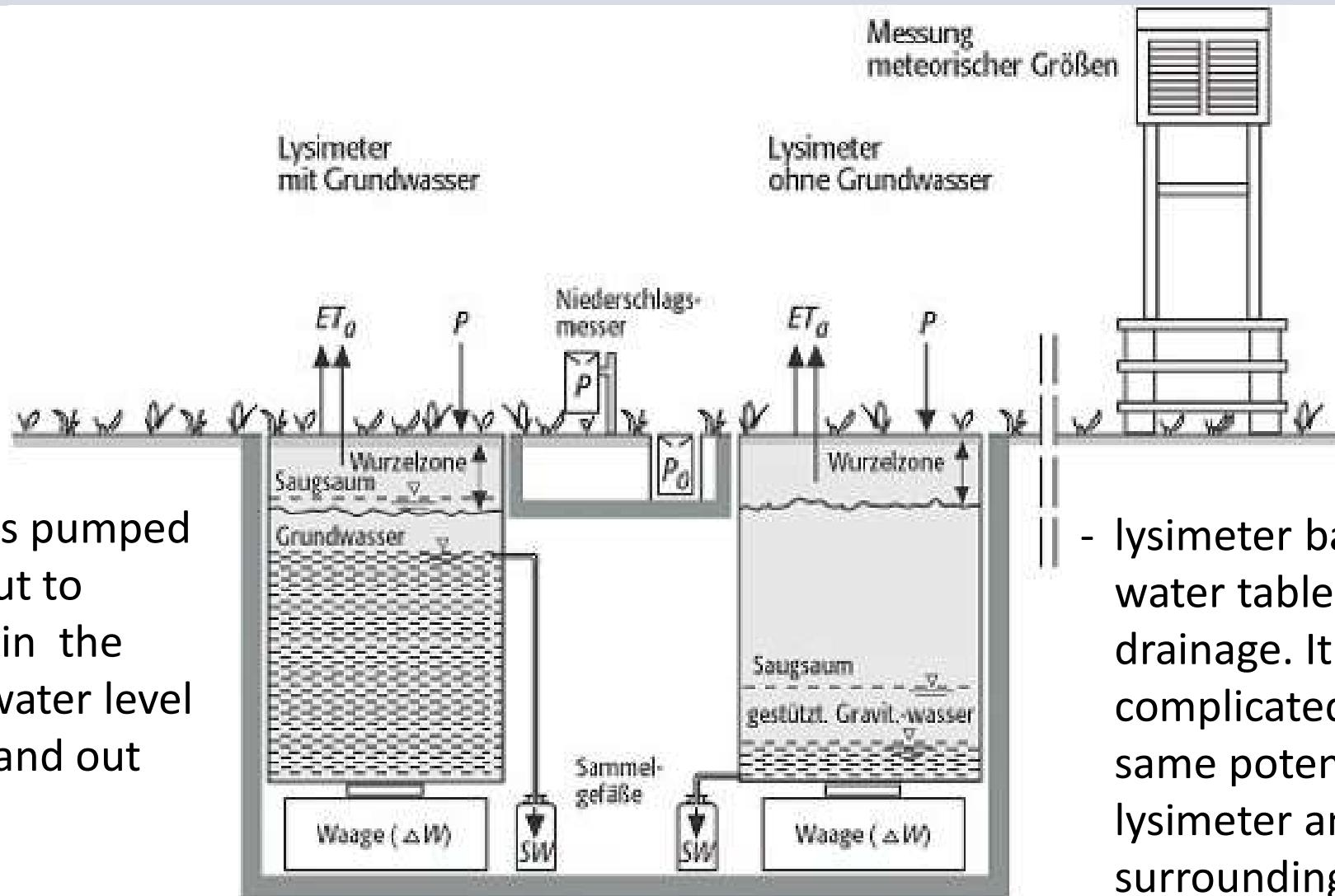
1. **Direct** measurements
2. **Water balance** calculations
3. **Darcy** approaches
4. Deconvolution of **hydrographs**
5. **Tracer** techniques
6. Other approaches, mainly **empirical methods**

# Direct Measurement of Recharge with Lysimeters

- requires expensive construction
- accurate
- point measurement
- do not account for loss by interflow
- upscaling to catchment area difficult



# Types of Lysimeters



- water is pumped in or out to maintain the same water level inside and out

- lysimeter base above the water table with suction drainage. It is relatively complicated to have the same potentials in the lysimeter and in the surrounding soil.

# Lysimeter Data



What must be measured in a lysimeter to calculate recharge?

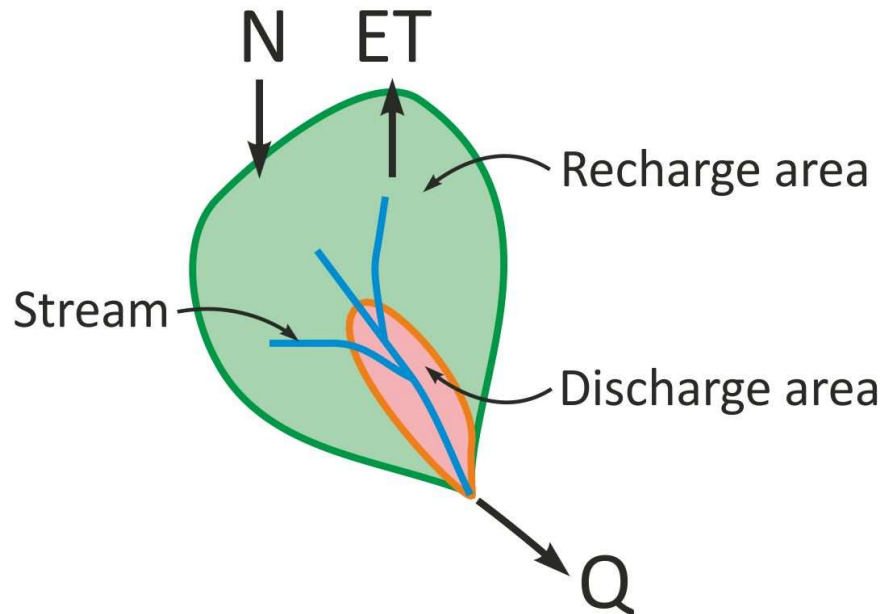


## Requirements of a lysimeter for estimating recharge

- **undisturbed** soil
- large enough to **minimize edge** effects
- large and deep enough to enclose complete **root systems**
- surrounded by **similar vegetation**
- surrounded by the **similar hydraulic conditions**
- watertight, except for the drainage to be measured



# Estimation of Recharge by Water Balance Approaches



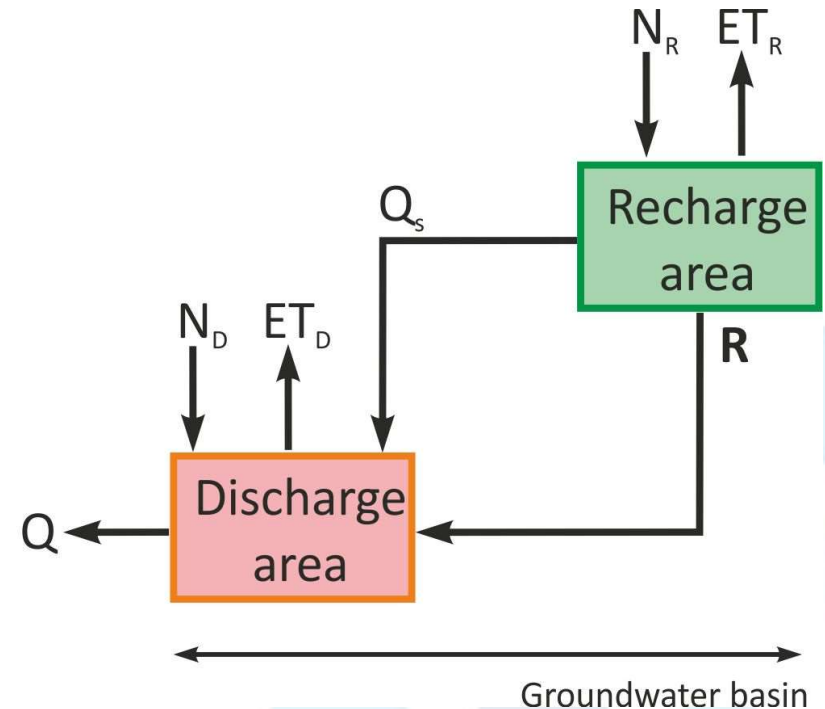
$$N = ET + Q \pm \Delta S$$

$N$  = precipitation

$ET$  = evapotranspiration,

$Q$  = discharge

$\Delta S$  = change in storage



$$R = N_R - ET_R - Q_s$$

$R$  = recharge to ground water

$Q_s$  = surface runoff

$ET$  = evapotranspiration

(Modified from Freeze & Cherry 1979)

# Precipitation

- includes various forms of water, that fall to the ground surface from the atmosphere
- highly variable in space and time
- examples: rain, drizzle, snow, ice crystals

## Precipitation is characterized by:

- precipitation depth  $h_N$  [mm]
- precipitation duration  $T_N$
- spatial distribution
- frequency  $n$ , annual occurrence  $T_n = 1/n$
- temporal change of intensity  $i_N(t)$
- type of precipitation



Source: ARD2010

# Measuring of Precipitation with a Weather Radar



Source: ARD2011

# Measurement of Precipitation

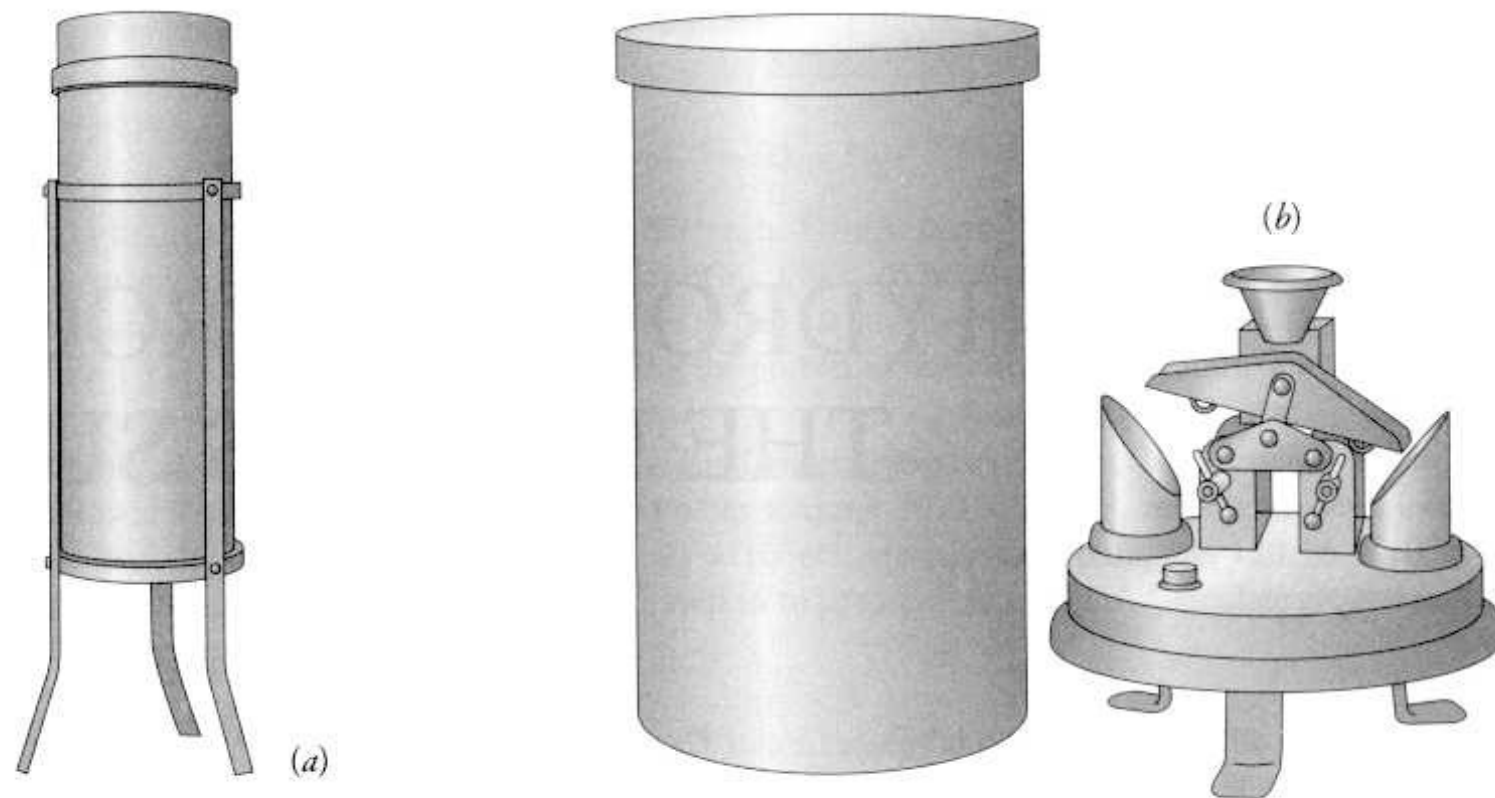
## Local measurement with rain gauges

- Simplest instruments are **containers** that collect precipitation through the storm event
- More **sophisticated rain gauges** record the time, duration and intensity of precipitation
- **Accuracy** of measurement is affected by the physical setting and by disturbances
- **Snow** measurements typically **underestimate** the actual precipitation, particularly at high wind speeds
- Precipitation data are available from **regional networks** (for example DWD, USGS)

## Spatial measurement

- Weather **radar**

## Rain gauges



**Figure 2.1** Examples of gages used for measuring precipitation. Panel (a) shows an 8-inch standard U.S. rain gage. The collection funnel inside directs the rain to a plastic measurement tube. Panel (b) shows a tipping bucket recording rain gage. The buckets are calibrated to tip after 0.01 in. of rainfall and are recorded by a data system.

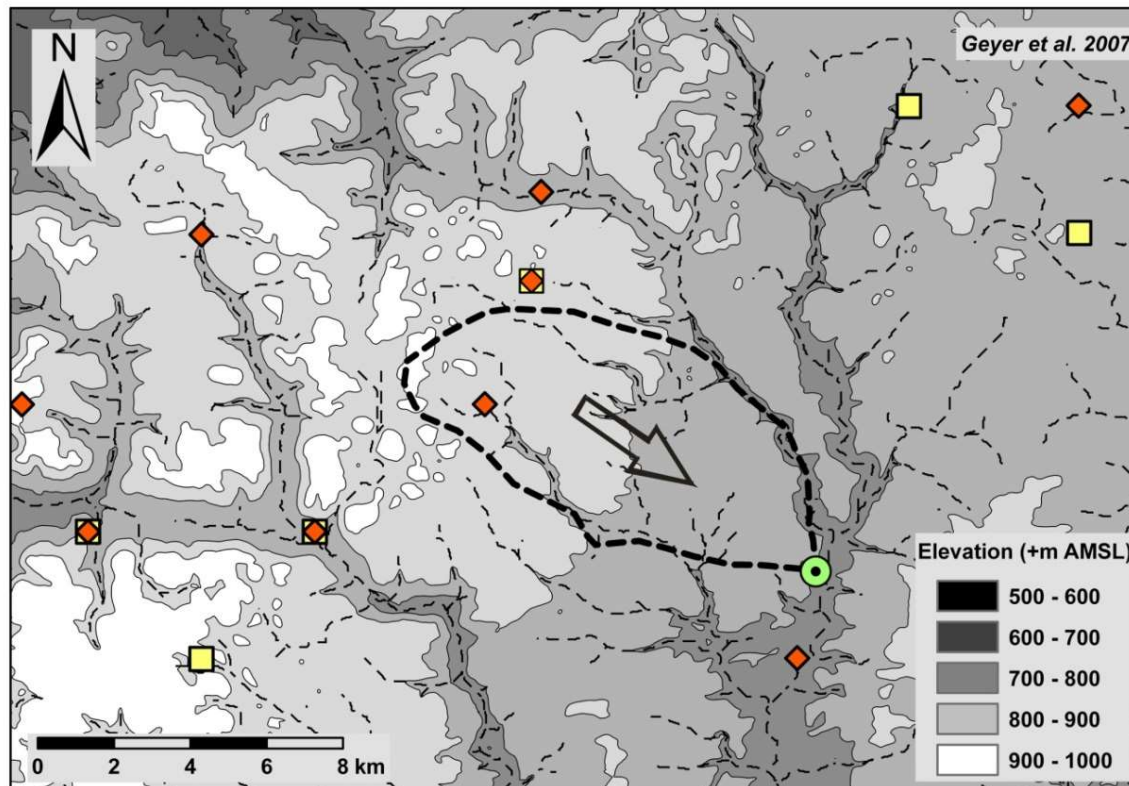
(Schwartz and Zhang 2003)

# Examples of Rain Gauges and Weather Stations



Fotos: Sebastian Schmidt

# Weather Station Networks



- Spring Gallusquelle
- Ground water flow direction
- Rain gauge
- Ground water catchment
- Valley system
- Weather station

## Approaches for areal estimates of precipitation from networks

Arithmetic average

$$P_a = \frac{1}{N} \sum_{i=1}^N P_i$$

Area-weighted average  
(Thiessen-weighted average)

$$P_a = \frac{\sum_{i=1}^N P_i a_i}{\sum_{i=1}^N a_i}$$

$P_a$  = average precipitation for the area

$P_i$  = precipitation at station  $i$

$N$  = number of weather stations

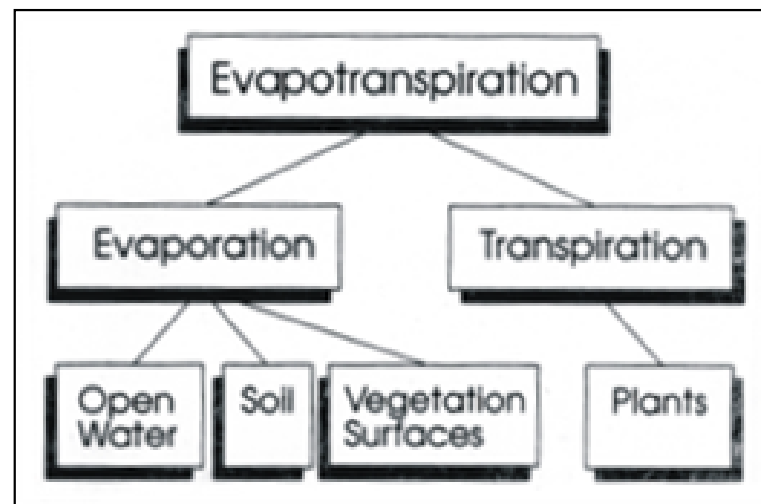
$a_i$  = area of the polygon around station  $i$

# Evapotranspiration

Evapotranspiration ETP = Evaporation E + Transpiration T                      unit [L/T] or (mm/a)

**Evaporation:** physical process by which liquid water is converted into water vapor and removed from the evaporating surface. The rate is controlled by the availability of energy at the evaporating surface and how water can diffuse into the atmosphere.

***Evaporation refers to the quantity of water loss from soils, rivers, and lakes.***



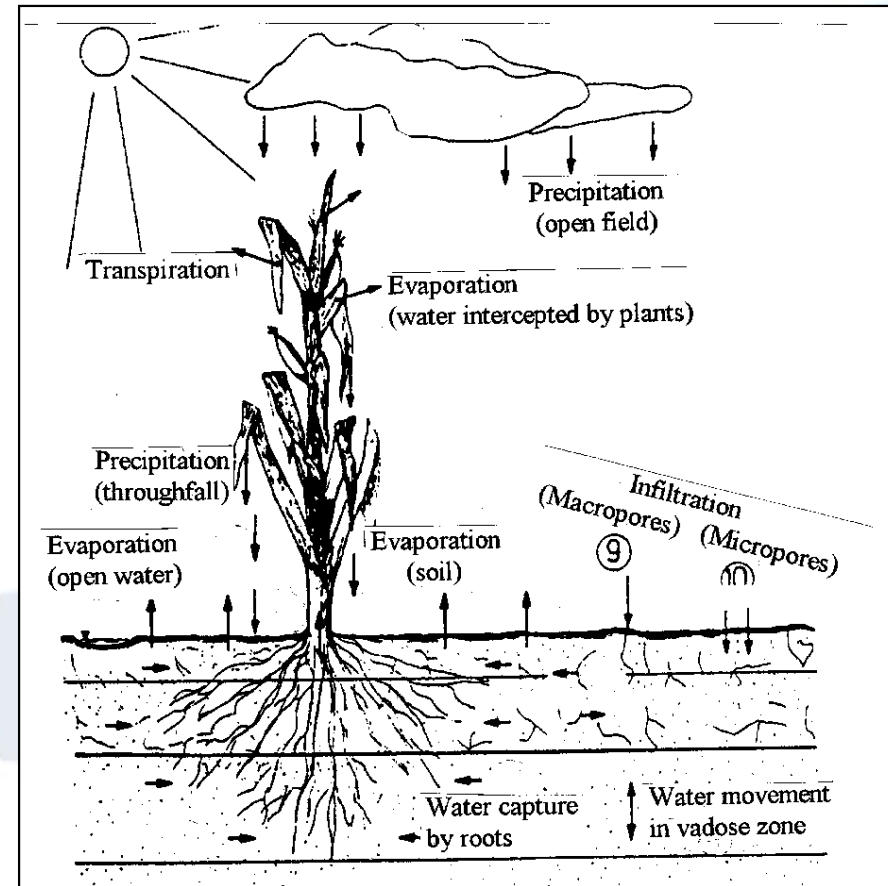
Ward and Elliot, 1995



# Evapotranspiration

Transpiration: "process by which moisture is carried through plants from roots to small pores on the underside of leaves, where it changes to vapor and is released to the atmosphere. Transpiration is essentially evaporation of water from **plant leaves**."

Studies have revealed that transpiration accounts for about 10 percent of the moisture in the atmosphere, with oceans, seas, and other bodies of water (lakes, rivers, streams) providing nearly 90 percent, and a tiny amount coming from sublimation (ice changing into water vapor without first becoming liquid). " (source: USGS)



Bronstert and Jürgens, 1995

# Potential and Actual Evapotranspiration

Potential evapotranspiration: assumes that the ET flux will not exceed the available energy and sufficient water is available , i.e. the potential ETP = maximum ETP.



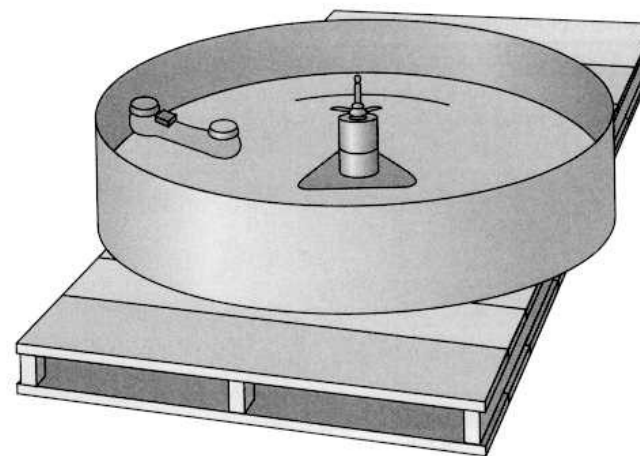
Actual evapotranspiration: Evapotranspiration under field conditions, for example, when not enough water is available from soil moisture.

# Measurement of Evaporation

Measurement of potential evaporation  $E_p$  with an evaporation pan

- shallow tank that contains water
- water depth is measured daily
- water is added or removed to adjust for evaporation losses
- in general, evaporation from a pan is higher than evaporation from a lake because the water can heat up by the metal sides considerably
- pan evaporation  $E_{pan}$  is corrected with an empirical pan coefficient  $C_{et}$  that varies depending on the month

$$E_p = E_{pan} \times C_{et}$$



**Figure 2.4** Sketch showing the U.S. Weather Service Class A evaporation pan. It has a diameter of 47.5 in. and is 10 in. deep. Mounted on the interior are a stilling basin, a hook gage for accurate measurements, and a min/max thermometer.

(Schwartz and Zhang 2003)

# Determination of Evaporation from Open Water Surfaces

## Dalton equation

$$E = f(u) \times (e_s - e)$$

$E$  = evaporation from a water surface

$e_s$  = saturation water pressure of the surface

$e$  = actual vapor pressure

$f(u)$  = function of wind speed in a given height

## Mass transfer equation

## Penman (1948)

$$E = \frac{s \times \frac{R_n}{L} + \gamma \times f(v) \times (e_s - e)}{s + \gamma}$$

$s$  = slope of the saturated vapor pressure curve

$R_n$  = net radiation

$L$  = latent heat of vaporization

$\gamma$  = psychrometric constant

Combination of energy balance equation with the Dalton equation

# Potential Evapotranspiration (ETp) – Calculation

## Formula by Thorntwaite

Thorntwaite and Wilen (1948) developed an equation to calculate the potential evapotranspiration **from grassland** that is based on the temperature.

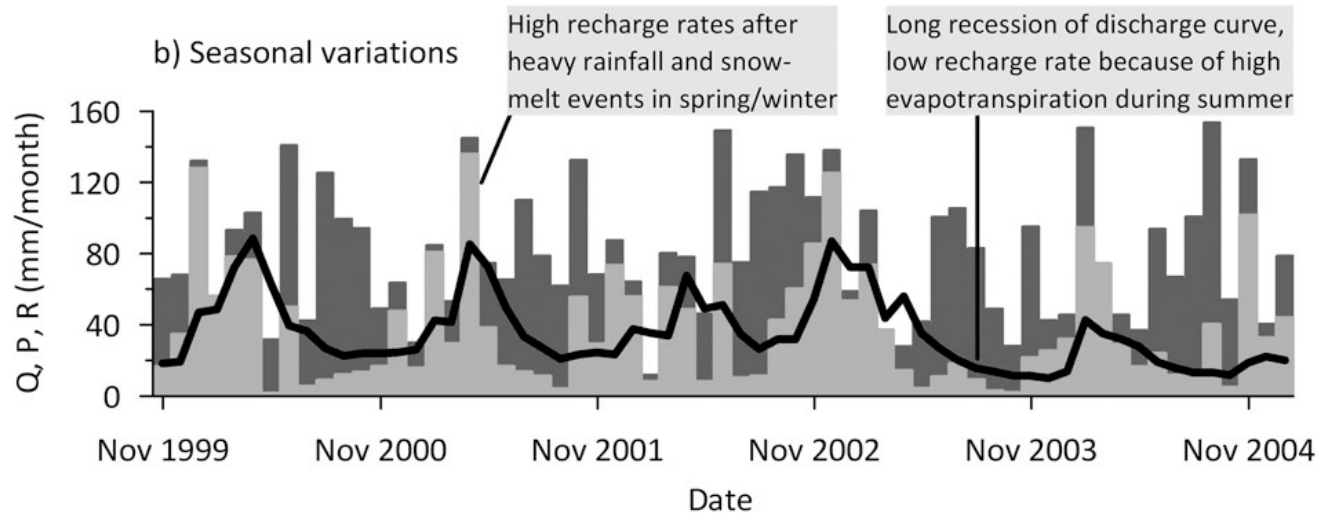
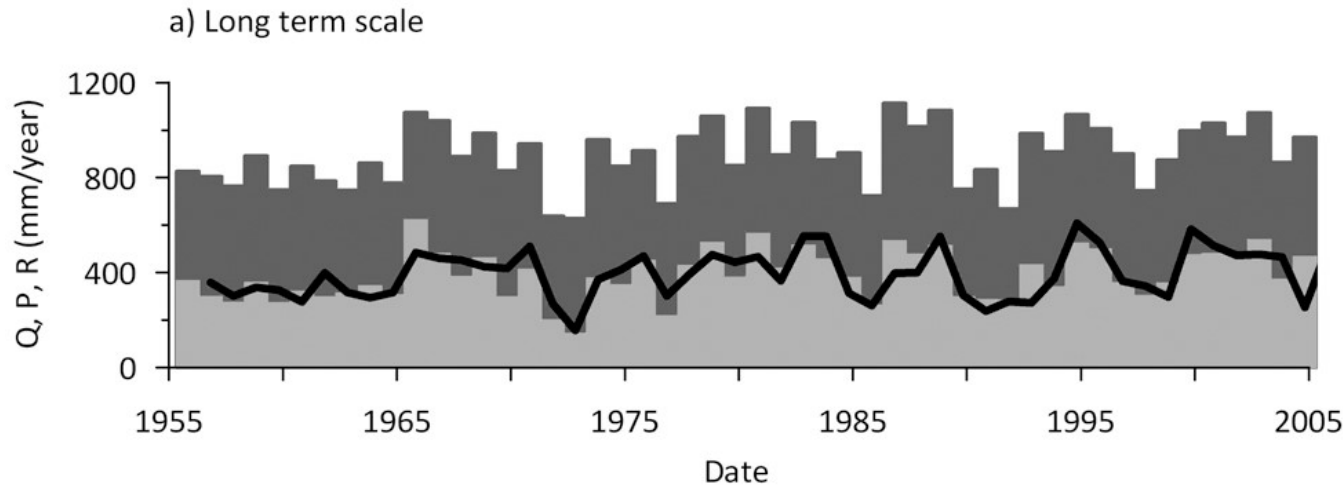
$$ET_p = k_1 k_m (10 T_M / I)^a$$

$$I = \sum_{M=1}^{12} (T_m / 5)^{1.514}$$

- $ET_p$  = **monthly** potential evapotranspiration (cm)
- $k_1$  = 16 mm/month
- $k_m$  = monthly correction factor, function of the latitude
- $T_m$  = mean monthly air temperature (° C)
- $I$  = annual heat index, the sum of the 12 monthly values of  $(T_m/5)^{1.514}$

$$a = 6.75 * 10^{-7} I^3 - 7.71 * 10^{-5} I^2 + 1.79 * 10^{-2} I + 0.49$$

# Example – water balance approach on different time scales



$N > E_{tpot}$ :  $E_{Ta} = E_{Tpot}$   
 Large soil moisture:  $E_{Ta} = E_{Tpot}$

Precipitation (P)    
  Recharge (R)    
  Spring discharge (Q)

# Groundwater recharge, methods, discharge methods

## Darcy- approach

If hydraulic potentials, hydraulic conductivities and boundary fluxes are known, groundwater recharge can be calculated by applying Darcy's law. Two approaches exist:

- a) the field measurement and
- b) numerical modeling.

A simple approach for homogeneous conditions and unconfined aquifers is presented.

## Groundwater recharge: Tracer methods

**Isotopes** and **chemical tracers** are commonly used to determine groundwater recharge in arid and semi-arid climate. They can be:

- *natural tracers*, i.e. tracers that are already in the geosphere
- *man-made tracers*, i.e. tracers of anthropogenic origin.

In general, tracers can be used to follow the travel path of a water molecule.

Application of a mass balance approach, the amount of water can be quantified for a given streamline (i.e. flow path).

### **Requirements:**

- conservative behaviour* of tracers, i.e. no interaction of the tracer with the solid phase of the aquifer,
- a *contrast*, i.e. differences of tracers in the different components,
- knowledge of concentrations* in the respective component.



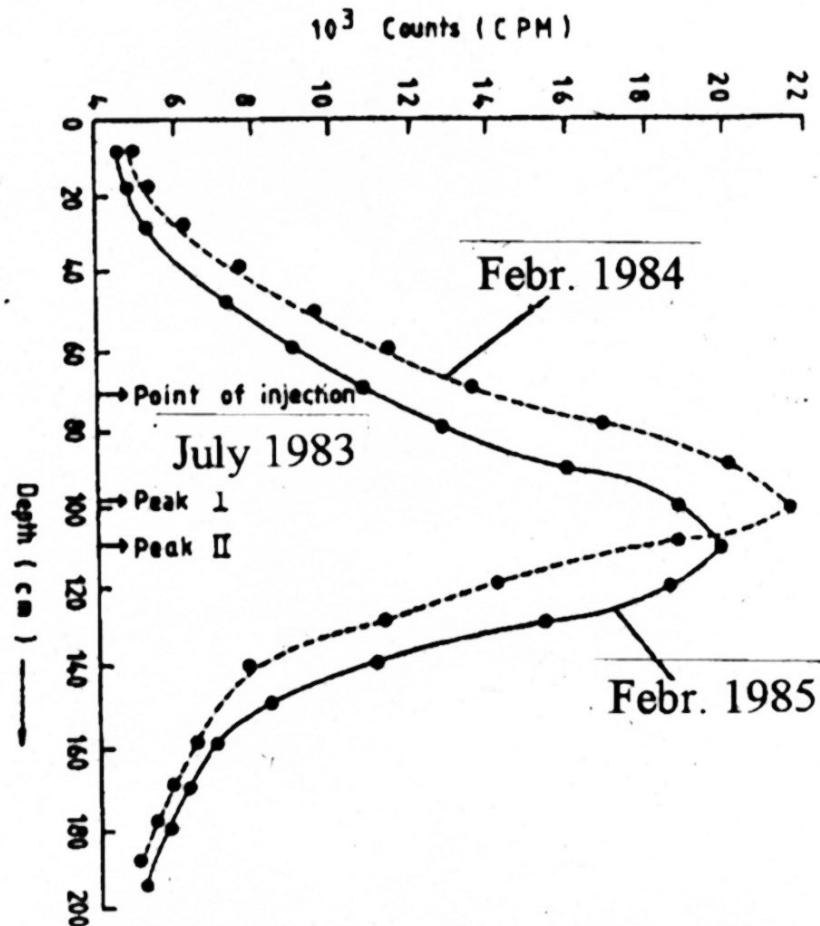
## Groundwater recharge: Tracer methods

Tracers can be used in different ways :

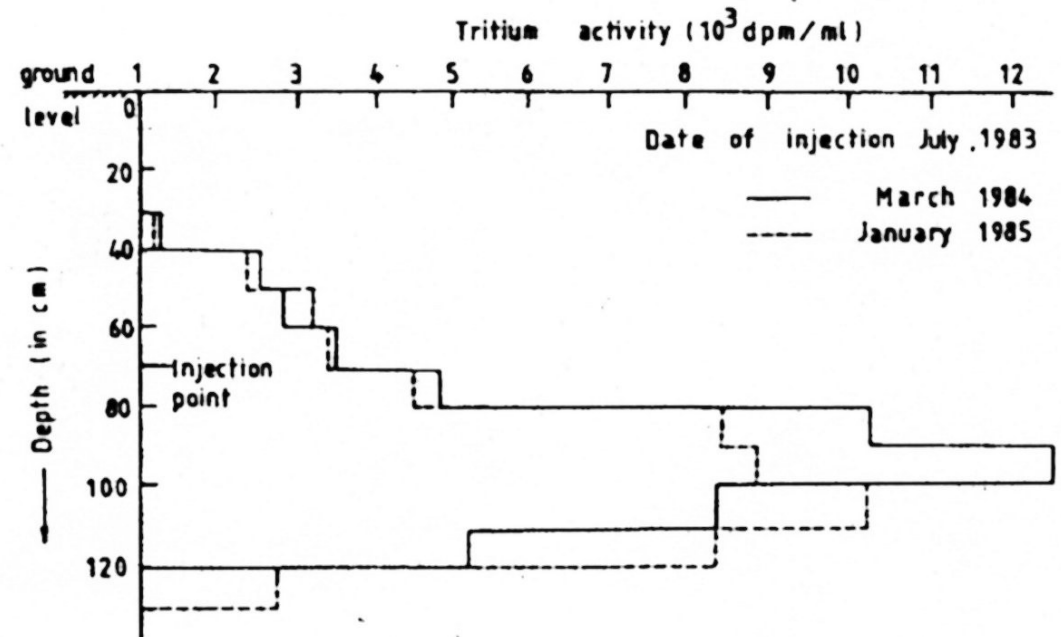
1. *Tracer method*: Tracer is added to a certain amount of water, and the concentration change of the tracer is observed in time (bomb tritium, artificially produced tracer). That method is usually applied to the unsaturated zone.
2. *Discharge method*: This method calculates fluxes of tracer and water for both the saturated and the unsaturated zone. Natural tracers are commonly used.
3. *Calculation of mean residence time*: This method is applied on the catchment area scale. Radioactive isotopes are commonly used.

# Groundwater recharge: Tracer methods

$^{60}\text{Co}$  application



Tritium application



Tracer- profiles (Chandrasekharan, 1988)

# Simulation variabel gesättigter Strömung

# Flow in the Unsaturated Zone

The quantity of water in a partly saturated zone is defined as volumetric water content  $\theta$  with:

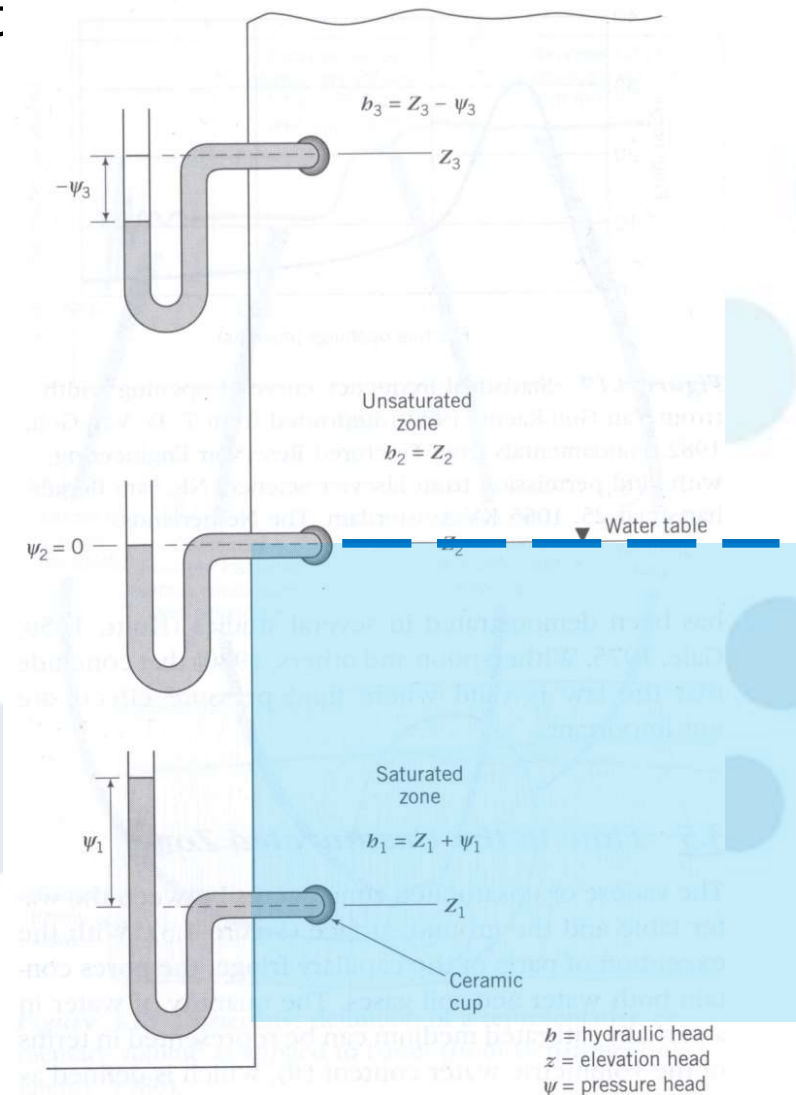
$$\theta = \frac{V_W}{V_T}$$

If the porous medium is fully saturated the volumetric water content equals the porosity.

$$0 \leq \theta \leq n$$

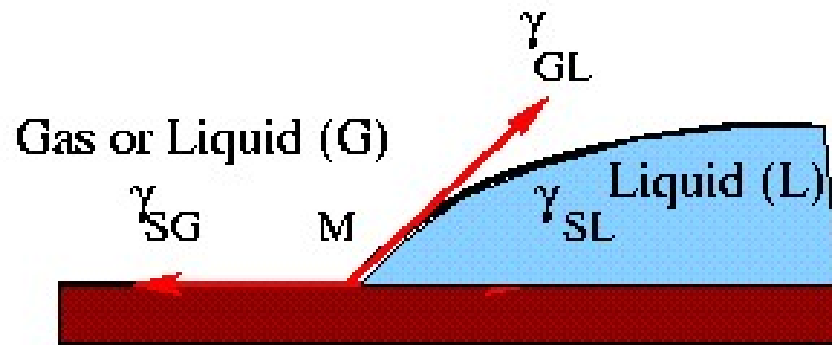
- ☞ in the saturated zone the water flow follows the head gradient and the pressure head  $\psi > 0$
- ☞ in the unsaturated zone the water pressure is below atmospheric pressure and the pressure head  $\psi < 0$

$\psi$  is called suction, matric suction, tension

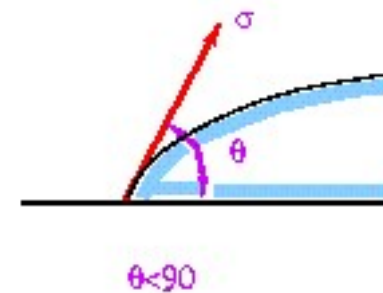


# Microscopic Consideration

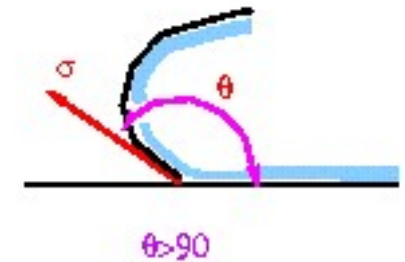
- If two or more fluid phases occupy a porous medium, one of the fluid is adsorbed on the solid surface more strongly than the other.
- The fluid that is most strongly adsorbed is called wetting fluid or wetting phase.
- The displaced phase is the non-wetting phase.
- Without interfacial forces (surface tension) separate phases would not exist.



(a)



wetting fluid



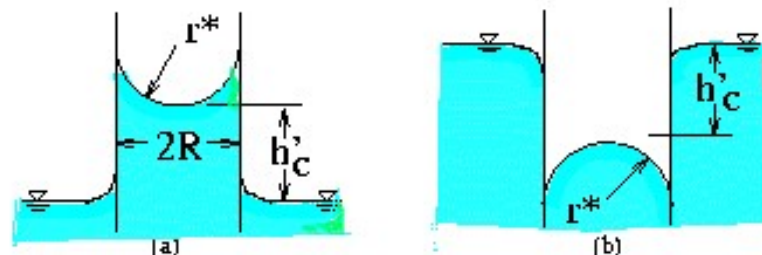
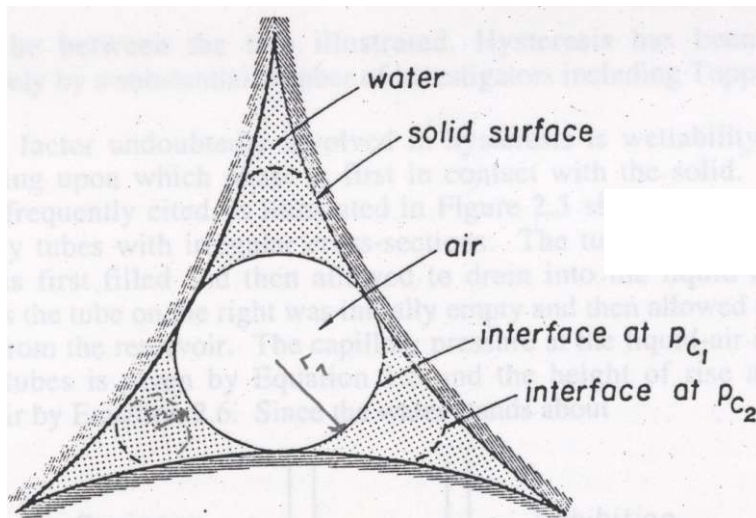
non-wetting fluid

# Capillary Pressure

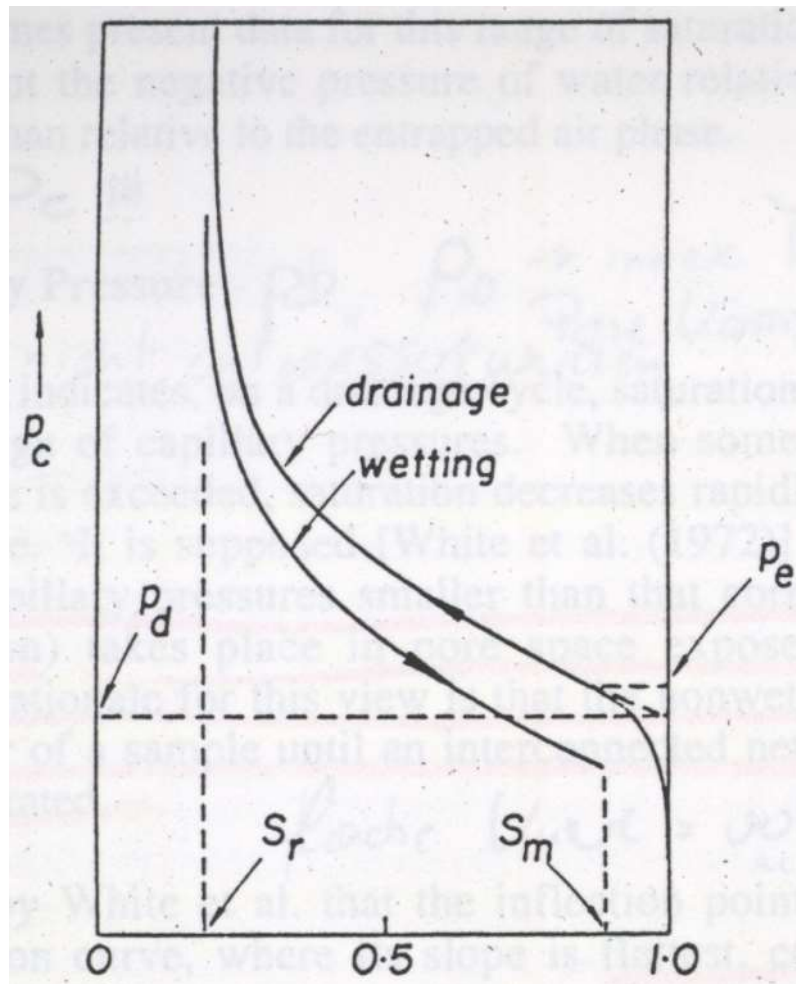
- The interfacial forces acting on the curved interface between the phases is balanced by a difference in pressure at the contact between the phases.
- The pressure difference is called capillary pressure  $p_c$ :  
$$p_c = p_{nw} - p_w$$
- The capillary pressure increases from  $p_{c1}$  to  $p_{c2}$

Laplace Equation:

$$p_c = \sigma \frac{1}{r_1} + \frac{1}{r_2} = \sigma \frac{2}{r^*}$$



# Capillary Pressure – Saturation Relationship



Increasing water saturation



Increasing air saturation



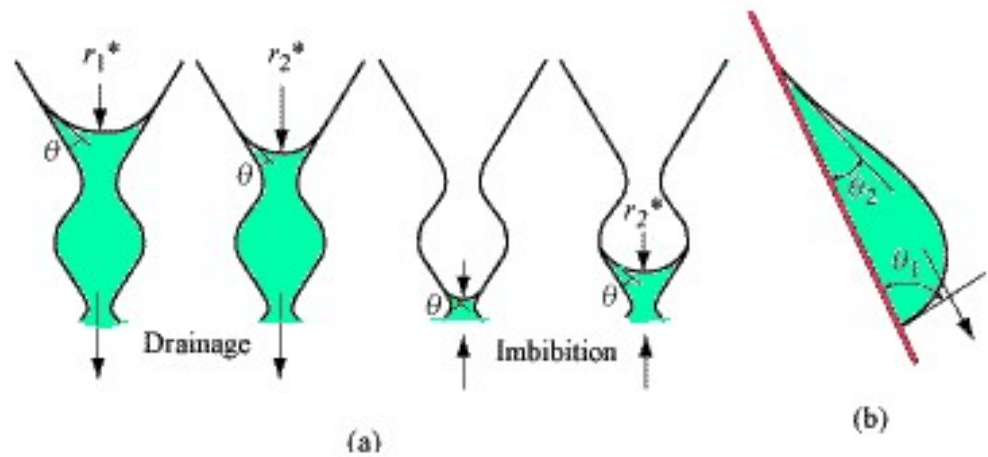
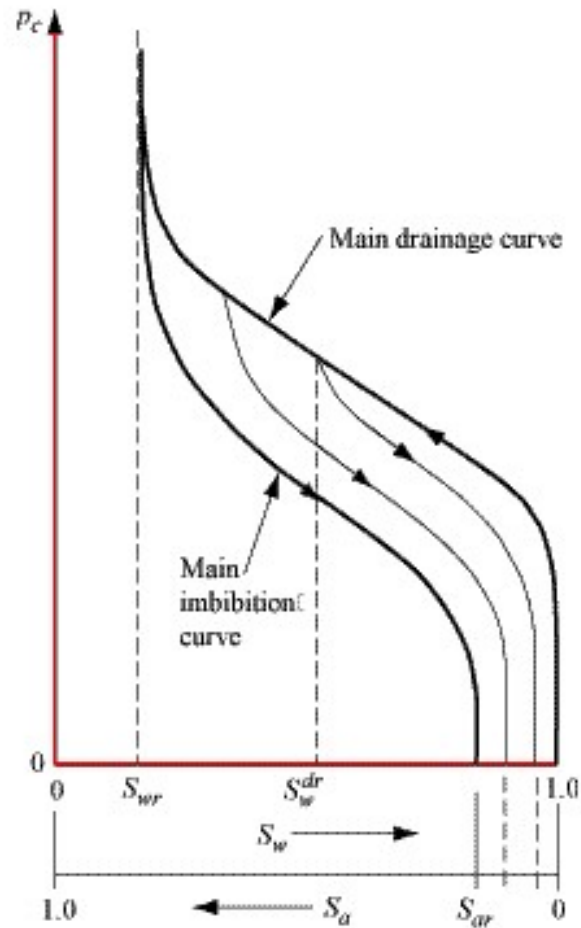
➡ Macroscopic consideration:  
 $p_c = f(S)$

- If the pore space is initially fully occupied with water a finite capillary pressure “ $p_e$ ” must be exceeded before air can enter the pore: *Entry pressure*

It is mostly at a saturation between 0.8. and 0.9

# Capillary Hysteresis

The  $p_c(S)$  relationship depends on the saturation history  
 → *Capillary Hysteresis*



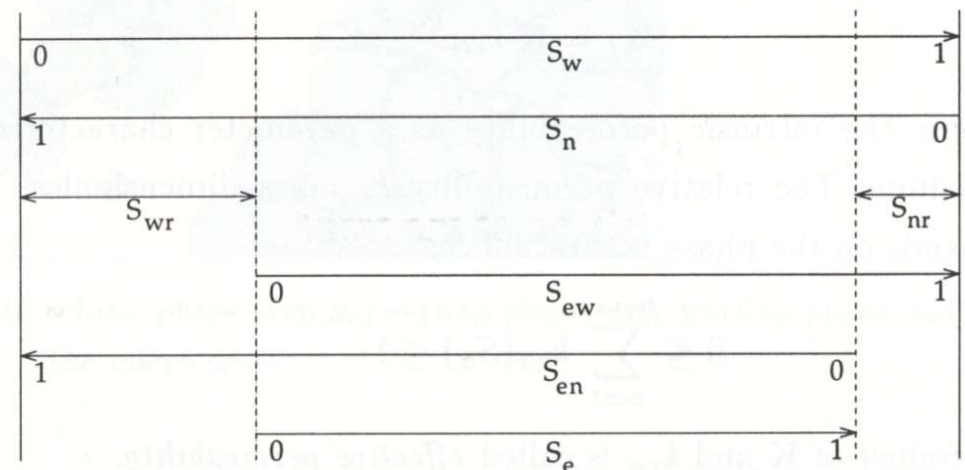
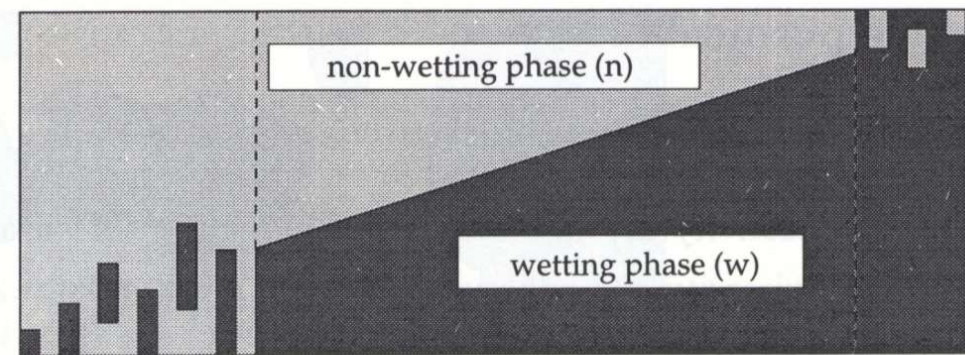
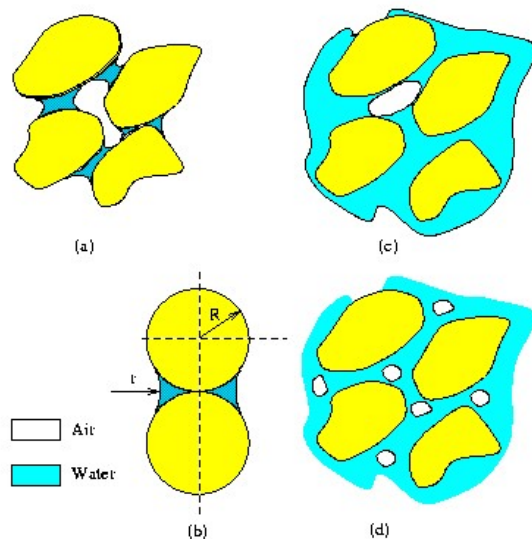


# Saturation Domains

- The *residual saturation*  $S_r$  is reached if the capillary pressure increases rapidly with negligible decrease in saturation.
- The residual saturation equals a saturation when the phase film over the solid phase are not connected anymore.

The *effective saturation*  $S_e$

is defined as: 
$$S_e = \frac{S - S_r}{1 - S_r}$$



# Empirical pc-S Relationships

*Brooks-Corey Model:*

$$p_C(S_W) = p_e S_e^{-\lambda^{-1}} \quad \text{for } p_c > p_e$$

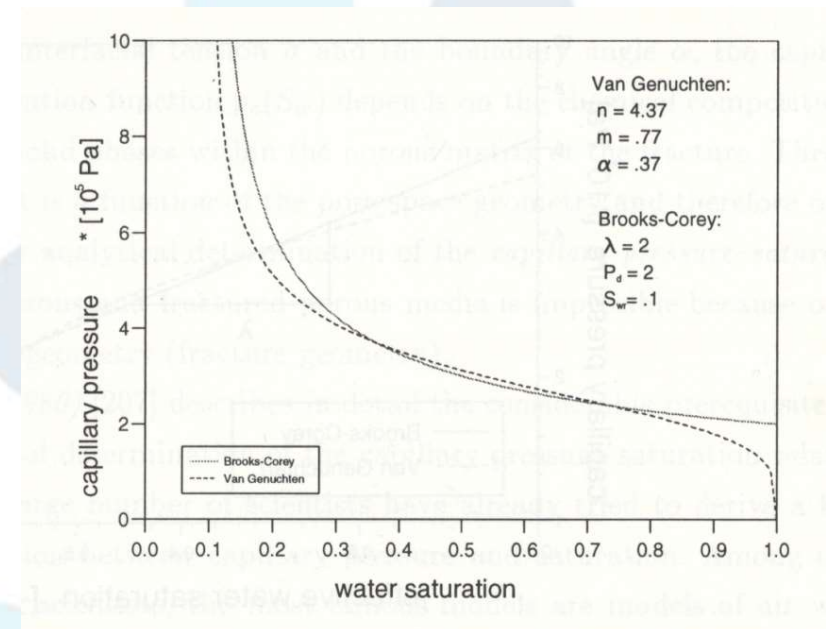
$$S_e(p_C) = \left( \frac{p_e}{p_c} \right)^\lambda$$

- $\lambda$  (BC - Parameter) and  $m, n$  (VG-Parameter) account for the pore size distribution.
- $\lambda$  can vary between 0.2 and 6.
- $\lambda$  increases with decreasing porosity
- Materials with uniform pores have large  $\lambda$  values.
- $\alpha$  accounts for the air entry pressure

*van Genuchten Model:*

$$p_C(S_W) = \frac{1}{\alpha} \left( S_e^{-1/m} - 1 \right)^{1/n}$$

$$S_e(p_C) = \left[ \frac{1}{1 + (\alpha p_c)^n} \right]^m$$



## Relative Permeability: Microscopic Consideration

In a two or more phase system the effective permeability  $K_f$  of a phase depends on the shape and pore size through the phase flows:

$$K_f = K k_{r\alpha} \frac{\rho_\alpha g}{\mu_\alpha}$$

$$0 \leq \sum_{\alpha=1}^{n_{phas}} k_{r\alpha} (S_\alpha) \leq 1$$

with:

$K_f$  = effective permeability [ $m^2$ ]

$K$  = intrinsic permeability [ $m^2$ ]

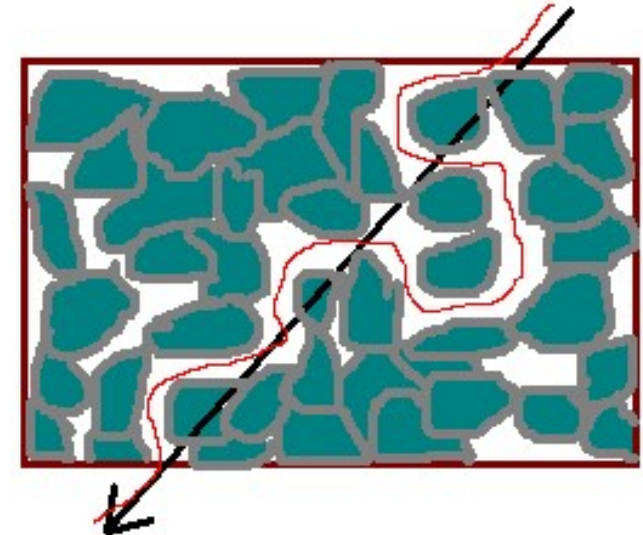
$k_{r\alpha}$  = relative permeability  
of each phase

# Relative Permeability: Microscopic Consideration

- If capillary effects exist the wetting fluid occupies the smaller pores and the non-wetting fluid the larger pores.
- Tortuosity has a strong influence on the flow, when the actual flow direction is not in the direction of the Darcy flow:

$$T_S = \left( \frac{l}{l_l} \right)^2 < 1$$

$$\left( \frac{T_1}{T_S} \right)_w \approx S_e^2 \quad \left( \frac{T_1}{T_S} \right)_{nw} \approx (1 - S_e)^2$$



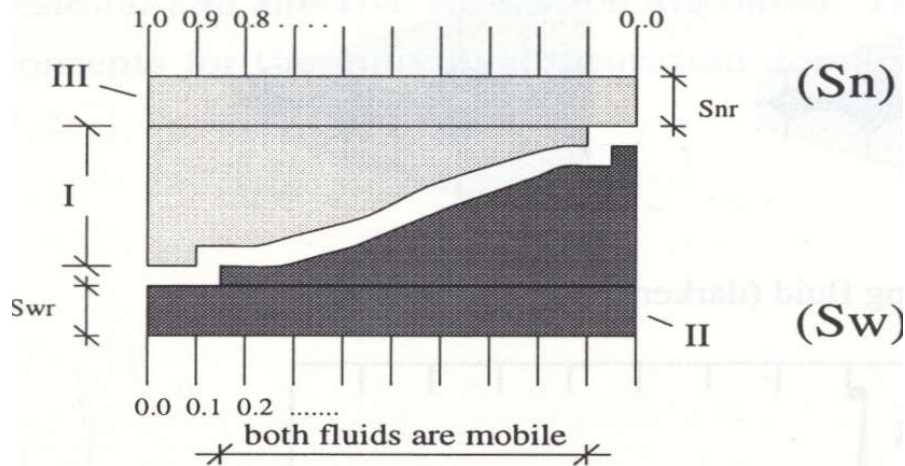
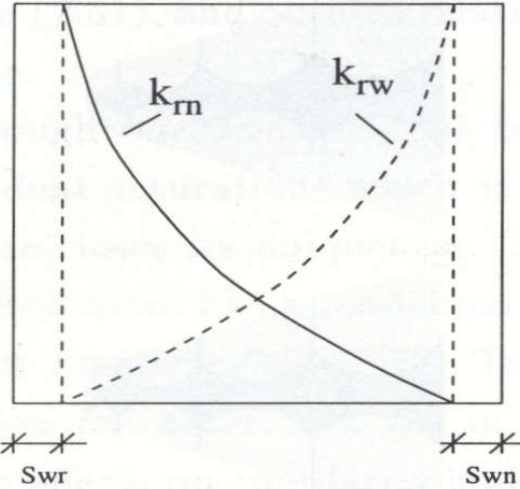
$l$  = straight line

$l_l$  = effective flow line

$T_S$  = Tortuosity under saturated conditions

$T_1$  = Tortuosity of the wetting/non-wetting fluid under unsaturated conditions

# Relative Permeability Macroscopic Consideration



The relative permeability also depends on the saturation at macro scale:

*Brooks & Corey model*

$$k_{relw} = S_e^{\frac{2+3\lambda}{\lambda}}$$

$$k_{relnw} = (1 - S_2)^2 (1 - S_e)^{(2+\lambda)/\lambda}$$

*van Genuchten model*

$$k_{relw} = S_e^\varepsilon \left[ 1 - \left( 1 - S_e^{m-1} \right)^m \right]^2$$

$$k_{relnw} = (1 - S_e)^\gamma \left[ 1 - \left( 1 - S_e^{m-1} \right)^m \right]^2$$

$\varepsilon$  and  $\gamma$  account for the connectivity of the pores