

Progress in Soil Science

Viliam Novák

Evapotranspiration in the Soil-Plant- Atmosphere System

 Springer

Evapotranspiration in the Soil-Plant-Atmosphere System

Progress in Soil Science

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Preface

Evapotranspiration is the process of water transport from evaporating surfaces to the atmosphere. Evaporating surfaces can be plant surfaces (intercepted water), substomatal cavities and cuticle tissue (transpiration), soil, the water table, or impermeable surfaces. The most important process is transpiration, the process of water movement from the soil to and through the plant, and further on to the atmosphere. This process is part of biomass production.

On average, about 60% of precipitation reaching the land surface evaporates; in dry regions this ratio is higher and can reach up to 90% of the annual rainfall.

Evapotranspiration is an invisible and complicated process; its study is difficult. Quantification of evapotranspiration involves numerous fields of science, such as hydrogeology, soil hydrology, plant physiology, and meteorology. The importance of the evapotranspiration process, particularly for biomass production, provoked its study and broad research. However, only a few books describe this process. Among them, those that strongly influenced specialists were *Evaporation in Nature* by Budagovskij (1964) for those who read Russian; and *Evaporation into the Atmosphere* by Brutsaert (1982) for those who read English. Within the framework of the series, benchmark papers were republished in hydrology (2007), and evaporation (Gash and Shuttleworth, eds), as well as basic literature about the evaporation process.

These publications analyze evapotranspiration as a process of water movement from evaporating surfaces to the atmosphere. However, water movement from the soil to the evaporating surface or roots, and water extraction by roots and water movement to a plant's leaves are mentioned only marginally.

A wide variety of methods for the calculation of evapotranspiration as a whole, as well as the components of its structure (e.g., transpiration, evaporation) have already been published.

The aim of this book is to focus attention primarily on water movement in the soil root zone and soil water extraction by roots. I also hope this volume will contribute to broadening study and research into the field of soil physics.

Finally, I would like to acknowledge the assistance of my colleagues. Completion of this interdisciplinary-oriented book required much of their effort and patience.

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List of Symbols

A	Area [L^2]
C_s	Specific volumetric heat capacity of soil [$L^2 M T^{-3} K^{-1}$]
D	Turbulent transport coefficient of height interval $(0, z)$ for neutral state of atmosphere [$L T^{-1}$]
D_1	Water vapor conductivity between leaf and its environment [$L^2 T^{-1}$]
D_m	Molecular diffusivity coefficient of water in the air [$L^2 T^{-1}$]
D_s	Turbulent conductivity of air layer between the soil surface and standard height [$L T^{-1}$]
D_o	Turbulent transport coefficient of height interval $(0, z_o)$ [$L T^{-1}$]
D_1	Turbulent transport coefficient of height interval $(0, z)$ [$L T^{-1}$]
D_2	Turbulent transport coefficient of height interval (z_o, z) [$L T^{-1}$]
E_t	Transpiration rate [$L^3 L^{-2} T^{-1}$]
E_{tp}	Potential transpiration rate [$L^3 L^{-2} T^{-1}$]
E	Evapotranspiration rate [$L^3 L^{-2} T^{-1}$]
E_{lp}	Leaf potential transpiration [$L^3 L^{-2} T^{-1}$]
E_l	Transpiration from the unit of leaf area [$L^3 L^{-2} T^{-1}$]
E_p	Potential evapotranspiration rate [$L^3 L^{-2} T^{-1}$]
E_{ep}	Potential evaporation rate [$L^3 L^{-2} T^{-1}$]
E_e	Soil evaporation rate [$L^3 L^{-2} T^{-1}$]
G	Soil heat flux [$M T^{-3}$]
H	Sensible heat flux [$M T^{-3}$]
I	Infiltration rate [$L^3 L^{-2} T^{-1}$]
K_c	Crop coefficient [-]
K_s	Saturated soil hydraulic conductivity [$L T^{-1}$]
L	Latent heat of evaporation [$L^2 T^{-1}$]
L_r	Root length of one plant [L]
L^*	Obukhov-Monin length [L]
P	Precipitation [$L^3 L^{-2}$]
P_z	Intercepted precipitation [$L^3 L^{-2}$]
M_r	Root mass per unit soil surface area [$M L^{-2}$]
R	Net radiation [$M T^{-3}$]

R_a	Extraterrestrial solar radiation [$M T^{-3}$]
R_g	Solar radiation on horizontal surface above the Earth [$M T^{-3}$]
R_s	Short-wave radiation [$M T^{-3}$]
R_{sn}	Net shortwave radiation [$M T^{-3}$]
R_l	Longwave radiation [$M T^{-3}$]
R_{ln}	Net longwave radiation [$M T^{-3}$]
Re	Reynolds number
$S(z,t)$	Root water uptake rate (sink term) [$L^3 L^{-3} T^{-1}$]
S_p	Potential root water uptake rate (sink term) [$L^3 L^{-3} T^{-1}$]
S_{dif}	Diffusion solar radiation [$M T^{-3}$]
S_{dir}	Direct solar radiation [$M T^{-3}$]
S_r	Root surface of a single plant [L^{-2}]
T	Air temperature [K]
T_0	Temperature at the bottom of dry soil layer [K]
T^*	Scaling parameter for air temperature [K]
T_1	Leaf temperature [K]
T_s	Soil surface temperature [K]
T_2	Air temperature at height z_2 [K]
V	Volume [M^3]
V_p	Water content of a canopy per unit soil surface area [$L^3 L^{-2}$]
V_r	One plant roots volume [L^3]
a_k	Convective heat conductivity of the air within the soil [$L^2 M^2 L^{-2}$]
a_s	Heat conductivity of the soil solid fraction [$L^2 M^2 L^{-2}$]
c	Specific heat capacity of soil at constant pressure [$L^2 T^{-2} K^{-1}$]
c_s	Specific heat capacity of the solid fraction of soil [$L^2 T^{-3} K^{-1}$]
c_p	Specific heat capacity of air at constant pressure [$L^2 T^{-2} K^{-1}$]
c_w	Specific heat capacity of water [$L^2 T^{-2} K^{-1}$]
c_i	Concentration (density) of an indicator, in mass unit per unit soil volume [$M L^{-3}$]
c_s	Solute concentration [$M L^{-3}$]
c_r	Solute concentration in roots [$M L^{-3}$]
d	Saturation deficit [$L^{-1} M T^{-2}$]
d_r	Root diameter [L]
d_e	Zero plane displacement height (level) [L]
e	Water vapor pressure [$L^{-1} M T^{-2}$]
e_o	Saturated water vapor pressure [$L^{-1} M T^{-2}$]
h	Hydraulic head [L]
h_a	Soil water potential (pressure head of soil water) at anaerobiosis point [L]
h_w	Soil water potential (pressure head of soil water) [L]
h_l	Leaf water potential [L]
k	Turbulent transport coefficient [$L^2 T^{-1}$], soil hydraulic conductivity [$L T^{-1}$]
l	Length of an element under study in the direction of wind speed [L]
l_r	Specific root length [$L L^{-3}$]
l_w	Effective leaf diameter [L]

m	Soil free porosity [-]
m_r	Root mass of a single plant [M]
n_r	Relative root characteristics (mass,length,surface) density [L^{-1}]
n_{rd}	Relative root mass density [L^{-1}]
n_{rc}	Relative cumulative water uptake rate by roots [-]
n_{ru}	Relative water uptake rate by roots [L^{-1}]
q	Specific humidity [-]
q^*	Scaling parameter for specific humidity
q'	Instantaneous fluctuations of specific humidity [-]
q_1	Saturated specific air humidity in inter- cells leaf space [-]
q_{so}	Saturated specific humidity at the temperature T_s [-]
q_s	Specific humidity at evaporating surface and temperature T_s [-]
q_o	Saturated specific humidity [-]
q_{2o}	Saturated specific humidity at the temperature T_2 [-]
q_w	Soil water flux [$L T^{-1}$]
r	Relative humidity [-]
r_a	Aerodynamic resistance [$T L^{-1}$]
r_c	Canopy resistance [$T L^{-1}$]
r_l	Leaf resistance [$T L^{-1}$]
r_s	Stomata resistance [$T L^{-1}$]
r_{sr}	Resistance to water flow at soil-root interface [$T L^{-1}$]
r_{sl}	Resistance to water flow between soil and leaf [$T L^{-1}$]
s	Thermal emissivity [-]
s_r	Specific root surface [$L L^{-3}$]
t	Time [T]
u	Wind velocity [$L T^{-1}$]
u^*	Friction velocity [$L T^{-1}$]
v_{is}	Indicator flux rate in soil [$ML^{-2} T^{-1}$]
v_{ir}	Indicator flux rate in roots [$ML^{-2} T^{-1}$]
v_l	Water content per the unit leaf area [$M L^{-2}$]
v_p	Rate of water flow from soil to the plant leaves [$L^3 L^{-2} T^{-1}$]
v_r	Water flux rate in roots [$L T^{-1}$]
v	Water flux rate in the soil [$L T^{-1}$]
w	Soil water content in mass units [$M M^{-1}$]
x,y,z	Coordinates [L]
z_p	Plant height [L]
z_r	Root system depth [L]
z_s	Thickness of dry soil layer [L]
z_0	Roughness length [L]
z_2	Standard height of measurements (usually at 2 m above the soil surface) [L]
α	Albedo [-]
γ	Psychrometric coefficient [$L^{-1} M T^{-2} K^{-1}$]
θ	Volumetric soil water content [$L^3 L^{-3}$]

θ_a	Soil water content corresponding to the anaerobiosis point [$L^3 L^{-3}$]
θ_s	Soil water content at the soil surface [$L^3 L^{-3}$]
θ_0	Soil water content at the bottom of dry soil layer [$L^3 L^{-3}$]
θ_{cr}	Critical soil water content [$L^3 L^{-3}$]
θ_r	Residual soil water content [$L^3 L^{-3}$]
λ	Heat conductivity [$L M T^{-3} K^{-1}$]
ν_a	Kinematic viscosity of air [$L^2 T^{-1}$]
ρ_a	Air density [$M L^{-3}$]
ρ_r	Dry root mass density [$M L^{-3}$]
ρ_s	Dry soil density [$M L^{-3}$]
σ	Stefan-Boltzmann constant [$M T^{-3} K^{-4}$]
χ	von Kármán constant [-]
η	Dynamic viscosity [$L^{-1} M T^{-1}$]
$\omega(z)$	Leaf area index as a function of z [-]
ω_o	Leaf area index [-]
τ	Quantity of motion [$ML^{-1} T^{-2}$]
<i>SPAS</i>	Soil-Plant-Atmosphere System
<i>BLA</i>	Boundary Layer of Atmosphere

Chapter 1

Evapotranspiration: A Component of the Water Cycle

Abstract Evapotranspiration as a process is part of the water cycle of the Earth; it is the most important consumer of energy, creating the link between water and energy cycles of the Earth. The physics of water phase change is briefly presented. Consumption of energy to change liquid water into water vapor cools the biosphere, thus allowing the creation of suitable conditions for life on the Earth. This chapter contains basic information about the Earth and continents' water cycle and its components, as well as the energy balance structure of the Earth. The kinetic theory of fluids is used to quantify the evaporation process because it depends on the properties of an environment, allowing us to find the most important properties of the environment influencing evapotranspiration. The kinetic theory of evaporation can help us understand evaporation as a process, but does not allow use in directly quantifying it; therefore other methods should be used.

1.1 The Evaporation Process and Its Basic Properties

Evaporation is a process in which matter changes its phase from the solid or liquid phase to the gaseous phase. The change of solid to gaseous phase is usually denoted as sublimation.

All matter can evaporate if its molecules have enough energy for phase transition. In this book, our interest will be focused on evaporation of water in nature and from artificial structures in the environment.

Evaporation of water from plants, which is transported through the plants from soil to the leaves, is of particular importance and is referred to as transpiration. It is a part of the plant production process. Evaporation of water accumulated directly on the plant surface can be defined as evaporation of intercepted water. Simultaneous evaporation from soil, water, and plants is known as evapotranspiration.

The different terms allow us to specify evaporation with respect to the evaporating surface. Evaporation of water is a term denoting phase transition from liquid to gaseous phase—laws describing it are equal for different evaporating substances.

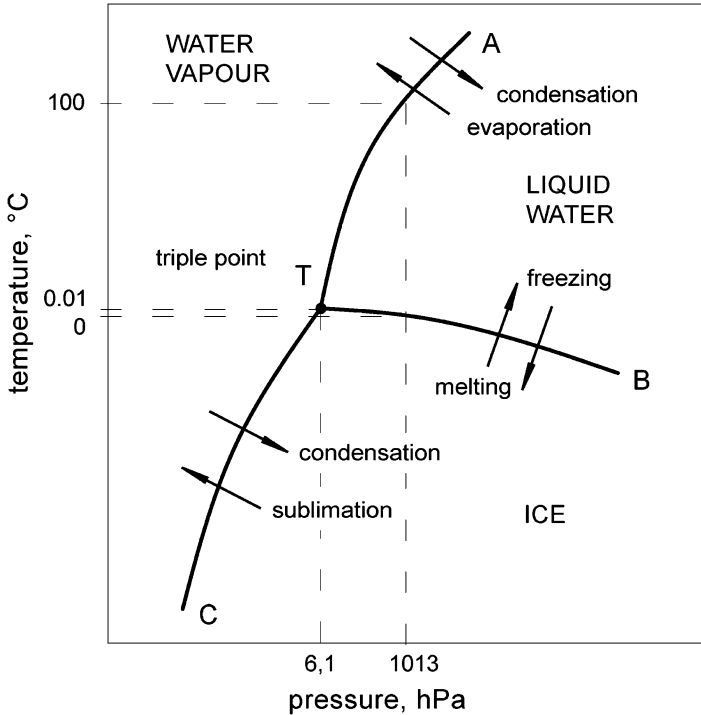


Fig. 1.1 Pressure versus temperature behavior of water near the triple point

It has been shown that the first stage of the process known as sublimation is phase transition from solid to liquid phase and then to gaseous phase. This means that “direct” phase transition from solid to liquid phase, in reality, does not exist. The term “evaporation” describes the process of evaporation as well as the quantity of evaporated mass. The evaporation process is characterized by high specific consumptive use of energy. Monteith compares the evaporation process to a transaction in which a wet surface is “selling” water to its environment in a gaseous phase. Any gram of water at 20°C is paid by 2,450 J of energy. The “transaction” can be performed in different ways: by the energy of solar radiation, or by hot air convection from places of higher temperature. The “price” of evaporated water is high. To increase the temperature of 1 g of water to 1°C, 4.18 J energy (at 20°C) is needed. To increase the temperature of 1 g of water from 0°C to 100°C, 418 J of energy is needed, which is 5.86 times less than is needed to evaporate 1 g of water! This is the reason why evaporation is important not only for the water cycle on the Earth, but is vital for the cycle of energy too.

Evaporation is a process in which energy is consumed; therefore, it is a thermodynamic process. For thermodynamic processes, the macroscopic state of matter typically can be characterized by two thermodynamic characteristics: temperature and pressure. This is demonstrated in the pressure–temperature diagram (Fig. 1.1) for pressures and temperatures around the so called “triple point,” where all three

phases of water are in equilibrium. The liquid phase of water exists between curves A-B, water vapor between curves A-C, and ice between curves B-C. Phase transition means crossing those curves; it can be done in both directions. The intersection of all three curves is the triple point, denoted as T , characterized by coordinates (0.01°C; 6.1 hPa). The Earth's surface has an average atmospheric pressure of 10,013 hPa, at which water boils at 100°C and water freezes at 0°C. Phase transition temperatures depend on atmospheric pressure; pressure fluctuation is small at a particular site, therefore it can be neglected in majority of cases.

Evaporation is the process linking water and energy cycles of the Earth.

1.2 Evaporation and the Kinetic Theory of Fluids

The theory of the evaporation process is based on the results of the kinetic theory of evaporation, originally published by Shulejkin (1926). The kinetic theory of fluids is based on the fact that all the molecules are in continuous, chaotic motion. Velocities of molecules' motion are distributed stochastically in a wide range, which can be described by Maxwell function. Mean quadratic velocity of an oxygen molecule (the most probable velocity) at 0°C is 461 m s^{-1} (supersonic velocity), number of collisions per second is 4.29×10^9 , and mean free path of a molecule (the distance between individual collisions) is $8.7 \times 10^{-6} \text{ cm}$.

Water molecules of the water table monomolecular layer (the water molecule dimension is approx. $2.72 \times 10^{-8} \text{ m}$) are attracted in one direction only (to the liquid) and they possess relatively low free energy, but molecules below this layer are of relatively high free energy and can overcome the energetic barrier and enter the adjacent atmosphere. To overcome this energetic barrier, one molecule needs energy that equals latent heat of evaporation of one molecule (W). During evaporation, water loses molecules with the highest kinetic energy. Therefore, the average kinetic energy of water molecules decreases and water temperature decreases too. This phenomenon leads to a decrease of the evaporation rate, because the maximum of the molecules energy distribution is shifting to the lower values.

To preserve evaporation rate, the flux of energy to the evaporating surface should be maintained; it equals the latent heat of evaporation. Energy flux to the evaporating surface increases kinetic energy of water molecules and temperature of water as well. The mean quadratic velocity of water molecules and the evaporation rate increases too.

Water molecules move randomly in the adjacent layer of the atmosphere and part of them come back to the liquid water. The ratio of the number of molecules evaporating and condensing depends on the number of molecules in the air layer adjacent to the evaporating surface. An air layer saturated with water vapor above the wet surface contains the maximum water vapor molecules at some temperature and effective evaporation does not exist; however, equilibrium exchange of water molecules between liquid water and air layer still exists.

Evaporation is a complicated phenomenon and its quantification by kinetic theory methods is difficult. The main reason is that basic postulates of kinetic theory are

valid only approximately in liquids (high density of molecules, finite dimensions of them). But, application of kinetic theory to evaporation can help us understand this process in relation to the conditions in which this process is occurring. The next part can be applied not only for water, but for the sake of simplicity liquid water will be used as a model.

The number of water molecules N_c condensed on the unit area of water surface is proportional to the density of water molecules in the air n and to the average molecules velocity v :

$$N_c = n \cdot v \quad (1.1)$$

Water molecules will evaporate if they gain energy W needed to leave liquid. Probability n_1 to gain the energy higher than W is (Feynman and Leighton 1982):

$$n_1 = \exp\left(-\frac{W}{kT}\right) \quad (1.2)$$

where T is liquid temperature, K; and k is Boltzmann constant, $k = 1.38 \times 10^{-23} \text{ J K}^{-1}$.

The number of water molecules N_e leaving the unit area of liquid per unit of time is proportional to the density of molecules per unit water surface area ($1/A$), to the time interval needed to pass the surface layer of molecules when escaping the liquid (d/v), and to the probability of water molecule escape n_1 :

$$N_e = \frac{1}{A} \frac{v}{d} \exp\left(-\frac{W}{kT}\right) \quad (1.3)$$

where v is velocity of water molecule; d is diameter of water molecule; and A is molecule surface area.

Volume of a spherical water molecule V_a can be approximately expressed as a product of molecule diameter d and maximum molecule cross-section area A_1 :

$$V_a = d \cdot A_1 \quad (1.4)$$

Then, Eq. 1.3 can be rewritten as:

$$N_e = \frac{v}{V_a} \exp\left(-\frac{W}{kT}\right) \quad (1.5)$$

During the state of equilibrium, the number of molecules condensing N_c and evaporating N_e are the same ($N_c = N_e$), and by combination of equations we get:

$$nv = \frac{v}{V_a} \exp\left(-\frac{W}{kT}\right) \quad (1.6)$$

because W is the energy needed to overcome bonds between molecules in the liquid, and the molecules' energy distribution (after evaporation) in the air is the same as it was in the liquid. Equation 1.6 expresses number of molecules leaving unit area of evaporating surface per unit of time. It is the maximum evaporation rate in a case in which all the molecules from the air layer adjacent to the evaporating surface are removed. In reality part of the molecules return to the water, and therefore actual evaporation rate is lower than expressed by Eq. 1.6. Evaporation velocity decrease can be expressed by the coefficient k_e , defined as the ratio of molecules number condensing at the liquid surface N_c and leaving liquid N_e :

$$k_e = \frac{N_c}{N_e} \quad (1.7)$$

Then, evaporation velocity can be expressed as:

$$N_e = n \cdot v \cdot k_e = \frac{v \cdot k_e}{V_a} \exp\left(-\frac{W}{kT}\right) \quad (1.8)$$

From Eq. 1.8 it follows that evaporation rate:

- is proportional to the molecules' motion velocity and is indirectly proportional to their volume
- is indirectly proportional to the energy needed to overcome energetic barrier in the liquid; it is property of liquid and is different for different liquids
- is proportional to the liquid temperature

From the kinetic theory of evaporation it follows:

- Water can evaporate if water vapor pressure in the air layer adjacent to the liquid surface is below saturated water vapor pressure corresponding to the air temperature.
- To keep evaporation rate constant, it is necessary to preserve energy flux to the area of evaporation in a rate needed for phase transition of liquid to vapor. Evaporation of water needs specific energy known as latent heat of evaporation (L), depending on liquid temperature. For $T = 20^\circ\text{C}$, $L = 2.45 \times 10^6 \text{ J kg}^{-1}$.

Results of kinetic theory are in agreement with our experiences. But direct use of the aforementioned equations to calculate evaporation in a field is difficult. Distribution of water molecules' velocity as well as temperature distribution near the evaporating surface should be known, as well as the value of coefficient k_e . Therefore, to estimate evaporation rate under different conditions, so called macroscopic methods are used, based on measurement of "macroworld" properties, which are phenomena such as air temperature or wind velocity integrating the effect of a large number of molecules.

1.3 Water Balance and Water Cycle

The term water balance can express algebraic sum of water fluxes to and out of the defined volume during given time interval. This term is used to quantify the ratio of individual components of water balance too. Water balance can denote also the process of estimating individual terms of the water balance equation. Water balance is application of the energy conservation statement to part of the hydrological cycle (Hillel 1982).

The water cycle starts by precipitation (irrigation) falling to the Earth surface. Precipitation can infiltrate to the soil at the rate of precipitation; if the infiltration rate is smaller, ponding on the soil surface occurs. Depending on soil morphological properties, part of the water can flow out (surface run off), part of the ponded water can infiltrate later, and part of the water volume can evaporate. A particular phenomenon of plant water retention is intercepted water, that is, plant surface retention of water, which will evaporate. Part of the infiltrated water can evaporate from the soil surface layer. The rest of the infiltrated water is retained by the soil and later extracted by plant roots to transpire. In the case of a shallow water table, part of the infiltrated water can reach and recharge it and then feed water streams.

The other source of water for soil can be surface water from the other parts of the territory or groundwater feeding. Snow precipitation accumulates on the soil surface and its melting and subsequent infiltration or runoff depend on the temperature regimen of the soil-plant-atmosphere system (SPAS).

Evapotranspiration is one of the most important water balance equation components not only for its quantity, but for its importance in biomass production process.

The basic water balance equation can be written for a catchment area and a period of 1 hydrological year. It expresses the distribution of annual precipitation total P to outflow O and evapotranspiration E :

$$P = E + O \quad (1.9)$$

Water balance of a territory for a short time interval can be expressed by the equation, in which rates [$\text{kg m}^{-2} \text{s}^{-1}$], (instead of totals) are used:

$$\frac{dS}{dt} = (P + I + O_i) - (E + O) \quad (1.10)$$

where S is water quantity in a catchment per unit area, P is precipitation rate, I is irrigation rate, O_i is rate of water flow into the catchment, E is evapotranspiration rate, O is outflow rate, and t is time.

Soil root zone water balance can be expressed (taking into account vertical flow components only):

$$\frac{dS}{dt} = (I_i + I_u) - (E_e + E_t - O_d) \quad (1.11)$$

Table 1.1 Average annual values of the Earth's water balance equation components Denmead (1973)

Continent	Area, 10 ³ km ²	Volume of water per year, km ³ year ⁻¹		
		Precipitation (<i>P</i>)	Runoff (<i>O</i>)	Evapotranspiration (<i>E</i>)
Europe	10,500	8,290	3,210	5,080
Asia	43,475	32,240	14,410	17,830
Africa	30,120	22,350	4,570	17,780
North America	24,200	18,300	8,200	10,100
South America	17,800	28,400	11,760	16,640
Australia	7,615	34,170	300	3,170

Continent	Water layer, mm/year			Ratio	
	<i>P</i>	<i>O</i>	<i>E</i>	<i>E/P</i>	<i>O/P</i>
Europe	789	305	489	0.62	0.38
Asia	742	332	410	0.55	0.45
Africa	742	151	591	0.8	0.2
North America	755	339	417	0.55	0.45
South America	1,600	650	940	0.59	0.41
Australia	455	40	415	0.91	0.09

where I_i is infiltration rate into soil, I_u is groundwater inflow to the soil, E_e is soil evaporation rate, E_t is transpiration rate, and O_d is outflow to the groundwater.

The most important component in the outflow part of Eq. 1.11 is usually transpiration and evaporation. In areas with groundwater depth 2 m below soil surface, the groundwater feeding term should be neglected. In areas with high precipitation totals (Northern Europe, Canada) and low air temperatures, surface and subsurface outflow are dominant components of the Eqs. 1.11. In Table 1.1 the average annual components of the water balance equation of continents can be seen. It can be seen that Africa and Australia are evapotranspiring at a greater part of precipitation than other continents. Differences in component structure of the water balance equation are observed in different areas of continents. In Europe, e.g., the evaporation ratio (ratio of evapotranspiration to precipitation) in Hungary is 0.91, but the ratio of contiguous state Slovakia is 0.647. Those differences are mainly caused by the morphological properties of both countries; Slovakia is mainly hilly, but lowlands are characteristic for Hungary.

The interface between precipitation and other components of the water balance equation includes soil-surface and soil-roots surface as well. Water transport processes in the soil-plant-atmosphere system (SPAS) are shown in Fig. 1.2. Typical seasonal courses of soil water balance equation components are shown in Fig. 1.3, for a site in Trnava (South Slovakia) with maize canopy. Ground water table was about 10 m below the soil surface; surface runoff was not observed during the vegetation period. Water content of the maize canopy and its changes were neglected too.

Our knowledge of evapotranspiration process is not satisfactory at this time. To calculate evapotranspiration flux, one needs complicated devices to estimate the

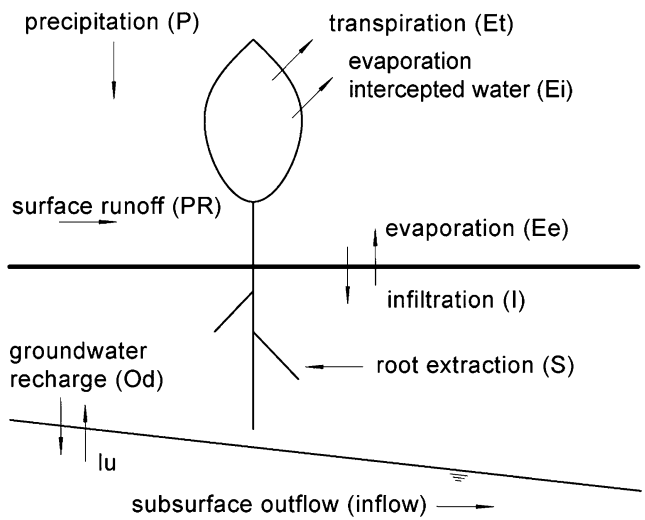


Fig. 1.2 Water transport processes in the soil-plant-atmosphere system (SPAS)

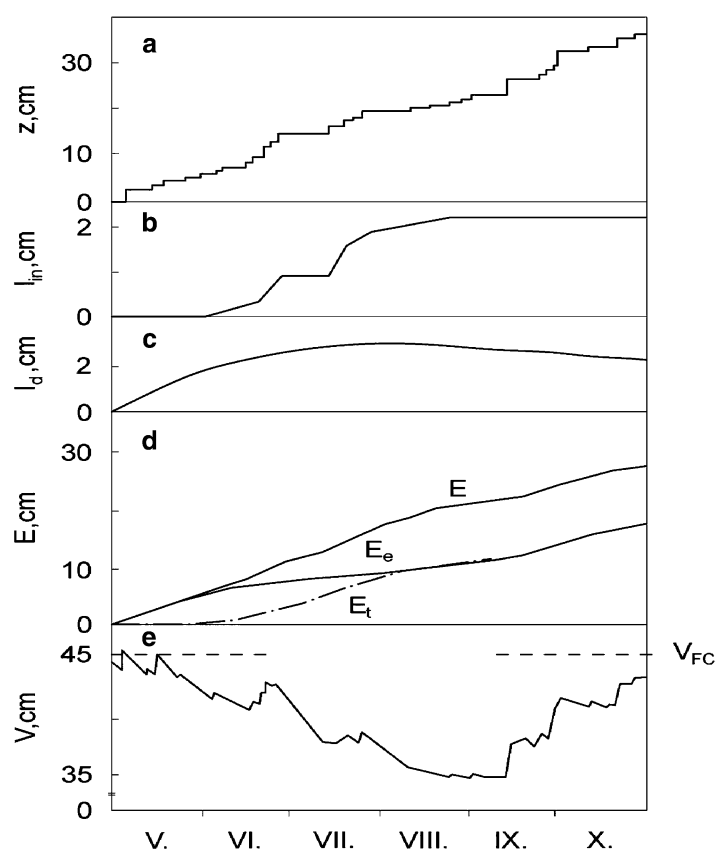


Fig. 1.3 Seasonal courses of soil water balance equation components during the vegetation period of maize. (a) precipitation total, (b) precipitation interception by canopy, (c) integral water flux through bottom boundary at a depth 150 cm below soil surface, (d) evapotranspiration (E), evaporation (E_e), transpiration (E_t), (e) water content in soil layer 0–150 cm (Trnava site, South Slovakia, 1981)

input parameters of the SPAS. To include plant properties in the calculations, fully understanding methods of evapotranspiration and its components is still the actual problem.

1.4 Energy Balance of the Evaporation Area

Energy and water transport processes are interconnected by evaporation. Therefore it is important to estimate energy that can be used for evaporation. As will be presented later, the energy transport to the evaporating surface is the deciding factor for evaporation rate from water surfaces, wet soils, and canopies grown on wet soils.

From a methodological point of view, it is suitable to perform energy water balance at the evaporating surface level, or at a height above it, where vertical fluxes of water and energy do not change significantly (Budagovskij 1981). Practical reasons dictate measurement of such fluxes at the height of 2 m above the soil surface; standard meteorological measurements, frequently used to calculate evaporation fluxes, are performed at this height.

The energy balance equation of this volume (between soil surface and 2 m height above it) can be written:

$$R = LE + H + G + A_f + A_r \quad (1.12)$$

where R is net radiation at reference level, sum of all radiation fluxes, W m^{-2} ; E is water vapor flux (evapotranspiration), $\text{kg m}^{-2} \text{s}^{-1}$; H is convective (turbulent, sensible) flux of heat from evaporating surface to the atmosphere, W m^{-2} ; L is latent heat of evaporation, $\text{J kg}^{-1} \text{s}^{-1}$; G is soil heat flux, W m^{-2} ; A_f is photosynthetic energy flux, W m^{-2} ; and A_r is change of canopy heat capacity, W m^{-2} .

Equation 1.12 does not involve the advective flux of energy and snow melting. Term A_f is very small (usually less than 2% of net radiation) (Budagovskij 1981) and term A_r is even smaller, therefore Eq. 1.12 usually is used in simplified form without terms A_f and A_r .

In the literature, conventionally downward net radiation flux is supposed to be positive; turbulent (sensible) heat flux and water vapor flux are usually positive in an upward direction. Conversely, downward soil heat flux is taken as positive (Budagovskij 1964; Brutsaert 1982).

Energy fluxes and their directions depend on the SPAS properties, and they possess typical daily and seasonal courses. Figure 1.4 presents typical daily courses of energy balance equation components during a clear day. Courses of some meteorological components corresponding to those courses are shown in Fig. 1.5.

The most important source of energy for evaporation is solar radiation. Intensity of solar radiation through the unit area perpendicular to the direction of radiation is approximately constant, and therefore it is referred to as solar constant $s = 1.4 \text{ kW m}^{-2}$. The majority of radiation flux occurs in the wavelength range 0.3–3 μm , in which the visible part of the solar radiation is in the range of

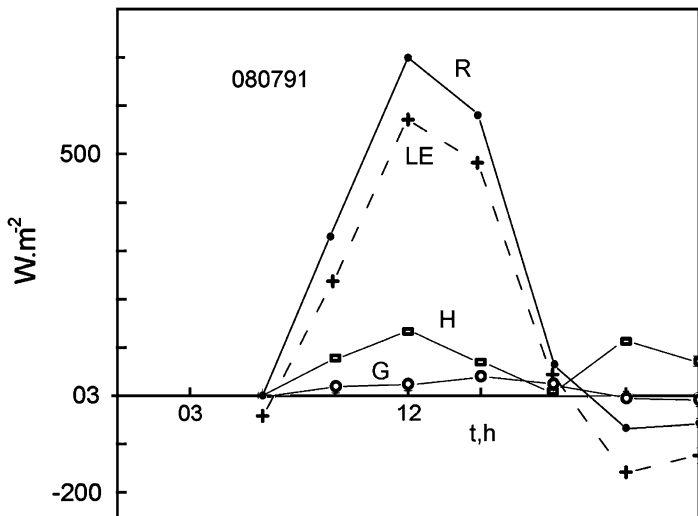


Fig. 1.4 Daily courses of the energy balance equation components over maize canopy during a clear summer day (Kursk, Russia, 1991)

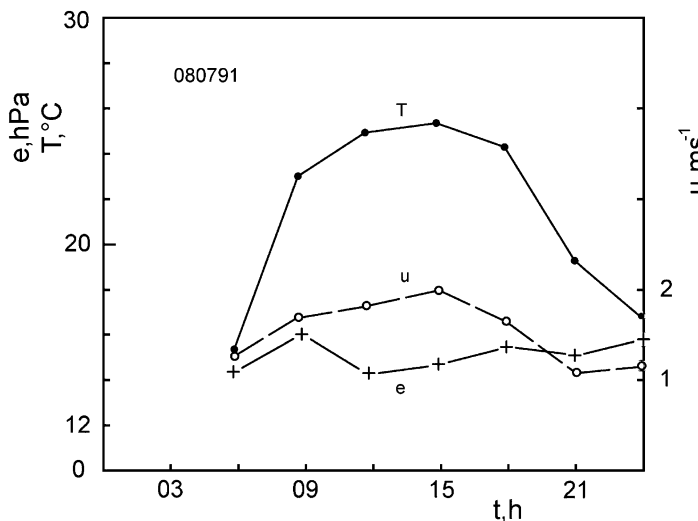


Fig. 1.5 Daily courses of air temperature T , wind velocity u , and water vapor pressure e at standard height above the maize canopy during a clear summer day (Kursk, Russia, 1991)

wavelength 0.4–0.7 μm (Brutsaert 1982). Solar radiation corresponds to the emission spectrum of a black body at the temperature approx. 6,000 $^{\circ}C$.

The Earth’s surface, with an effective temperature of approximately 300 K, radiates with relatively low intensity and with longer wavelengths than the Sun,