Chapter 1

WATER OCCURRENCE AND ITS FUNCTION IN NATURAL SYSTEMS

1.1 SYSTEMS OF THE NATURAL ENVIRONMENT

Water exists as scattered humidity and as spatially limited water formations below, on and above the Earth's surface. Water resources are water formations which can be utilized by human society. Water and water formations are dynamic; they are always in motion and their state of aggregation is forever changing. These processes continue without interruption, change in space and time and transform the natural environment.

The natural environment is formed by a number of systems, or complexes of mutually interrelated elements, whose relationships within the framework of these complexes are more important than their relations with the elements of other systems. In the important part of the natural environment which constitutes the object of the present investigation it is possible to distinguish:

(a) abiotic systems, created by water, soil and air elements and characterized by:

- morphological (topographical) data
- pedological and geological data (soil is a mixed abiotic-biological element)
- hydrogeological and hydrometeorological data

(b) biotic-biological systems (ecosystems), originating with the development of living matter in a defined part of the abiotic environment and

(c) socio-economic systems, i.e. administrative, economic and technical systems (Fig. 1.1) originating with the formation of human society and possessing important interconnections with the above two systems.

The natural environment of Earth represents a semi-closed system. The input of matter into this system from outer space is negligible. The movement of matter inside this system is enabled by an input of energy, consisting mainly of solar energy and the internal energy of Earth itself. This system, due to its own homeostatic mechanisms and detectors, tends to achieve a state of equilibrium balancing accidental deflections from this state.

The material couplings which form interrelations among these systems include:

- biotic-abiotic couplings, e.g. the quantity of dissolved oxygen caused by the decay of a biomass
- abiotic-biotic couplings, e.g. the dependence of the intensity of biological processes on water temperature
- socio-abiotic couplings, e.g. as manifested especially by the impact of urbanization, industrialization and agricultural production on runoff and sediment transport

- socio-abiotic-biotic couplings, e.g. as manifested by the role of urbanization, industrialization and intensive agricultural production in polluting certain ecosystems.

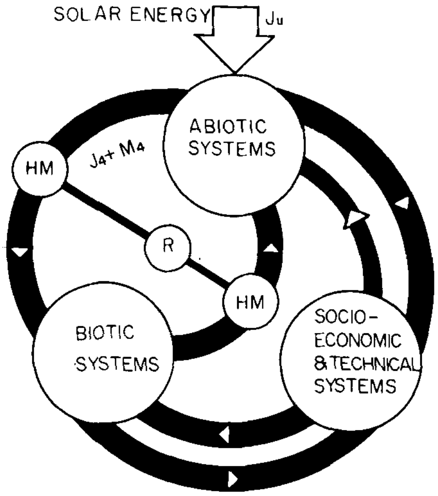


Fig. 1.1. The penetration of matter and energy through the abiotic and biotic (also socio-economic) systems. The equilibrium of relevant systems and its recovery depend on the energy and matter input: R - detector, HM - homeostatic mechanisms.

These material couplings also form complex interrelations, such as the socio-abiotic-biotic-social interrelation manifested by the influence of industrialization and the subsequent water pollution and eutrophication on water utilization.

The task of analyzing these interrelationships among the various systems is complicated not only by the complexity of the couplings and interrelations concerned, but also by the lack of data available (Fig. 1.2).

As only selected couplings are operationally controllable, only a few can be checked systematically. Moreover, because data monitoring is neither complex nor fully systematic, the relevant series of data in the different categories do not mutually correspond and are therefore inadequate. Furthermore, frequently undesirable secondary couplings occur and have a negative influence on the function of the system in question, sometimes bringing about a gradual change in the system's behaviour.

The movement of water and other matter within and between these systems changes in time and space: The importance of individual relations is variable. Regarded in this way the doctrine of water management concerns the structure and the function of systems, thus enabling the water to fulfill its natural functions and to be utilized for the various present and future requirements of society.

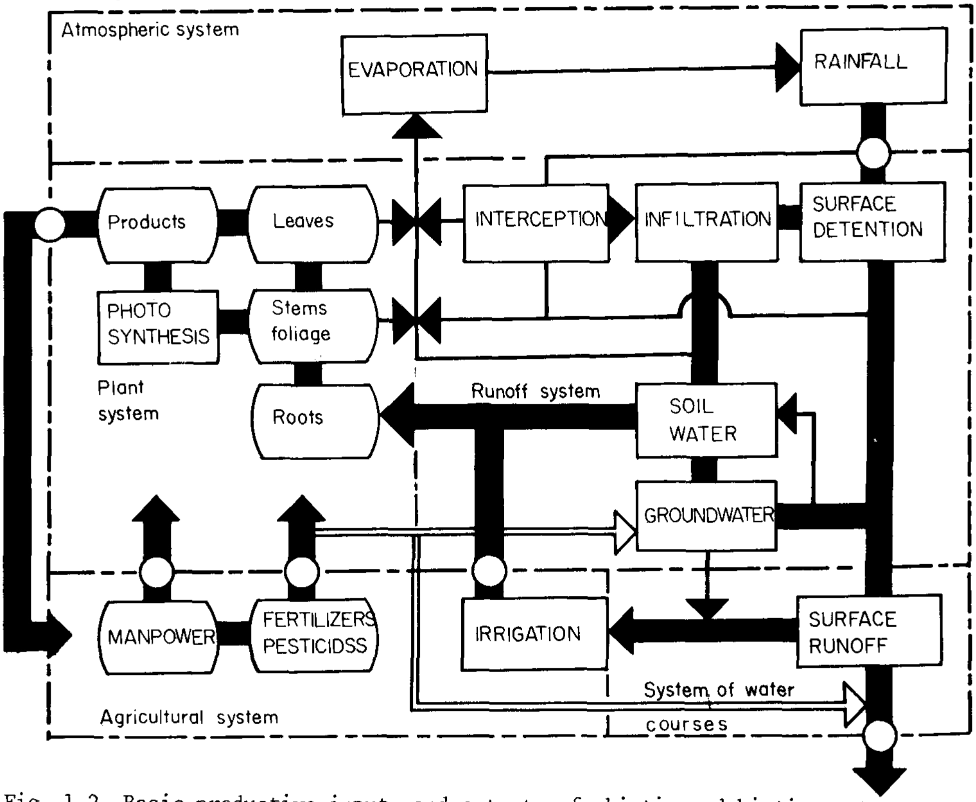


Fig. 1.2. Basic productive inputs and outputs of abiotic and biotic systems. Monitored, i.e. systematically checked inputs are marked by a circle; accidental, undesired outputs are marked by white arrows.

1.2 ENERGY INPUT AS A CAUSE OF THE HYDROLOGIC CYCLE

The Earth reflects a part of the external energy input which it receives, distributing the rest between the air, water, soil and geological formations and radiating part of it back into the Universe. The basic equation of the energy balance expresses the law of the conservation of energy (Fig. 1.3)

$$J_u \cdot (1 - \varrho) - J_g = \sum_{k=1}^{4} J_k \qquad (m^2.kg.s^{-2}) \qquad (1.1)$$

J_u – sun and other radiation from the Universe (short wave)

ϱ – albedo (coefficient of reflection)

J_g – global radiation (long wave)

J_1 – energy supply to the atmosphere

J_2 – energy supply to the hydrosphere

J_3 – energy supply to the pedosphere and lithosphere

J_4 - energy supply to the biosphere

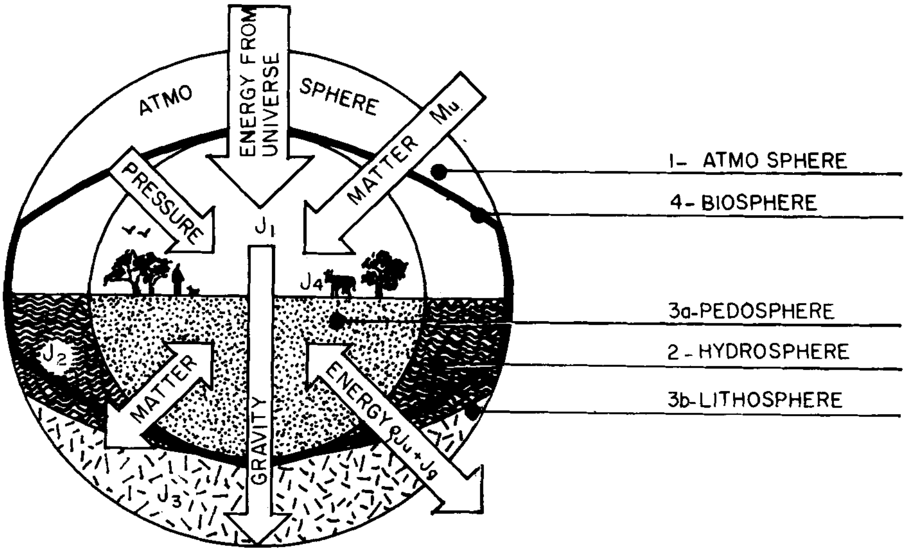


Fig. 1.3. Basic interrelations of the systems of atmosphere, lithosphere and pedosphere as well as the hydrosphere: movement of matter in the gravitational field, enabled by the supply of energy, forming the main input. The input of matter from the universe is negligible.

The Earth reflects on average 34% of the energy input. The coefficient of reflection, the albedo, depends essentially on the character and morphology of the surface, the state and quality of the atmosphere above, as well as on the angle of incidence of the rays. Stretches of water reflect 10% of the energy on average, lawns 15%, forests 20%, deserts 30% and snow 80%. The type of energy utilization changes with the character of the surface: 90% of the energy input is consumed by evaporation above oceans, while above continents the figure is only 50%.

The global average temperature of the air is not changing at present. The energy balance does not demonstrate any increment in the component: $J_1 = 0$. In average the basic equation of the energy balance, also taking into account the fact that the energy supply to the biosphere is relatively small, can be simplified as follows:

$$J_e = J_2 + J_3 \qquad (m^2.kg.s^{-2}) \qquad (1.2)$$

J_e - effective radiation

The result of the acceptance of the effective radiation are fluctuations in the soil and water temperature, accompanied by evaporation with sublimation.

These processes change the state of water aggregation into a gaseous one. The specific weight of water vapour is lower than that of air. Water vapour rises and in this way it acquires position energy. The thermal energy thus regenerates the mechanical energy of water and causes the circulation of water. The hydrological cycle is an uninterrupted process of water motion and changes of aggregation in the systems of the biological and abiotic environment. The difference between the specific and latent heat of fusion and vaporization, whose values are approximately two and three orders high respectively, balances this process during a higher or lower energy input.

The mechanical energy consists of the position energy, the pressure energy and the kinetic energy.

The position energy	$J_h = m.g.h$	$(m^2.kg.s^{-2})$	(1.3)
m - mass		(kg)	
g - gravitational constant		$(9.81\ m.s^{-2})$	
h - head		(m)	
Pressure energy	$J_p = m.\frac{p}{\varrho}$	$(m^2.kg.s^{-2})$	(1.4)
p - pressure		$(m^{-1}.kg.s^{-2})$	
- water density, unit mass of water		$(kg.m^{-3})$	
Kinetic energy	$J_k = m.\frac{v^2}{2g}$	$(m^2.kg.s^{-2})$	(1.5)
v - velocity of flow ,		$(m.s^{-1})$	

The quantity of water in water courses forms only 0.002% of the total global water reserves. The proportion of water power potential of water courses is only 0.4% of the $6.4\ .\ 10^{30}$ of energy which the Earth continuously receives from the Universe. But it is twenty times higher than the percentage of water courses volume in relation to total global water reserves because of the high head formed by geomorphological conditions.

Position energy acts as pressure energy and changes into kinetic energy, depending on the physical conditions. This kinetic energy together with chemical energy of water and changes in volume during ice formation, transforms soil and rock formations also forming and changing river beds. The growth and changes in ecosystems are also enabled by the effect of the mechanical, thermal and chemical energy of water.

By accepting and emitting energy, water molecules change their position and state of aggregation during their course through the biological and abiotic systems of the natural environment. The law of conservation of energy during this cycle expresses the equation of hydrological equilibrium:

$$\sum_{k=1}^{2} P_k = \sum_{k=1}^{3} Q_k + \sum_{k=1}^{4} E_k + \sum_{k=1}^{5} R_k \qquad (m^3) \qquad (1.6)$$

P_1 - vertical precipitation

P_2 - horizontal precipitation (see paragraph 1.3.2)

Q_1 - surface outflow (channel and overland flow)

Q_2 - subsurface outflow (groundwater runoff)

Q_3 - deep percolation and juvenile water inflow

E_1 - evaporation from bare soil surface

E_2 - evaporation from free water surfaces

E_3 - evaporation from snow and ice

E_4 - evapotranspiration

R_1 - water increment (or decrement) in soils and rock formations

R_2 - water increment (or decrement) in water courses and reservoirs incl. depression and detention storage

R_3 - water increment (or decrement) in the atmosphere

R_4 - water increment of the flora

R_5 - water increment of the fauna.

The area and period of application of this equation can be established in such a way that relevant increments or decrements in volume and the deep percolation or water supply from deep strata are negligible, thus simplifying the formula:

$$P_1 + P_2 = Q_1 + Q_2 + \sum_{k=1}^{4} E_k \qquad (m^3) \qquad (1.7)$$

This hydrological equation simply states that the total evaporation and the difference between the total inflow and outflow (concentrated and overland runoff, groundwater runoff) is formed by the precipitation and the dew deposit. Data on the earth's water reserves vary within a range of ±10%. KORZUN and SOKOLOV (1976) estimated them at 1,386 mld. km^3, of which some 2.53% or only 35 mil.km^3, are fresh water reserves. The total annual evaporation is 577,000 km^3: 505,000 km^3 on sea surfaces, 72,000 km^3 on continental surfaces. Groundwater reserves exceed five thousand times the amount of water in all rivers, brooks and creeks. 50% of the groundwater is below the level of 1000 meters under the earth's surface (Tab. 1.1).

Of basic importance in this context is the recycling rate, which indicates the duration of the natural exchange of the relevant volume of water: in the case of water courses this rate is $3.4 \cdot 10^4$ times as high as for groundwater.

TABLE 1.1

Type of formation	Area (10^6 km^2)	Volume (10^3 km^3)	Layer (m)	Share (%) on total water	fresh reserves	Period of replenishment (years)
Ocean	361	1 338 000	3700	96.5	0	2500
Brackish groundwater	134.8	12 870	96	0.94	0	1400
Lakes	0.822	85.4	103.8	0.006	0	17
Total	497	1 351 000	3660	97.45	0	-
Groundwater	134.8	10 530	78	0.76	30.1	1400
Soil water	82	16.5	0.2	0.001	0.05	1
Icebergs:						
Antarctic	13.98	21 600	1540	1.56	61.7	9700
Greenland	1.80	2 340	1298	0.17	6.7	9700
Arctic	0.23	83	369	0.006	0.24	9700
Mountain	0.22	41	181	0.003	0.12	1600
Permafrost	21	300	14	0.022	0.86	10000
Fresh water:						
Lakes	1.236	91.0	73.6	0.007	0.26	17
Marshes	2.682	11.5	4.28	0.009	0.03	5
Watercourses	148.8	2.1	0.014	0.0002	0.006	16 d
Biosphere	-	1.1	0.002	0.0001	0.003	1 h
Atmosphere	510	12.9	0.025	0.001	0.04	8 d
Fresh water reserves	148	35 029	235	2.55	100	-
Total water reserves	510	1 386 000	2719	100	2.55	-

World water reserves and the share of different water bodies in the reserve of the total volume of water and in the volume of fresh water according to Sokolov (1981). The period of exchange of their volume by natural recharge (d - days, h - hour)

Of the 145.10^3 km^2 of continental surfaces only two thirds (100.10^3 km^2) are suitable for water development, 25.10^3 km^2 being permafrost, 14.10^3 km^2 icebergs and 6.10^3 km^2 extremely arid land.

1.3 HYDROLOGIC CYCLE SYSTEM

The complicated processes of the hydrologic cycle include evaporation, precipitation, interception and surface storage, infiltration and percolation, surface and groundwater runoff.

The catchment area, i.e. the area which drains into one place and thus contributes to the runoff in the profile in question, is an open system whose boundary crosses energy, water, air and soil/rock particles. Potential energy of position, thermal and chemical energy within this system, is transformed into kinetic energy and heat. Water, suspended, wash and bed load as well as floating debris are transported from the upper elevations towards the sea (and partially vice versa e.g. by sand dune movement) and transformed. Erosion, crushing, chemical and biochemical processes are an integral part of the water cycle (Fig. 1.4).

The system of the catchment area tends to achieve a steady state of operation, corresponding to the conditions of climate, topography, geology and ecology, characterized also by the fact that the water and debris output corresponds to a specific energy input. Any change e.g. by river training, reservoir construction, land cultivation, urbanization and industrialization, is a change of system elements and of the energy input, thus resulting in the change of output.

The state of this system is to be followed in its spatial elements (Fig. 1.4). Three equations of balance can be formulated for each of these elements:

- hydrologic balance (Fig. 4.1)
- fall-out, erosion and debris balance
- energetic balance.

In any spatial element due to the equilibrium of the atmosphere branch, the precipitation, the evaporation and the increment of atmosphere moisture equal to the difference of water vapour entering and leaving the element:

$$A_n - A_{n+1} = R_n + P_n - E_n \qquad (m^3) \qquad (1.8)$$

A_n - water vapour entering the spatial element n

A_{n+1} - water vapour leaving the spatial element n

P_n - precipitation and dew deposit in the element n

E_n - evaporation in the element n

R_n - air moisture increment in the element n.

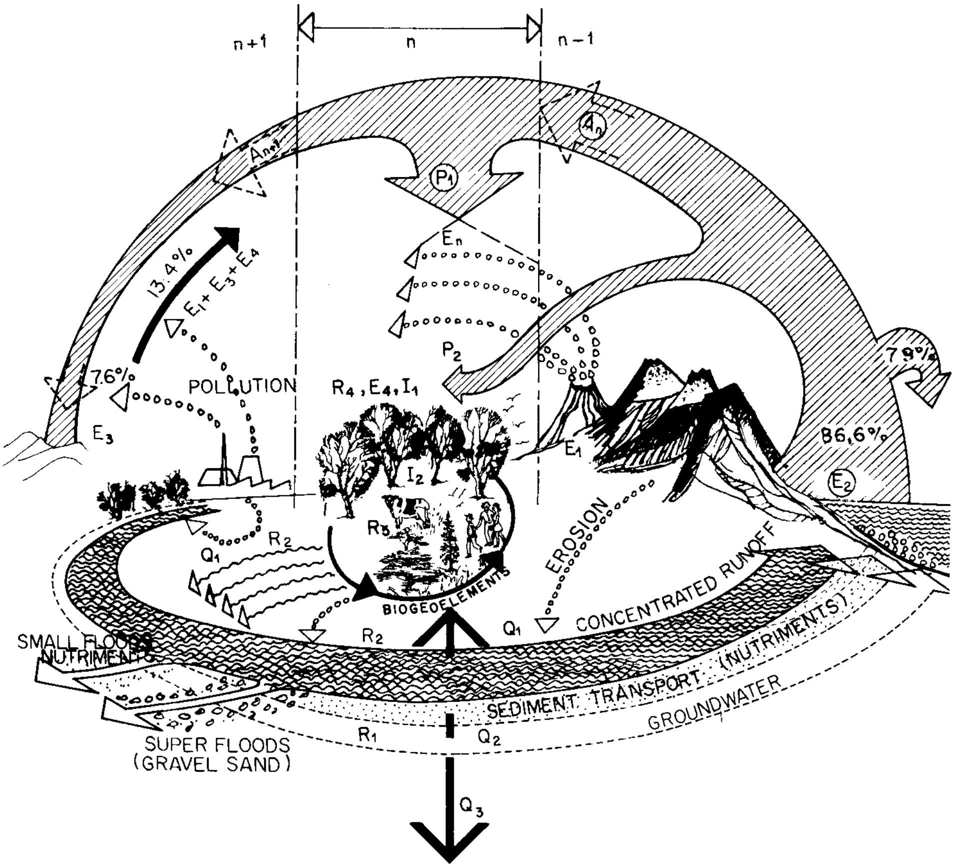


Fig. 1.4. Equation of hydrological equilibrium, the transport of mass and biogeochemical cycles as subsystems of the water cycle. Explanation of symbols and main equations is given in the text.

Precipitation is formed by external water vapour supply $A_n - A_n+1$ and by internal evaporation E_n:

$$P_n = A_n - A_{n+1} + E_n - R_n \tag{1.9}$$

but $P_n = Q_n + E_n$ (see equation 1.7) and, therefore,

$$A_n - A_{n+1} + E_n - R_n = Q_n + E_n$$

$$A_n - A_{n+1} = Q_n + R_n \tag{1.10}$$

Internal evaporation over continents is generally lower than the external water supply, requiring more energy and resulting in the generally prevailing evaporation on the sea surface. This fact can be expressed by the water circulation ratio, defined as

$$m = \frac{\sum_{n=1}^{N} P_n}{\sum_{n=1}^{N} A_n - A_{n+1}} = \frac{\sum_{n=1}^{N} P_n}{\sum_{n=1}^{N} Q_n + R_n} \qquad (1.11)$$

The value of this ratio over vaste areas is quite stable, e.g. for North America $m_{na} = 1.25$, Asia and Europe $m_{ae} = 1.51$.

1.3.1 Evaporation

Evaporation is a physical process by which water changes from the liquid state to the vapour state through the transfer of thermal energy. The change from solid state without passing the usual intermediate liquid aggregation is called sublimation.

Evaporation is the key process in the water cycle:

(a) it is the only one of the processes in this cycle during which the energy input exceeds the energy output,

(b) it accounts for the creation of living matter.

Evaporation takes place particularly on the boundary of the atmosphere and the hydro-, pedo-, and biosphere, thus making it possible to distinguish:

- evaporation from open water surfaces,
- evaporation from bare soil surfaces,
- evaporation from snow and ice,
- evaporation of water intercepted by vegetation,
- evapotranspiration from soil and vegetative cover,
- evapotranspiration of vegetative cover on water surfaces,
- evaporation from organic bodies and moist materials,
- evaporation in the atmosphere.

Evapotranspiration includes soil evaporation and the evaporation of water which is absorbed by crops, used in the building of plant tissue and transpired. The quantity of water evapotranspirated by plants and relevant soil surfaces per annum with the increment in plant tissue is the consumptive use of plants.

The hydrologic balance can be expressed generally or for a limited element of the lithosphere by the following equation:

$$ET = E + T = P + R_i + W - Q - F \quad (m^3/year, M^3.s^{-1}) \qquad (1.12)$$

ET - evapotranspiration

E - evaporation

T - transpiration

P - precipitation

R_i - irrigation

W - increase of water moisture caused by capillary forces from the groundwater

(+) or decrease by water consumption of crops (-)

Q - deep percolation and drained water

F - water chemically and biologically absorbed and used in the building of the plant tissue (+), or eliminated from the organic matter (-), e.g. by guttation.

The rate of evaporation depends on the state of the systems whose interaction enables its course. The atmosphere influences this course by meteorological factors, namely by solar radiation, humidity and by air movement leading away water vapours. The relationship between radiation and evaporation can be expressed for a limited part of the hydro- or the lithosphere (reservoir, forest, field etc.) by the equation

$$J_e = J_h + (J_a + J_x) \quad (J.m^{-2} \text{ per day}) \qquad (1.13)$$

J_e - effective solar radiation

J_h - heat accepted by the hydro- or litho- and biosphere

J_a - heat transferred back to the atmosphere

J_x - latent heat used for evaporation and evapotranspiration

The value and ratio of all these factors also change at the same place in the course of the year (Fig. 1.5)

$$\frac{J_a}{J_e} = \frac{\Delta T}{\Delta P} \qquad (1.14)$$

Albrecht (1951) simplifies the equation for evaporation

$$E = \frac{J_e - J_h}{1 + \frac{\Delta T}{\Delta P} \cdot \alpha} \quad (\text{mm}) \qquad (1.15)$$

ΔT - temperature increment (^{o}K)

ΔP - air pressure increment (Pa)

α - coefficient

The transfer of solar energy is greatly influenced by overshadowing and by meteorological factors, namely by the humidity of the air, precipitation, air temperature and pressure as well as the velocity of its movement. Temperature and air pressure do not influence evaporation directly. They characterize the quantity of accepted energy. The humidity and air flow which accelerate the evaporation by the exchange of saturated air strata above the evaporating surface are also incidental phenomena of the energy transfer to the atmosphere. They function as regulating factors of the evaporation and transpiration rate.

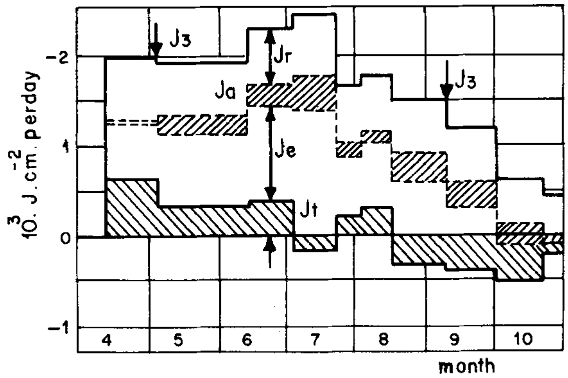


Fig. 1.5. The annual course of the energy balance (Lake Haussee according to measurements of Czepa and Schellenberger (1956). The energy input J_3 is partially radiated (J_r), partially transferred to the atmosphere (J_a) by the contact of the water table and the air mass. J_e is used for evaporation and J_t causes the change of water temperature.

The immediate cause of the evaporation process is the difference in humidity between the internal and external - or the saturated and unsaturated - system of environment. It can be characterized by the evaporation from surface as relative humidity, i.e. the ratio of the actual water vapour pressure e (Pa) and the maximum pressure E (Pa) which the air is able to accept at the actual temperature. The difference between these values is the saturation complement

$$d = E - e \qquad \text{(Pa)} \qquad (1.16)$$

Braslavskij and Vikulina (1954) assessed the following practical formula for the computation of the evaporation from open water surfaces on the basis of the air humidity and wind velocity

$$EV_a = 0.013 \,.\, (e_o - e_2) \,.\, (1 + 0.72\, v_2) \qquad \text{(mm per day)} \qquad (1.17)$$

EV_a - monthly average of evaporation

e_o - maximum water vapour pressure, corresponding to the average temperature of the water surface (m)

e_2 - monthly average of water vapour pressure 2 m above the water surface (m)

v_2 - average wind velocity at an elevation of 2 m above the water surface($m.s^{-1}$)

Šermer (1960) established the relation between the temperature and the evaporation from open water surface for the conditions of Central Europe as follows

$$EV_d = 10^{(0,0452\ T\ -\ 0.104)} \qquad \text{(mm per day)} \qquad (1.18)$$

T - average monthly temperature 2 m above the water surface (^oC)

Evaporation from snow and ice is five to ten times lower than evaporation from free water surfaces. Due to the lower temperature and solid state of snow and ice, much more energy is needed for the same intensity of its course. The evaporation rate from ice is about 50 to 100% higher than the evaporation from snow under the same conditions, because the heat conductivity of snow is lower. Therefore

$$EW > EI > ES \qquad (mm) \qquad (1.19)$$

EW - evaporation from open water surface
EI - evaporation from ice
ES - evaporation from snow surface.

The value of evapotranspiration from overgrown water surfaces depend on the kind and the total quantity of the vegetable matter. The evaporation from overgrown surfaces does not differ greatly from the evaporation from open water surfaces, when the water surface is only covered by floating leaves. But it exceeds it twice in the case of the densely overgrown edges of reservoirs. Therefore

$$EO \geqq EW \qquad (mm) \qquad (1.20)$$

EO - evapotranspiration from overgrown water surface.

Evaporation from bare soil does not depend on heat input only, characterized by meteorological factors, but also on the soil factors, namely on

- the structure and other physical properties of the soil,
- the soil moisture,
- the contact of the soil layer with the groundwater surface (Fig. 1.6).

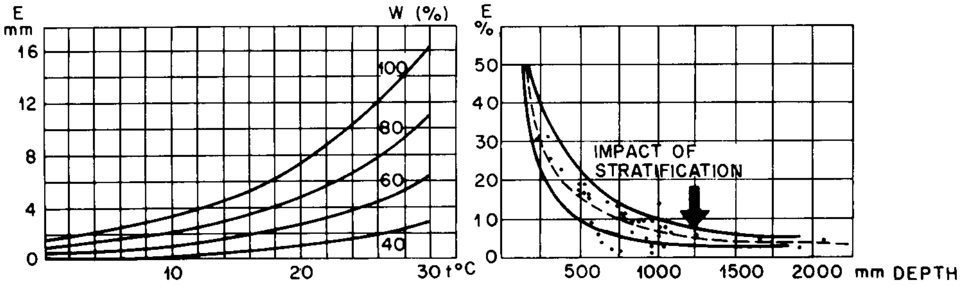


Fig. 1.6. (a) Interrelation of the evaporation from the free water surface, the air humidity and the air temperature according to Dub (1957). (b) Relation of the evaporation from the groundwater table on its depth, expressed as the ratio of the evaporation from free water surface. Derived according to White (1970).

These internal factors function clearly in the case of lower soil saturation. The evaporation from bare soils also depends on the velocity of water inflow to the surface. Inadequate water inflow lowers the evaporation rate.

Actual evaporation from bare soils is, therefore, lower than the potential rate, whose value depends on the energy supply only.

Values of the potential evaporation from bare soils almost equal those from open water surfaces, when water evaporates directly from the wetted topographic surface. Under conditions of a dry surface layers, as Penman (1940) proved, water vapour penetrates this layer by diffusion, which lowers the values of evaporation.

Evaporation from bare soils takes place

(a) in contact with the groundwater surface, regulating the soil moisture, or, more frequently,

(b) without outstanding contact with the groundwater surface, when the suspended capillary water of the soil profile is not connected with the capillary water supported by the groundwater level, and the root system does not penetrate into this space, i.e. when the groundwater level is influenced by the conditions of the soil surface by means of hygroscopic and osmotic forces and by the gas pressure only.

Evaporation from groundwater surfaces depends namely on the depth of the groundwater level. The course of groundwater level changes is not the same under the conditions of evapotranspiration: Relevant forces differ, especially during the day. They also depend on the kind of vegetation, its root system, stage of growth and quantity of leaves.

Evaporation within reach of a well can be calculated by neglecting the evaporation during the night, which is comparatively low, anticipating that the rise in water level is uniform:

$$E_1 = Q_1 \cdot (24 \cdot v - s) \qquad (1.s^{-1}) \qquad (1.21)$$

E_1 - evaporation from the groundwater surface within the reach of the measured well (1.s^{-1})

Q_1 - well yield during the decrease of water level by 1 m (1.s^{-1})

v - velocity of water level rising during night (m per hour)

s - total increase of water level per day (m per day)

In the most frequent case of evaporation from bare soils without outstanding contact with the groundwater level, the value of the evaporation rate in the initial stage is almost equal to the potential evaporation. The following stage, beginning with a substantial lowering of the moisture of the soil surface, is characterized by the decreasing velocity of evaporation, which stabilizes in the final stage at a low value. Anticipating a uniform distribution of the perpendicular velocity, Kutílek (1978) proves that

$$e = (W_i - W_o) \cdot \left(\frac{\bar{D}}{\pi \cdot t}\right) \qquad (m.s^{-1}) \qquad (1.22)$$

e - evaporation rate $(m.s^{-1})$

$W_i - W_o$ - difference of moisture content in the soil surface during time t (%)

$\overline{D}$ - average difussivity of the soil water, depending on the soil category and moisture $\overline{D} = \frac{k_f \, . \, dh}{dW}$

k_f - hydraulic conductivity

h - pressure height

t - time

π - constant of soil properties.

Transpiration, the evaporation of water absorbed by a crop and not used in the building of plant tissue, may be above all stomatal but also appears as cuticular or as guttation or exudates from cut surfaces of the plant. Transpiration depends on physiological and environment (meteorological) factors, e.g. daytime. Physiological factors include

(a) the physiological structure of relevant plant types, the age of their organs and the nature of their cellular membranes,

(b) the actual state of the relevant individual plant, i.e. the degree of nutrition, namely water content of its cells and water vapour content in the transpiratory organs.

Under conditions of sufficient moisture and nutrition for the development of individual plants, the intensity of transpiration depends especially on environmental factors, namely on solar radiation, wind velocity and soil moisture. The temperature of the leaf exposed to the sun is higher than that of the air. In the case of an insufficient supply of water and nutrition, the intensity of transpiration depends more on the above-mentioned internal physical factors (Tab. 1.2).

The plant exercises a limited control on the transpiration rate. Stomata usually open in the light. They close with reduced moisture and when the sugar content decreases, changing to starch, as happens in the dark or at the end of the vegetation season, when leaves turn yellow. Transpiration is also reduced in the case of abundant water.

The movement of water from the root zone, through the stem and leaves, is enabled by diffusion and osmosis. The rate of both these processes is influenced by air moisture and energy supply, resulting in the removal of water vapour next to the leaf surface. Van Den Honert (1948) expressed transpiration by physiological analogy with OHM's law

$$T = \left(\frac{\psi_s - \psi_e}{r_s - r_p}\right) . \, \gamma_v \qquad (m.s^{-1}) \qquad (1.23)$$

T - transpiration rate

TABLE 1.2

Term, sign	Definition	Usage
Potential evaporation (evaporativity) ET_1	Theoretical value derived from energy input $ET_1 = \frac{\Delta p}{\Delta p + \gamma} (J_s + J_g)$ Δp – saturated vapour pressure increment γ – psychrometric constant J_s – energy input J_g – energy transfer	Energy budget
Potential evapotranspiration ET_p	Evapotranspiration from soil and vegetation system, saturated with water and nutriments, derived from local hydrometeorological conditions.	Water balances for long term planning
Optimal evapotranspiration ET_{opt}	Evapotranspiration from a surface whose soil moisture is managed in order to increase agricultural and forestry production	Determination of plant water and irrigation requirements
Maximum evapotranspiration ET_{max}	Highest evapotranspiration that relevant vegetation system is able to achieve, dependent on its stage of growth and actual state	Resistance against moisture. Dimensioning of the drainage.
Minimum evapotranspiration ET_{min}	Evapotranspiration of a plot irrigated only for survival of relevant plant species	Resistance against drought
Actual evapotranspiration ET_a	Real evapotranspiration dependent on the growth stage and state of the plant, measured by soil-moisture sampling, large-size lysimeters, groundwater fluctuations	Determination of actual irrigation rates

Glossary of evapotranspiration.

TABLE 1.3

Cover	Bare soil			Grassland			Pine forest		
	mm	%	Share	mm	%	Share	mm	%	Ratio
October	9.6	5.4	1.00	18.2	5.1	1.75	28.6	6.6	2.98
November	2.6	1.5	1.00	7.0	1.9	2.69	13.0	3.0	5.00
December	0.6	0.3	1.00	2.0	0.5	3.33	8.5	1.9	14.20
January	0.2	0.1	1.00	3.7	1.0	18.50	12.1	2.8	60.50
February	4.2	2.4	1.00	5.1	1.4	1.21	13.4	3.1	3.18
March	10.2	5.7	1.00	16.7	4.7	1.64	25.4	5.8	2.49
April	23.1	13.0	1.00	37.6	10.5	1.63	41.0	9.4	1.78
May	23.9	13.5	1.00	62.6	17.5	2.62	69.9	14.7	2.92
June	24.4	13.7	1.00	51.6	14.6	2.12	58.6	13.4	2.40
July	29.9	16.8	1.00	57.4	16.4	1.91	61.5	14.2	2.05
August	27.0	15.2	1.00	55.2	15.5	2.04	61.1	14.0	2.26
September	22.1	12.4	1.00	39.5	11.1	1.78	48.4	11.1	2.18
Yearly total	177.8	100.0	1.00	355.6	100.0	2.00	435.5	100.0	2.44
Share from free surface evaporation	Loamy soil		50%	Loamy soil		87%	Depending on forest density and age		70%
	Sand		26%	Sandy soil		26%			100%

Seasonal distribution of evaporation and evapotranspiration, its dependence on the soil surface and cover according to Wechman (1963, Eberswalde, GDR). The share of evapotranspiration compared with the free surface evaporation according to Krecmer (1980).

e_t - evaporation rate $(m.s^{-1})$

Ψ_s - soil moisture potential - suction pressure of the soil water $(J.kg^{-1})$

Ψ_e - moisture potential of the leaves $(J.kg^{-1})$

r_s - flow resistance of the soil $(Pa.s^{-1})$

r_p - flow resistance of the plant $(Pa.s^{-1})$

γ_v - unit mass of the soil water $(kg.m^{-3})$

The ratio of transpiration and evaporation from soil changes at one point in time, namely during the vegetation season. At the beginning of this season, the evaporation from bare soils dominates. Transpiration increases with the growing vegetation. Under the conditions of coherent plant cover, transpiration generally prevails. Overshadowing of the soil surface by the vegetative canopy decreases the soil surface temperature and the rate of water vapour removal, thus causing a decrease in the evaporation rate. Transpiration also drops during the period of ripening.

Evapotranspiration is a complicated phenomenon, explicable by a few values only (see Tab. 1.2) complying with the following unevenness

$$ET_l > ET_p > ET_m > ET_o > ET_a > ET_{min} \tag{1.24}$$

Under Central European conditions evapotranspiration by vegetation is generally higher than evaporation from bare soils, but lower than evaporation from open water surfaces. Evapotranspiration from arid zone plants is often lower than evaporation from bare soils. Gilimeroth (1951) and Kramer (1969) state that evapotranspiration from vegetation never exceeds evaporation from saturated soils at the same level of exposure.

1.3.2 Precipitation

The precipitation process is the transfer of water eliminated from the atmosphere system to the system of the hydro- and lithosphere, characterized by an output of the latent heat of vaporization.

The number of different forms of precipitation is very large, but basically precipitation can be

(a) vertical, i.e. precipitation from the upper part of the atmosphere system, characterized by a vertical movement of drops: drizzle, rain, snow, glaze, hail, sleet etc.,

(b) horizontal, i.e. water eliminated by condensation or sublimation on the ground: dew, hoar frost, rime, diamond dust etc.

Depending on its stage of agglomeration, precipitation is either liquid or solid. It is measured in terms of depth (mm), rainfall intensity in terms of mm per minute.

Rainfall is produced by a cooling of the air as the result of a decrease in the barometric pressure, by radiation, by contact with a colder land or sea surface or during mixing of air masses. Condensation of water vapour into cloud droplets takes place on condensation nuclei, formed by hygroscopic salt particles. The falling speed of droplets is a function of their size and of the speed of the air stream. The coalescence of the droplets to form raindrops is accounted for by the coexistence of the ice crystals and water droplets and by the differences in speed between large and small drops. The latent heat of evaporation regulates the process of condensation (Fig. 1.7).

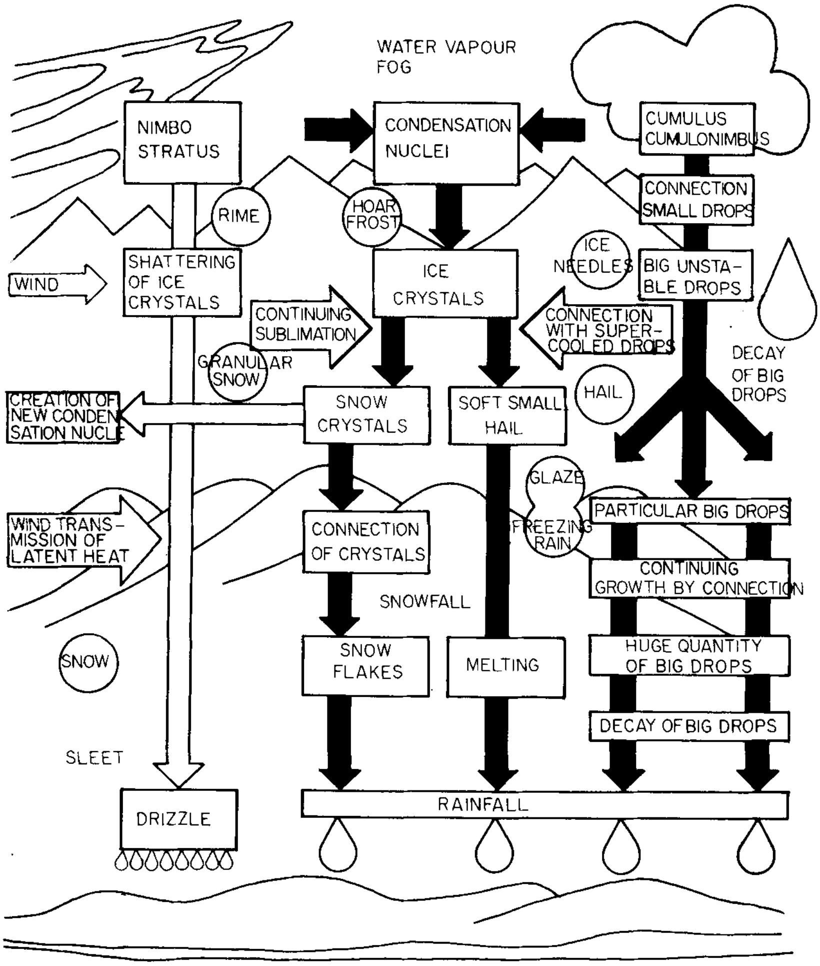


Fig. 1.7. The subsystem of precipitation and its processes in dependence on the altitude according to Mason (1957).

To summarize the tendency of rain to occur in definite patterns, it is possible to distinguish between:

(a) local convective rainfall - caused by the upward movement of air masses,

(b) orographic rainfall - influenced by the morphology of the exposed part of the higher mountain ranges and characterized by longer duration and lower intensity,

(c) cyclonic rain - caused by the air-mass contrast of cold and warm rain, i.e. by the excess of surface heating in lower latitudes and of cold in higher latitudes, characterized by moderate lasting rainfall over a large area, but also by heavy rain, hail or snowfall over a small area.

Rainfall often occurs as a combination of the above mentioned forms. Its occurrence, intensity and frequency depends on zonal, regional and local factors.

The contrast between surface heating in the equatorial and polar zones, or the difference in temperature between the continent and the sea, causes substantial movement of the air, which manifests the homeostasis of this system, i.e. its tendency to achieve a balanced, stable state. This movement of air is influenced by the rotation of the Earth and by regional thermal and orographic factors. Regions with a high horizontal inflow of water vapour and an upward movement of air masses are characterized by frequent precipitation.

The important influence of a region's latitude on the rainfall frequency is evident:

(a) the equatorial zone has the highest annual precipitation on average (50% of the global rainfall occurs between 20° N.L. and 20° S.L.),

(b) areas which have considerable rainfall cover middle and higher latitudes, oceans and the western parts of continents,

(c) areas with passat winds and strips along the tropics are rainless.

(d) polar areas are also rainless, obtaining 4 % of the total global precipitation.

In certain climatic regions, precipitation and its state of aggregation depends to a great extent on the altitude (Fig. 1.8). The increment in the rainfall total, corresponding to the difference in altitude, is the rainfall gradient.

Local factors which influence the spatial distribution of rainfall include geographical and morphological factors, such as the area exposure (to the direction of wind), the characteristic of its surface (roughness, vegetative canopy), and the gradient of the slope. Rainfall intensity, very often unevenly distributed in space and time, decreases on average with the areas affected and with the rainfall duration. The unevenness of the space-time distribution of the rainfall results in a fluctuation of precipitation during the year and in a fluctuation of annual average. During periods of defficient precipitation the deviation from average is greater for river runoff than for rainfall.

The complicated mechanism of precipitation can be influenced in particular by

- changes in the heat input,
- changes in the air mass movement,
- increasing or decreasing the quantity of condensation nuclei.

The result of any of these measures is not explicit, because the relevant interrelationships of the atmospheric system are complicated and a certain feedback exists, such as the atmospheric system's external relations with the hydro- and lithosphere and the solar system, whose energy supply is not uniform.

1.3.3 Interception

Interception is a process of precipitation transmission and redistribution on the boundary of the systems of the atmosphere and the lithosphere by the vegetative canopy. The quantity of precipitation which actually reaches the ground, effective rain- and snowfall, consists of the

- throughfall, which reaches the ground directly through intershrub spaces and drips from the leaves, twigs and stems, and of the
- stemflow which reaches the ground by running down the stems.

Interception loss is the part of precipitation retained by the vegetative canopy and then evaporated or absorbed. Therefore

$$P_e = P - I \qquad \text{(mm)} \qquad (1.25)$$

P_e - net precipitation
P - actual precipitation (above the vegetative canopy)
I - interception loss

$$I = I_1 + I_2 \qquad \text{(mm)} \qquad (1.26)$$

I_1 - interception by the aerial portion of the vegetative canopy
I_2 - interception of the layer of shedded leaves and needles

$$P_e = T + S - I_2 \qquad \text{(mm)} \qquad (1.27)$$

T - throughfall
S - stemflow

Interception loss during one single rainfall consists of the interception capacity of the surface of leaves, twigs etc. and the amount evaporated and absorbed by plants:

$$I = I_o + I_{ea} \qquad \text{(mm)} \qquad (1.28)$$

I_o - interception capacity of leaves, twigs etc.

I_{ea} - interception loss by evaporation and absorption

Absorption of water by plants during one single rainfall is negligible. For this reason Linsley derives the following equation for the interception during one single rainfall:

$$I = I_o + \alpha . ET . t \qquad (mm) \qquad (1.29)$$

α - ratio of the total evapotranspiration and the evaporation from the vegetation surface, depending on the ratio of the vegetative and non-vegetative surface (s^{-1})

ET - evapotranspiration (mm)

t - rainfall duration (s)

The interception capacity depends on the composition of the relevant levels of the vegetative canopy, its morphology and development stage. This capacity, which can be reduced by preceding rainfall, influences the net precipitation in dependence upon the actual rainfall intensity, duration and course as well as upon the wind velocity. An overfulfilling of this interception capacity is characterized by a remarkable increase of stemflow and throughfall (dripping). It goes without saying that the interception loss may exceed the interception capacity.

Zinke (1967) estimates, without including the capacity of the shedded leaves and needles, the average interception capacity of most grasses, trees and shrubs at 1.3 mm during one single rainfall and 3.8 mm during snowfall. He also states that the interception loss is twice as high in 20% of the observed cases. The average interception loss of a certain area depends not only on the composition of the vegetative canopy, its development stage and actual state, but also on the time distribution of the precipitation and the interplay of the rainfall occurrence with the course of temperature, humidity and wind velocity.

1.3.4 Depression and Detention Storage; Overland Flow

The effective precipitation reaching the earth's surface is partly stored

(a) after snowfall as snowpack, whose further effect on runoff depends on energy supply, i.e. on

- radiant heat from the sun
- latent heat of vaporization released by the condensation of water vapour,

(b) by depression storage in surface puddles and by surface detention formed by a sheet of water on the soil surface.

(c) as channel storage in stream channels, ponds, swamps, etc.

Depression storage is not directly measurable and even detention storage is usually derived from hydrograph analysis rather than from observation. Surface runoff usually commences from one part of a catchment area before the interception and depression storages in other parts are satisfied. Detention storage depends on the slope and surface roughness of the area, i.e. on the soil conditions, the vegetative cover and its state. The difference between types of vegetation are caused by the effects of the litter, which appears to be more significant than irregularities in the soil surface.

The surface runoff does not occur whenever the rainfall intensity does not exceed the infiltration and evaporation intensity. In this case the surface runoff does not occur only during the first part of the storm, when the interception, depression and detention storage capacities are not exceeded. As the rain continues, puddles become full and the soil surface becomes covered with a sheet of water and downhill flow begins towards an established surface channel.

A level plain can accumulate 3-18 mm of water, meadows and fields 12-42 mm and forests much more water, which gradually infiltrates and evaporates. When these limits are exceeded, spatially varied unsteady flow during rainfall occurs, in which the rate and depth of flow increase down the length of the flow path. This depth also increases with time, even when the intensity of the rainfall remains unchanged. For these conditions the relationship becomes

$$D_e = K \,.\, L \,.\, q \qquad (m^3) \qquad (1.30)$$

D_e - volume of detention when equilibrium flow condition is established (m^3)

L - length of flow (m)

q - discharge per meter width at equilibrium $(m^3.s^{-1})$

K - coefficient of rainfall intensity, slope and roughness of the surface

Where steady uniform overland flow is considered (Tab. 1.5), the following relationship between rate of discharge and depth of overland flow can be theoretically derived:

$$Q = K \,.\, H^m \qquad (m^2.s^{-1})$$

Q - overland flow $(m^2.s^{-1})$

H - depth of flow (m)

m - coefficient of slope and roughness (involving viscosity,

$m = 3$ for laminar flow, $m = 1.67$ for turbulent flow)

1.3.5 Infiltration

Infiltration is a process of unsaturated or saturated flow during the movement of water into the pedo- and lithosphere, detructing the soil water and groundwater from the net precipitation. Water tries to achieve a state of minimum energy in these systems and moves from levels of higher energy to levels of lower energy.

Saturation depends on the porosity of soil or rock and the moisture content. When the moisture content is smaller than the porosity, the flow is unsaturated. When it equals the porosity, the flow is saturated. Its rate depends on the effective porosity, which is usually expressed as a percentage and defined by

$$n = \frac{V_v}{V_o} \cdot 100 \qquad (\%) \qquad (1.30)$$

n – effective porosity (%)

V_v – volume of water governed by gravity forces in the saturated soil or rock (i.e. the volume of all connected effective pores and voids) (m^3)

V_o – total volume of the soil or rock (the volume of all pores and voids plus the volume of all the grains and solids) (m^3)

The effective porosity is a part of the total porosity which enables the gravitational movement of water. It depends on soil texture and structure (grain-size distribution, mutually connected pores and cracks etc.) Infiltration also depends on the state of the soil surface incl. density of vegetation, moisture distribution in the soil layer, the air content in non-capillary pores, the temperature, the depth of the groundwater table and the intensity of the rainfall (high intensity rainfall causing compaction of the surface level).

The infiltration rate is the maximum rate at which the soil can absorb precipitation in a given condition. The initial high rate of infiltration decreases exponentially: rapidly at the beginning and then more slowly until it approaches a constant rate after a period of 20 to 120 minutes.

Philip (1958) expresses the actual infiltration rate by the formula

$$v_i = \frac{1}{2} \cdot S \cdot t^{-\frac{1}{2}} + v_k \qquad (m.s^{-1}) \qquad (1.33)$$

v_i – actual infiltration rate ($m.s^{-1}$)

S – sorptivity ($m.s$)

t – time of the beginning of infiltration (s)

v_k – final stable infiltration rate ($m.s^{-1}$)

Sorptivity can be defined by the equation

$$S^2 = 2\, k_f' \cdot (\varpi + H) \cdot (W_o - W_i) \qquad (m^2.s^{-1}) \qquad (1.34)$$

k_f' – coefficient of the hydraulic (unsaturated) conductivity

$$k_f' = \frac{k_f \cdot W^3}{n^3} \qquad (m.s^{-1}) \qquad (1.35)$$

ϖ – soil characteristic (m)

W_o - final moisture content (%)

W_i - moisture content at the beginning (%)

H - depth of groundwater table (m)

Philip (1969) expresses the total value of infiltration by the sequence in which the first two component prevail

$$W_n = S \cdot t^{\frac{1}{2}} + v_k \cdot t \quad (\mathrm{mm}) \qquad (1.36)$$

W_n - infiltration total (mm)

The following processes of unsaturated subsurface flow are interconnected with the infiltration

- redistribution, when soil water enters the layers with lower moisture content,
- percolation, when water leaves the saturated soil layers and enters the groundwater,
- capillary rise, when the moisture of the upper layers is supplemented from the lower ones, or from the groundwater.

1.3.6 Subsurface Water Movements

Subsurface water forms the subsurface hydrosphere in the heterogeneous environment of the soil and hydrogeological structures, which occurs in different forms (Tab. 1.5). The subsurface hydrosphere is formed by:

- soil water, occurring in the upper 2-4 m layer on the boundary of the atmosphere and the lithosphere. Water is retained in soil by surface-tension forces, which are molecular (electrical) by nature, i.e. other than those of gravity. The outflow of free water from soils occurs only if the pressure in the soil water exceeds the atmospheric pressure. Soil water is, therefore, unsuitable for water extraction, but indispensable for the photosynthesis of all plants.
- groundwater in the permeable formations of the Earth's crust, retained especially by gravitational forces and, therefore, usable for extraction (Fig. 1.8).

Permeable geological formations are known as aquifers and water occurs in their internal void space, forming

(a) voids or pores, i.e. subtle, microscopic spaces, which originated simultaneously with the associated rocks,

(b) cracks, i.e. breaches and other generally multi-directional spaces of secondary, tectonic origin,

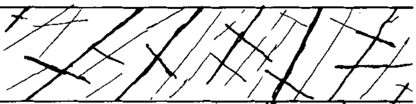
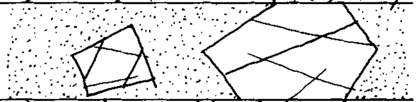
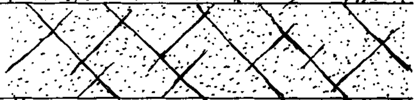
(c) cavities, or spaces of exceptional dimensions, originating mainly in carstic formations.

TABLE 1.5

Form	Prevailing forces	Occurrence	Movement
Water vapour	pressure	gaseous state	by pressure gradient, wind power
Ice, snow	gravity	solid state	by gravity, in soil after heat input
Gravitational	gravity	earth surface, pores, fractures, cavities	by gravity and tidal forces, pressure, osmotic and temperature gradient
Capillary	surface tension	- suspended - supported - no contact with gravitational	held in the interstices, available for plants
Adsorbed	attractive (surface potential)	hygroscopic viscous	oven-drying at 105°C
Structural	molecular	crystallic chemically combined	after disintegration/integration
	biochemical	biological	after disintegration

Modes of water occurrence above and under the ground.

TABLE 1.6

Types of interstices	Graphical interpretation
fractures-fractures	
fractures-pores	
pore-fractures	
pores-pores	
pores-cavities	
fractures-cavities	

Categorization of combined permeability according to Landa (1980).

Part of the water infiltrated into the soil flows laterally at shallow depths as interflow owing to less pervious lenses below the soil surface.

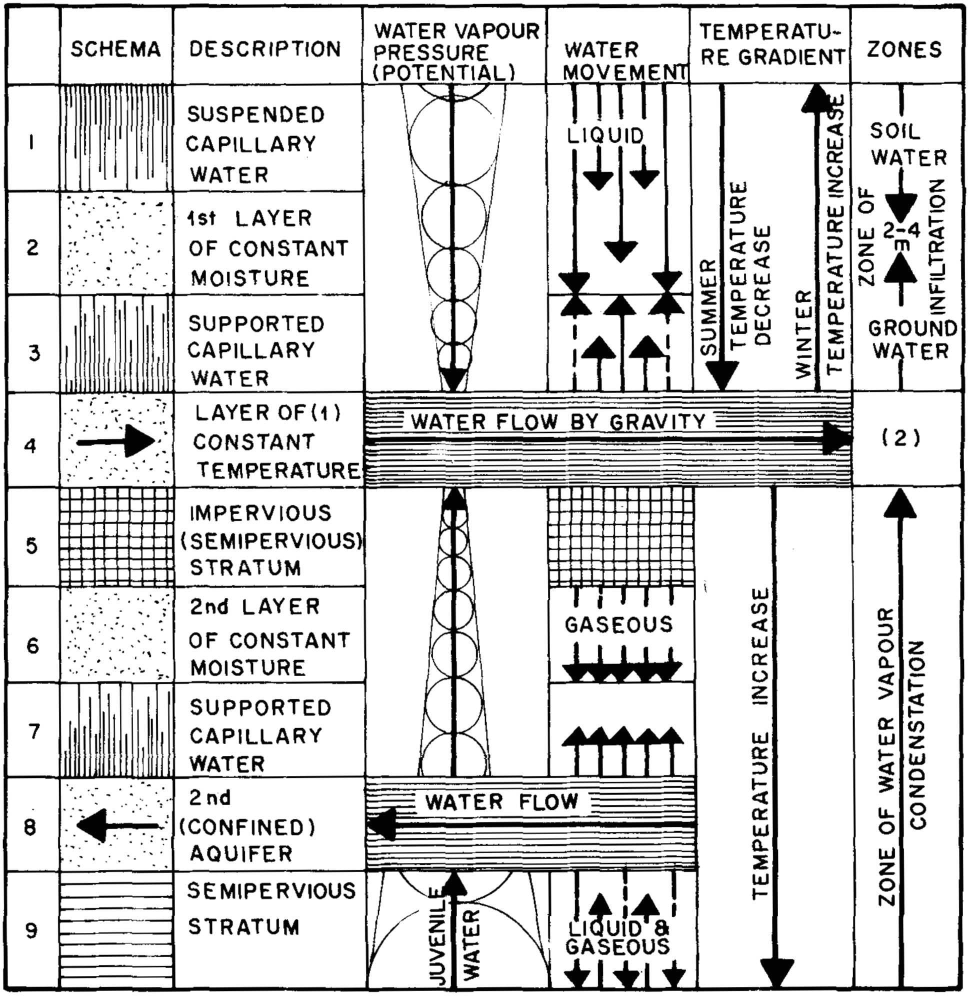


Fig. 1.8. Schematic representation of a characteristic arrangement of groundwater strata. (1) 1st (unconfined) aquifer, (2) depends on geographical length and geological structure. The size of circles is proportional to the pressure (potential).

Voids, cracks and cavities form extremely complicated underground spaces, which are separated or interconnected and which communicate effectively or non-effectively. Water in these internal spaces, whose permeability is combined (Tab. 1.6), is influenced by

(a) gravity - acting as the water weight, - acting as the pressure of surrounding geological formations,

(b) pressure of gases emitted by the water

(c) surface-tension forces (capillarity)

(d) molecular forces of the soil or rock particles (hygroscopic forces etc., primarily electrical in nature)

(e) osmotic forces, caused by the different quality (chemical composition) of water in different parts of the geological formations.

Unless these forces are in a state of equilibrium, groundwater is in movement and also influenced by

(f) friction forces, caused by the roughness of the surface of the soil or rock particles,

(g) internal friction forces caused by the fluid viscosity.

The flow in mutually communicating voids, cracks and cavities is in detail non-uniform and unsteady. For practical purposes it can be considered as uniform and steady on average. For groundwater movement Darcy's law is applicable within the laminar range of flow where resistive forces govern flow and the soil/rock environment is saturated

$$v_f = \frac{Q}{A} = k_f \cdot I \qquad (\text{m.s}^{-1}, \text{ m per day}) \qquad (1.37)$$

v_f - apparent velocity of flow

Q - flow rate $(\text{m}^3.\text{s}^{-1}, \text{ m}^3 \text{ per day})$

A - cross-sectional rate

$I = \frac{dh}{dL}$ - hydraulic gradient

k_f - coefficient of hydraulic conductivity (Tab. 1.6)

As velocity increases, inertial forces change the linear relation to the apparent velocity of flow at the hydraulic gradient to the

$$v_f = k_f \cdot I^{\frac{1}{m}} \qquad (\text{m.s}^{-1}, \text{ m per day}) \qquad (1.38)$$

m - coefficient $\rightarrow 2$

Similar equations can be derived for unsaturated flow

$$v_f = - k'_f \cdot \text{grad}\,\Psi \qquad (\text{m.s}^{-1}, \text{ m per day}) \qquad (1.39)$$

$\text{grad}\,\Psi$ - gradient of the total potential of the groundwater

k'_f - coefficient of the unsaturated flow (m.s^{-1})

TABLE 1.6

Soil	Coefficient of hydraulic conductivity k_f ($m.s^{-1}$)	Capillarity (mm)	Average porosity
Clay	1.10^{-8}	2000 - 4000	50 - 95
Silt	$5.10^{-6} - 1.10^{-7}$	700 - 1500	40 - 60
Compacted loamy sand	$1 - 5.10^{-6}$	350 - 700	15 - 25
Fine and loose sand	$1 - 5.10^{-5}$	50 - 350	20 - 45
Coarse sand	$1 - 5.10^{-4}$	10 - 50	25 - 35
Sandy gravel	$2.10^{-4} - 1.10^{-3}$	-	20 - 40
Clean gravel	1.10^{-2}	-	25 - 35

Coefficients of hydraulic conductivity and average porosity and capillarity of different soils.

The total potential of the groundwater is the amount of energy needed to transfer a unit of water quantity from one place in the system water-rock/soil to another one:

$$\Psi = \sum \Psi_k = \Psi_k \cdot F_k \cdot dl \qquad (J.kg^{-1} = m^2.s^{-2}) \qquad (1.40)$$

Ψ - total potential of water in the force field ($J.kg^{-1}$)

F_k - unit force of the force field ($N.kg^{-1}$)

dl - distance

On the basis of the definitions and formulas in Tab. 1.7, the total potential of groundwater under isothermic conditions is

$$\Psi' = g \cdot (x+z) + \frac{1}{\delta} \cdot (\Delta P - \delta) \qquad (J.kg^{-1}) \qquad (1.41)$$

The groundwater movement is spatial in character. Depending on the governing potential, the regime of flow can be

(a) hydrodynamic - where gravitational and pneumatic forces are governing,

(b) hydrothermal - where the difference in temperature is governing,

(c) hydrochemical - where osmotic forces are governing.

The hierarchy of these groundwater movement regimes is interconnected with the values of the associated potentials, which used to be remarkably different. The regime of groundwater flow depends on the homogeneity of the geological formations. The ratio of permeability of the relevant formations and their integrated parts influences the creation of the flow regime. Several regimes e.g. local, areal and regional, can occur in a heterogeneous environment (Fig. 1.9).

TABLE 1.7

Symbol	Potential	Forces	Equation	Explanatory notes
Ψ_g	gravitational	gravity	$\Psi_g = g.dz = g.z$	g - gravitational constant z - head (m)
Ψ_c	capillary	capillarity	$\Psi_c = g.x$	x - capillary rise
Ψ_p	pneumatic	pressure gradient of soil gases, atmospheric	$\Psi_p = \frac{1}{\gamma}.\frac{\partial p}{\partial l} . dl = \frac{1}{\gamma}. \Delta P$	γ- unit weight of water ($kg.m^{-3}$) ΔP- pressure gradient ($Pa = kg.m^{-1}.s^{-1}$)
Ψ_t	thermic	water density gradient	$\Psi_t = \frac{1}{\gamma} .\Delta T$	ΔT- temperature gradient
Ψ_o	osmotic	difference in chemical composition	$\Psi_o = \frac{1}{\gamma} . \delta$	δ - osmotic pressure (Pa)

Categorization of soil water potential.

The function of different forces in the heterogeneous system of hydrogeological formations depends on external factors, including

- climatic and meteorological factors,
- surface run-off,
- variations and oscillations in the interconnected surface water levels of water courses, reservoirs, lakes and seas,
- external load.

Where the surface water is not in contact with an unconfined aquifer, the precipitation produces the governing influence. Seasonal variations in rainfall and changes of groundwater in storage, manifestated by changes in groundwater tables, are closely correlated. This correlation is heavily influenced by the surface run-off: the groundwater recharge depends on the rainfall intensity and distribution. The same monthly averages may produce different fluctuations in the water table.

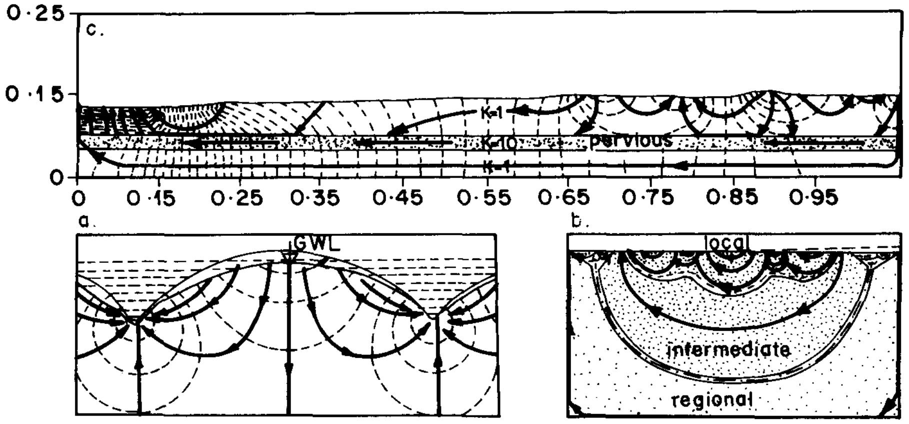


Fig. 1.9. Regional flow of groundwater (flow direction marked fully, equipotentials dashed, system boundaries dash- and dotted): a - homegeneous permeable strata according to Hubbert (1940), b - homogeneous isotropic strata according to Toth (1962), c - heterogeneous strata according to Freeze, Witherspoon (1966): local regimen dotted densely, intermediate medium, regional regimen dotted scarcely.

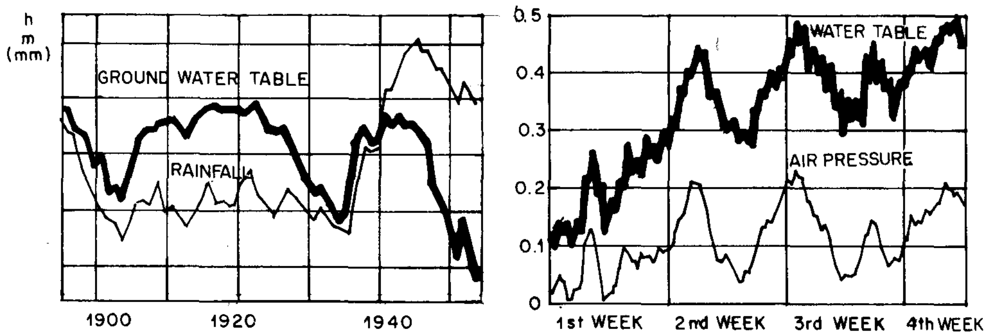


Fig. 1.10. Reduction of values and time delay of groundwater fluctuation in relation to the rainfall occurrence (deviation from the average) according to Todd (1970). Relationship of the air pressure and the water table fluctuation in an artesian well according to Robinson (1939).

Atmospheric pressure has no effect on unconfined aquifers. In the case of confined aquifers, increases in atmospheric pressure cause a decrease in water tables and vice versa (Fig. 1.10):

$\Delta h = \eta . \Delta p_a$ (m) (1.42)

Δp_a - change in atmospheric pressure (m of water)

Δh - water level decrease or increase

1.3.7 Flow in Channel Network

Overland flow is gradually concentrated by the topography of the Earth's surface. Flow in natural channels whose profile and head is not stable due to erosion and siltation is generally non-uniform and unsteady. The discharge and the medium average velocity are functions of time and space.

The discharge in natural channels can for practical purposes be considered as gradually varied. In this case, the total head Δh at a channel section can be expressed as

$$\Delta h = \frac{v_s^2 \,.\, L \,.\, n^2}{R} + \frac{(v_2^2 - v_1^2) \,.\, k}{2g} \tag{1.43}$$

$$v_s = \frac{v_1 + v_2}{2} = \frac{R^{\frac{2}{3}}}{\sqrt{2g.L}.n} \cdot \left[2g.\,\Delta h + k(v_1^2 - v_2^2)\right]^{\frac{1}{2}} \quad (m.s^{-1}) \tag{1.44}$$

Δh - total head at a channel section (m)

v_1, v_2 - mean velocity in the upper and lower profile $(m.s^{-1})$

L - length of the channel section (m)

R - hydraulic radius ($= \frac{A}{0}$) (m)

A - area of the cross section (m^2)

0 - wetted perimeter (m)

k - reduction coefficient (≤ 1)

n - coefficient of channel roughness (smooth 0,01, variable sections 0,05)

Under conditions of a stable profile, uniform slope and roughness in a channel without barriers, the equation (1.43) can be simplified as follows

$$v_s = \frac{1}{n} \,.\, R^{\frac{2}{3}} \,.\, I^{\frac{1}{2}} \qquad (m.2^{-1}) \tag{1.45}$$

$I = \frac{\Delta h}{L}$ - slope of the channel

The roughness coefficient depends on geomorphological conditions: the riverbed material, the unevenness of its surface, the character of the profile changes, the barriers in the riverbed, the riverside vegetation, the meandering and sediment transport (Tab. 1.8). The total roughness coefficient can be assessed on the basis of the supplemented formula of Cowan (1957)

$$n = m \,.\, s \,.\, \sum_{k=1}^{4} n_k \tag{1.46}$$

TABLE 1.8

Coefficient of	Characteristics	Value	Coefficient of	Characteristics	Value
n_1 material roughness	earth	0,020	n_4 barriers	negligible	0,000-0,010
	rock			small	0,010-0,015
	fine gravel			medium	0,020-0,030
	coarse gravel			high	0,040-0,060
n_2 bed roughness	smooth, plain	0,000	n_5 vegetation canopy	low	0,005-0,010
	small ripples	0,005		medium	0,010-0,025
	medium ripples	0,010		high	0,025,0,050
	dunes	0,020		high and dense	0,050,0,100
n_3 cross section changes	gentle	0,000	m meandering	low	1,000
	occasional	0,005		medium	1,150
	frequent	0,010 -0,015		high	1,300
			s sediment transport	low	1,000
				high	1,500
				muddy discharge	2 - 100

Partial coefficient for estimation of the roughness coefficient for various boundaries according to Cowan (1957) supplemented by the coefficient of the sediment transport impact.

1.4 INTERRELATIONS OF SURFACE WATER AND GROUNDWATER RUNOFF

Runoff is a hydrologic process of rainfall distribution by the Earth's surface, which takes place in the system of the litho- and hydrosphere. This system consists of natural (morphological, geological, soil, vegetative) and anthropogenetic elements (urban, rural and other constructions, dykes, reservoirs, drainage and sewerage networks etc.). The output of this system depends on the input, which is characterized by

(a) meteorological data, especially rainfall distribution

(b) climatological data, or the supply of solar energy,

and on the actual state of this system, which depends on its previous function (degree of saturation) and anthropogenetic factors (water management activities).

Under natural undisturbed conditions, the surface outflow can be characterized by meteorological and climatological factors (Tab. 1.9).

Surface runoff equals precipitation minus interception, depression and detention storage, changing into infiltration and evaporation. The ratio of the sur-

TABLE 1.9

Category	Area (climate)	Sub-category	Characteristics of discharge occurrence
A. Rivers whose flow depends mainly on rainfall	arid and semiarid	1.	discharges only in rainfall periods
		2.	high discharges in winter, extremely low in summer
	tropical and subtropical	3.	high discharges during summer
	humid	4.	high discharges out of the summer season
B. Rivers whose flow depends mainly on snowmelt and glacier run-off	cool humid	1.	peak discharges especially in spring, influenced by rainfall
	hilly, northern	2.	high discharges influenced by rainfall
	high mountains	3.	high discharges from snowmelt in summer
	permafrost	4.	temporary streams downstream of glaciers

Categorization of rivers according to meteorological and climatological factors.

TABLE 1.10

Type of area			Slope	
		flat 1%	average 1-5%	steep 5%
Residential Appartment dwelling	closed blocks paved courts	0.70	0.80	0.90
	closed blocks with yards	0.60	0.70	0.80
	open blocks	0.50	0.60	0.70
	detached multi-units	0.40	0.50	0.60
Single-family houses with gardens	attached	0.30	0.40	0.50
	detached	0.20	0.30	0.40
Industrial	old type densely covered	0.60	0.90	-
	modern with laws	0.40	0.50	-
Parks, cemeteries, playgrounds		0.10	0.20	0.30
Streets, drives, walks, roofs		0.70	0.80	0.95
Railway areas		0.20	0.30	0.40
Unimproved areas		0.10	0.20	0.30
Grassland, fields	sandy soil	0.05	0.10	0.15
	heavy soil	0.17	0.22	0.35
Forests		0.00	0.05	0.10

Values of runoff coefficient in relation to the type of the drainage area.

face runoff and total losses (recharge and evaporation) does not change when the state of the elements and the energy supply into the system remains constant. For practical reasons this ratio is considered as stable even in the case of a single rainfall. Such hypothesis leads to the following simplified equations for each of the elements i:

$$Q_{si} = P_{gi} - (I_i + D_i) = P_i - (R_{gi} + E_i) \qquad (m^3) \qquad (1.47)$$

$$\frac{Q_{si}}{R_{gi}+E_i} = k \qquad (1.48)$$

and for the total area

$$Q_s = 1000. \sum_{i=1}^{n} C_i \cdot P_i \cdot A_i \qquad (m^3) \qquad (1.49)$$

Q_s - surface outflow (m^3)

G_g - groundwater/soil recharge (m^3)

P_i - precipitation in element i (mm)

C_i - runnoff coefficient of element

A_i - area of element (km^2)

This simplification neglects the time distribution of the input data and the changing state of the runoff system. The actual runoff coefficient is not stable, it is not only a function of the drainage area roughness r (which changes e.g. with the season), its shape and slope i, geology g but also a function of the soil state

$$C = \Phi (r , i , g , s_f) \qquad (1.50)$$

Factors r, i, and g are relatively stable and are almost independent of meteorological conditions. Factor s_f depends on frost and saturation of soil. It determines the actual runoff in the specific hydrologic situation.

The total annual runoff can be determined on the basis of the climatological input data. Data on the left side of the simplified equation of the hydrologic balance i.e. precipitation P and surface runoff Q_s

$$P - Q_s = G_g + E \qquad (mm, m^3 \text{ per year}) \qquad (1.51)$$

can be measured quite easily and precisely. They are in most cases measured in the long term and systematically, and also analyzed statistically. Data on evaporation E and groundwater recharge G_g, on the right side of the equation, are difficult to measure and, therefore, not systematically followed up. Groundwater recharge and evaporation can be expressed as a function of the left side of the equation 1.51

$$E = f_1 (P - Q_s) \qquad (mm,m^3) \qquad (1.52)$$

$$G_g = f_2 (P - Q_s) \qquad (mm,m^3) \qquad (1.53)$$

The maximum possible evaporation for the measured long-term difference of rainfall and the surface runoff is

$$E_m = P - Q_s \qquad (mm,m^3) \qquad (1.54)$$

In this case $G_g = 0$, the groundwater is without recharge.

This phenomenon occurs in desert catchment areas, where all the infiltrated water evaporates. This can be graphically illustrated by a straight line with an angle of 45° (Fig. 1.11).

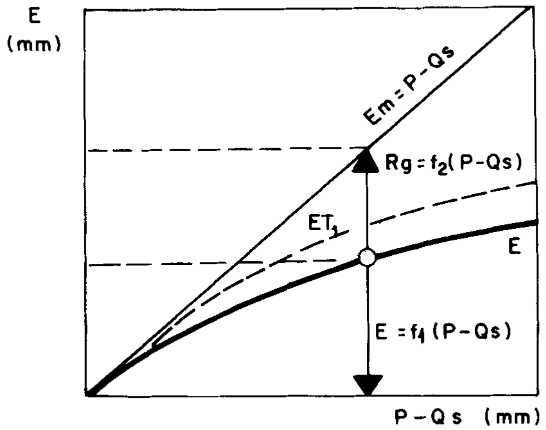


Fig. 1.11. Regional characteristics of the runoff: E - evaporation, ET_1 - evaporativity, P - rainfall total, G_g - groundwater recharge, Q_s - surface water runoff.

The second limiting stage could theoretically be reached when the difference of rainfall and surface runoff recharges the groundwater without any evaporation. This case is graphically illustrated by the horizontal axis. The practical values of the function f_1 migrate between these two limiting stages. They are also limited by the ET_1 value of potential evaporation corresponding to the supply of solar energy in the area. Curves f_1 and f_2 express the average influence of the input data of the relevant runoff system and can, therefore, be used as regional characteristics for the assessment of the groundwater runoff and evaporation.

The system of the rainfall/runoff process can be modelled on a physical or mathematical basis. The basis of the mathematical Tank Model assembled by SUGAWARA (1974) is hydraulic. This model represents the catchment area by a set of tanks, arranged vertically in a row. The number of tanks, their grouping and configuration depend on the catchment characteristics. Experience shows that the following two basic systems are suitable for any practical case

(a) four tanks arranged vertically for humid areas,

(b) several rows of four tanks arranged vertically for semi-arid and arid areas.

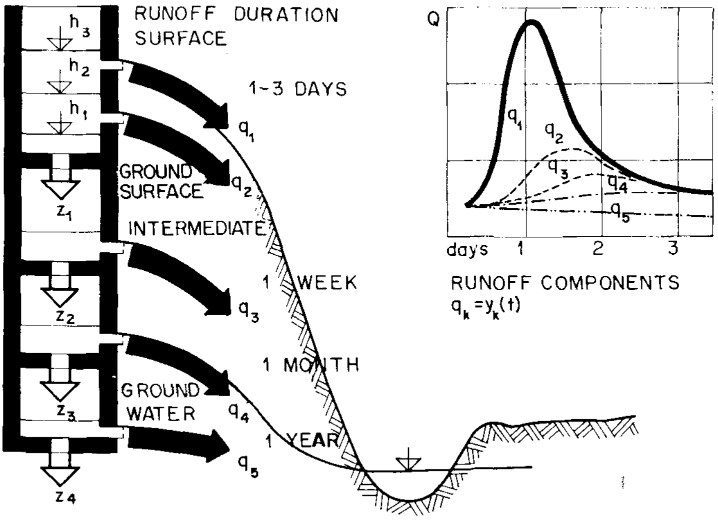


Fig. 1.12. Separation of the runoff components, its course and physical principles of the mathematical model of the runoff process according to SUGAWARA (1974): q_1 - surface runoff, q_2 - ground surface, q_3 - intermediate outflow (above the groundwater table), q_4 - groundwater runoff with short-term and q_5 - with long-term delay before penetration into the river, z_1 - infiltration, z_2, z_3 - percolation into groundwater, z_4 - deep percolation. Water tables: h_1 - at low, h_2 - at medium, h_3 - at high rainfall.

Tanks are equipped with side outlets and bottom outlets. The outflow from the side outlets simulates the following components of the surface runoff (Fig. 1.12):

- the top tank the surface and ground surface runoff, reaching the channel in one to three days,
- the second tank the intermediate runoff, reaching the channel in a week's time.
- the third and fourth tank the groundwater runoff, reaching the channel in one month, or in a year's time.

The top tank generally has two side outlets, while the other tanks are equipped with one side outlet only. The bottom outlets of all tanks simulate the infiltration or, in the case of the fourth tank, the deep percolation. The outflow from outlets is simply expressed by a linear or square relation on the storage amount:

$$q_k = \alpha_k \cdot X_k = f_k(t) \qquad (m^3.s^{-1}) \qquad (1.55)$$

q_k - outflow from the outlet $(m^3.s^{-1})$

X_k - storage amount (m^3)

α_k - outlet coefficient (s^{-1})

The following simple square relation is used whenever the linear relation does not generate satisfactory results:

$$q_k = \alpha_k \cdot X_k^2 = F_k(t) \qquad (m^3.s^{-1}) \qquad (1.56)$$

The nonlinear course of the output data is a consequence of a summarization of the partial results:

$$Q = \sum_{k=1}^{k} q_k = \sum_{k=1}^{k} \alpha_k \cdot X_k \qquad (m^3.s^{-1}) \qquad (1.57)$$

Low rainfall, not filling up the top tank up to the first side outlet, does not produce any runoff. Short floods with a rapid increase in discharge can be modelled more precisely by using more side outlets in the top tank: The saturation of the soil layer can be expressed by a restriction of the bottom outflow. Evaporation produces the decrease in the storage amount in the first tank. The number of tanks, their equipment and arrangement represent the behaviour of the catchment area. This arrangement has to be calibrated to achieve the desired relation of the input and collected output data.

Catchment areas in semi-arid and arid regions have to be divided into zones and represented by several rows of vertically grouped tanks. Evaporation in the period without precipitation has to be modelled by a space without an outlet in the top tank. Deformations of the discharge by river channels can be modelled by similar tanks.

The Tank Model is able to extend runoff data series or to complete missing data by simulation. The most important act to achieve the required accuracy is the assessment of mean precipitation: areal fluctuations in precipitation are high, both in small and large catchment area.

Purely mathematical models of the rainfall/runoff process can be derived from equations describing the physical substance of the hydrological process (Fig. 1.13, 1.14) and the hydrologic balance. Input data for such a model include rainfall records, catchment and meteorological characteristics, while the output data cover surface and groundwater discharges, i.e. entry data for the erosion process (Fig. 1.22).

Mathematical simulation models require detailed information about numerous coefficients (Fig. 1.14), which is very seldom available in the precise form required. These models are deterministic, but a stochastic approach has to be adapted to make them function reliably.

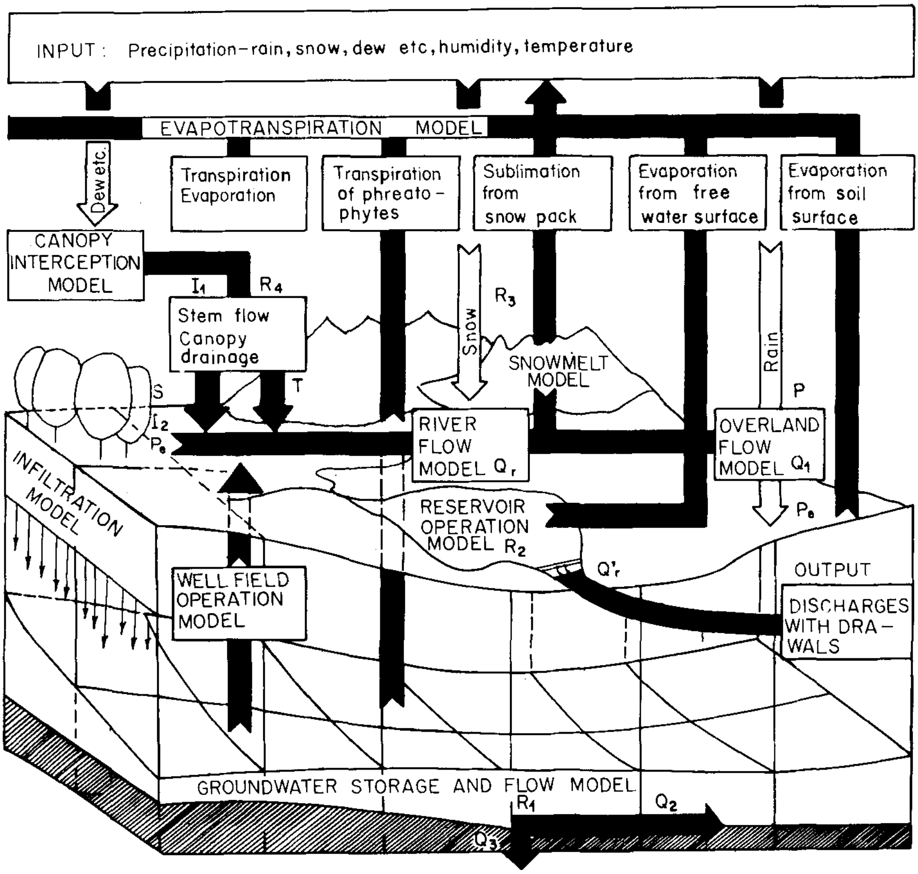


Fig. 1.13. Schematic representation of the model components of the hydrological cycle in rectangular grid: evaporation and canopy interception model, snowmelt (layered) and oveland flow model (two dimensional), river flow and reservoir operation model, root zone infiltration and recharge model (layered, one dimensional unsaturated flow for each grid element), groundwater model (layered, storage and saturated flow).

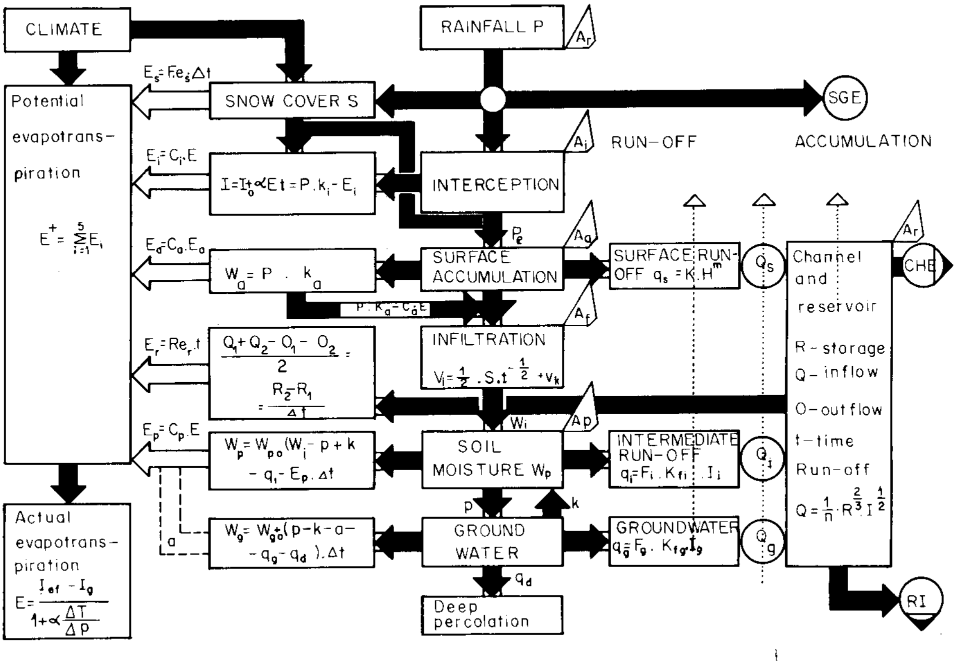


Fig. 1.14. Mathematical model of the water cycle. Explanation of main equations is given in text. Outputs of the run-off process form inputs of the erosion process model (Fig. 1.24). SGE - sheet/gully erosion, CHE - channel erosion, RI - recharge/infiltration, A_r, A_i, A_a, A_f, A_p - anthropogenetic factors.

1.5 GROUNDWATER LEVEL REGULATION, SOIL MOISTURE AND SOIL STRUCTURE FORMATION

Groundwater flow into streams/reservoirs, or vice-versa, depending on the relative altitude of the relevant water levels. In this case of the higher surface water level, infiltration i.e. water supply to aquifers occurs. Under conditions of the lower surface water level groundwater which drains into a stream forms its base flow.

Changes in this regime of infiltration or drainage can occur with time as stream or groundwater level shift (Fig. 1.15). The low rate of groundwater flow causes a shift in time and a reduction in the values of groundwater table fluctuation. The filtration results in a gradual improvement of the quality of the infiltrated water.

The period of progression of the groundwater into the stream channel/reservoir depends on the coefficient of hydraulic conductivity, effective porosity, distance and head

$$t > \frac{n_e \cdot L^2}{k_f \cdot h} \qquad \text{(days)} \qquad (1.58)$$

t - period of progression (days)

n_e - effective porosity (%)

L - shortest distance from the stream channel (m)

k_f - coefficient of hydraulic conductivity (m per day)

h - head (m)

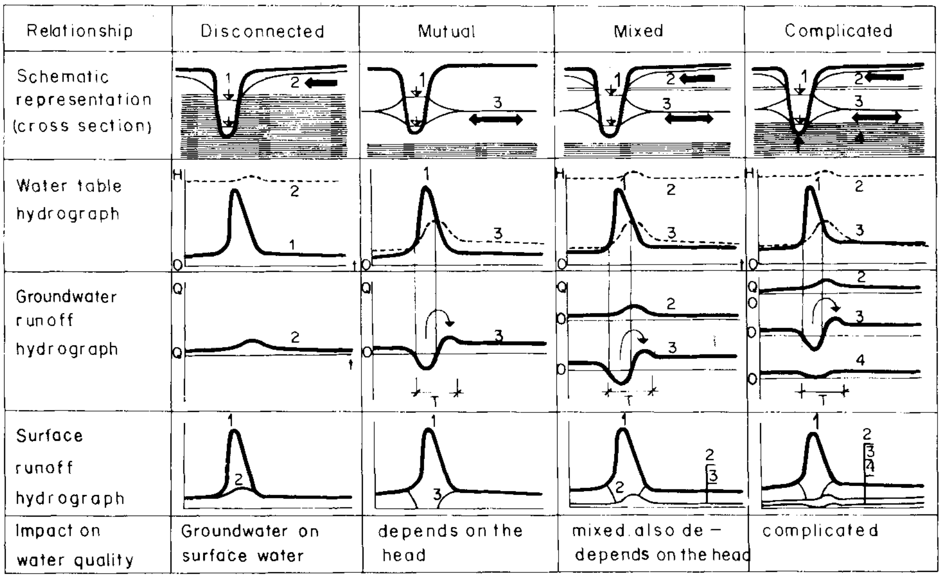


Fig. 1.15. Interrelationships of runoff, water tables and quality of the groundwater. Water tables: 1 - water course, 2 - independent aquifer, 3 - aquifer dependent on the stream, 4 - confined aquifer.

The direction of the flow between the surface and the groundwater influences the water quality: intruding water changes the water quality of the effluent. Both processes, and especially infiltration, the process of surface water penetration into the groundwater can be slowed down by clogging, i.e. the blocking of pores and cracks by suspended matter. During drainage, a reverse process occurs, suffusion, which gradually speeds up the groundwater movement.

Soil moisture, which is of basic importance for the water supply of plants, is less supplemented by groundwater than by precipitation. Gravitational potential is less important for its exploitation by plants than the capillary and osmotic forces, whose function can be measured as the total soil suction. This total soil suction consists of the matric suction, numerically equal to the capillary pressure, and the solute suction, numerically equal to the osmotic pressure. These depend on the soil texture and structure, on the interconnection of pores and cracks, on the chemical properties and temperature of the soil and water, but especially on the moisture content expressed by the ratio

$$w = \frac{V_w}{V_o} \cdot 100 \qquad (\%) \qquad (1.59)$$

w - moisture content (volumetric)

V_w - total volume of water in pores and voids (m^3)

V_o - total volume of the rock or soil (m^3)

The relationship of the moisture content and the total soil suction has to be expressed using the natural logarithm

$$pF = \ln \text{ of the total soil suction} = 3 + 981 \text{ Pa (mm of water)} \qquad (1.60)$$

(Tab. 1.11)

TABLE 1.11

Hydrolimits	Suction pressure pF	Moisture content %	Definition
Full capacity (saturated)	0	25-60	State in which pores and cracks are filled up with gravitational and other water. Corresponds to the capillary porosity.
Field capacity FC	2.5-3.0	10-40	State in which water is held in the soil after the gravitational water has drained away.
Point of decreased availability	3.1-3.5	4-35	State in which the capillary conductivity was interrupted.
Wilting point WP	4.18	2-30	State in which evapotranspiration exceeds the water input (depends also on the plant species).
Adsorption water capacity	4.8-5.2	1-15	State in which the soil does not contain either gravitational or capillary water.

Categorization of soil state according to its suction pressure and moisture retention from plant cultivation.

The changing soil moisture and the total suction pressure form characteristics intervals. On the boundaries of these intervals the water supply of the plants changes in a way which qualitatively influences the development of the plants. The centre of these intervals has a characteristic value of the total soil suction and moisture, depending on the soil category (Fig. 1.16).

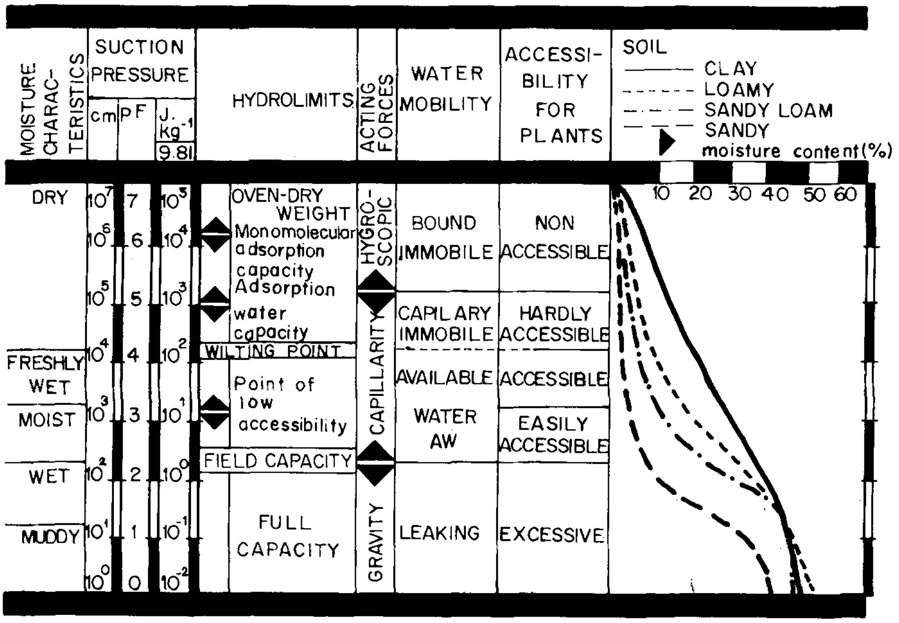


Fig. 1.16. Interrelations of the soil porosity and humidity with the suction pressure, potential and soil moisture retention data according to Kutílek (1979).

The water regime of soils depends on

- soil category, its texture, structure and permeability, characterized by the coefficient of the hydraulic conductivity, or by the course of the total soil suction,
- the position of the soil profile with regard to the groundwater level and the reach of the capillary zone,
- the root system of the plants,
- anthropogenetic factors, including especially changes in the soil texture, structure and moisture during its exploitation,
- climatological characteristics.

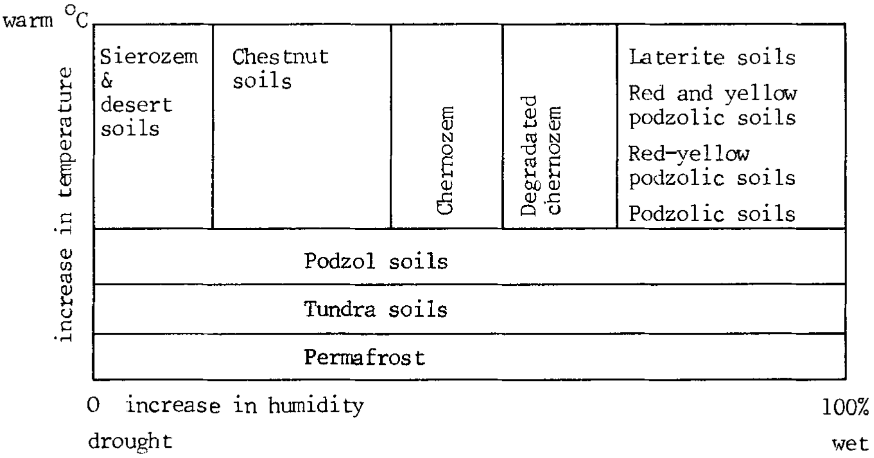
Rode (1956) classifies the water regime of soils on the basis of the annual ratio of the rainfall and evapotranspiration (Tab. 1.12). It is useful to add the swampy regime to this classification, which occurs when the water level permanently penetrates above the soil surface. The gradual development of soil is interrelated with the water regime and the action of climatological factors (Tab. 1.13).

TABLE 1.12

Regime	Characteristics	Ratio of annual precipitation and evaporation total
1. Permafrost	The soil water is permanently frozen	>1
2. Flushing	The soil is completely wetted several times a year	>1
3. Intermittent	Soil is not flushed every year	$\doteq 1$
4. Unflushed	Infiltration does not recharge groundwater reserves, the soil profile being only partially wetted	<1
5. Evaporative	Soil profile moisture is supplemented from groundwater	<1
6. Irrigation	Soil regularly wetted by irrigation	<1 $\frac{P + I}{E} \geq 1$
7. Marshy	Capillary rise permanently reaches the soil surface, oversaturated by groundwater	-

Categorization of soil water regimes according to the ratio $r = \frac{P}{E}$ of the average annual precipitation total P and annual mean evaporation total E according to Rode (1956). I - annual irrigation rate

TABLE 1.13



The impact of moisture and temperature on the formation of the soil profile according to BLUMENSTOCK and THORNWAITE (1941)

1.6 CLIMATOLOGICAL FUNCTIONS OF WATER

The water cycle is one of the basic processes which regulate the climate. The characteristic properties of water which influence the climatic conditions depend on its molecular structure. The climatological regulating functions of water depend on a number of unique physical characteristics: the high value of the specific heat, the high values of the latent heat of fusion and evaporation, the high coefficient of heat reflectance, the extremely low heat conductivity, and the anomalous density decrease below $4^{o}C$, differing from the fusion point (Tab. 1.14).

TABLE 1.14

Climatic impact of water	Fundamental cause
1. Accumulation and slow transfer of heat	High specific heat $I_s = 4.19 \ . \ 10^3$ J.kg.K^{-1} (at $20^{o}C$) 4-5 times exceeding that of air and rocks High latent heat of fusion $I_f = 3.55 \ . \ 10^5$ J.kg^{-1} (at $0^{o}C$) High latent heat of evaporation $I_e = 2.26 \ .10^6$ J.kg^{-1} (at $100^{o}C$)
2. Limitation of energy input from universe	High coefficient of heat reflectance $k_{hr} = 0.582$ W.m^{-1}.K^{-1}
3. Limitation of heat radiation	
4. Regulation of energy transfer between the lithosphere, hydrosphere and atmosphere	Low heat conductivity Fresh water ($20^{o}C$) 0.557 W.m^{-1}.K^{-1} Sea water ($18^{o}C$) 0.561 W.m^{-1}.K^{-1} Ice 1.173 W.m^{-1}.K^{-1} Snow Highest density at $4^{o}C$
5. Influence on temperature inversion	High specific heat and low heat conductivity
6. Influence on microclimate	High specific heat and high latent heat of fusion and evaporation. Irregularity of the highest density occurrence (at $4^{o}C$).

Climatic impact of water and its fundamental causes as a function of its exceptional molecular characteristics.

The coefficient of heat reflectance occurs in the reflectance of solar energy and in the heat transmissivity of the atmosphere. The Earth reflects 34% of the solar energy on average. Two thirds of this value, i.e. 23% of the total, is reflected by clouds, or accumulated vapour in the atmosphere. The increase in the water vapour pressure leads to a decrease in the heat transmissivity of the atmosphere. The difference between the average values of the heat transmissivity of the equatorial and the polar zone, with lower humidity, exceeds 13%.

Water vapour is the main cause of the hot-house effect of the atmosphere. It intercepts infrared radiation in the sphere of the wave length $4,4\mu < \lambda < 8,5\mu$ and $11\mu < \lambda < 80\mu$. The hot-house effect of the atmosphere consists in the input of radiation from the Universe to the Earth with small losses during daytime and in the output of the heat from the lithosphere to the Universe with high losses during nighttime. The atmosphere retains and reflects back to the Earth's surface the infrared radiation, forming 66 - 75 % of the radiated energy. Carbon dioxide CO_2 and ozone O_3 also take part in this effect, but in a very limited part of the spectrum, thus having far less influence.

Large water formations such as seas, icebergs, marshes and dambos, reservoirs and water courses act as thermal regulators. Evaporation and transpiration need a large quantity of thermal energy. This energy is released during condensation and accepted by the surrounding air, thus causing an equalizing of thermic differences. Due to the high specific heat of water, the process of its warming and cooling is slower than the process of the warming and cooling of soil. The differences in winter and summer temperatures, or day and night temperatures are, therefore, smaller in the vicinity of water formations. The local climate is influenced by the thermal inertia manifested by a slow transfer of heat to the environment. The degree of this influence depends on the volume of the functioning water mass and also acts over considerable distances, demonstrated by the continental nature of the climate.

Water reservoirs, marshes, dambos, snow and ice, also functioning as runoff regulators, thus propagate this thermal influence along water courses. The regulating function of marshes and dambos qualitatively differs from the influence of reservoirs. Marshes and dambos, containing 80 to 97% of water, bond on organic matter. The content of water in marshes cannot be decreased below 70% by drainage. Evapotranspiration from a marshy area forms 30 - 50% of evaporation from an open water surface, not taking away as much energy.

Snow and ice, thermal and runoff regulators, basically occur in two forms

(a) temporary snow and ice, thawing in the warm season,

(b) permanent snow and ice, especially Antarctic and Greenland icebergs, sea icebergs and mountain glaciers, permanent snow , permafrost.

Permanent or perennial ice and snow occur depending on the climate, altitude, latitude and morphology. If the climate is very rigorous, a layer of frozen ground, permafrost, may be formed, persisting from year to year. Thick snowpack tends to impede the formation of permafrost.

In areas where more snow and ice accumulate than actually melt and adjoining areas of lower altitude where the wastage of frozen water exceeds its accumulation, a slow movement of snow and ice mass from the upper area to the lower one may occur when the thickness of the glacier exceeds 15 to 30 m. The speed of this movement depends on the glacier thickness, its temperature and the slope of the ground. In this way, frozen water is transferred to areas with more favourable conditions for its participation in the hydrologic cycle (see also Tab. 1.1).

The hydrologic and microclimatic influence of snow and glaciers also functions by means of supplied rivers. Glaciers thaw more slowly than snow, thus offering high river discharges before the beginning of the summer season, when irrigation requirements may be expected to be highest (Tab. 1.9, Tab. 1.15).

TABLE 1.15

Season	Albedo	Snowpack thickness	Diurnal fluctuation	Runoff	Runoff after rainfall	Longterm storage impact
Spring	High	Highest	Slight	Moderate increasing	Subdued delayed	Dry & warm years = = Increase in total runoff
Summer	Moderate	Moderate	High	High	Slight delay	
Autumn	Low	Low	Moderate	Moderate	No delay	Wet & cool years = = Decrease in total runoff
Winter	Very high	Moderate	Nil	Slight	Stored	

Seasonal change in glacier-runoff characteristics according to MEIER (1964).

The regulating functions of water are also a consequence of its low heat conductivity. Heat flow is the most effective form of heat propagation in liquids. Under conditions of restricted flow below the ice cover, the heat losses of the water formation are low. The snowpack has an extremely low value of the heat conductivity. The insulating capability of water therefore appears in the winter season, remarkably enough, when this is urgently required for the maintenance of the temperature of the soil surface.

The low heat conductivity of the snowpack means that the heat output to the atmosphere does not exceed the heat input from the deeper formations. The soil temperature below the snowpack can be up to 15°C higher in comparison to the temperature of soils without the snowpack. The insulating effect of the snowpack does not only depend on its thickness, but also on the snow quality.

The snowpack has a cooling effect on the adjoining, warmer layer of atmosphere, thus causing ground inversion of temperature, especially in the spring season, and an associated increase in relative humidity. Such inversions above the open water surface are not as intensive and occur less often.

The air humidity can be characterized by the water vapour pressure. Its value oscillates between 0 and 40 milibars (0 - 4000 Pa) (Fig. 1.17). This pressure decreases with altitude: at 1750 m a.s.l. at one half and at 5000 m at one fifth of its original values at sea level. The content of water vapour in the atmosphere is on average 0.2% in the polar zone and 2.5% in rainy, tropical regions. It may, locally, exceed 4%.

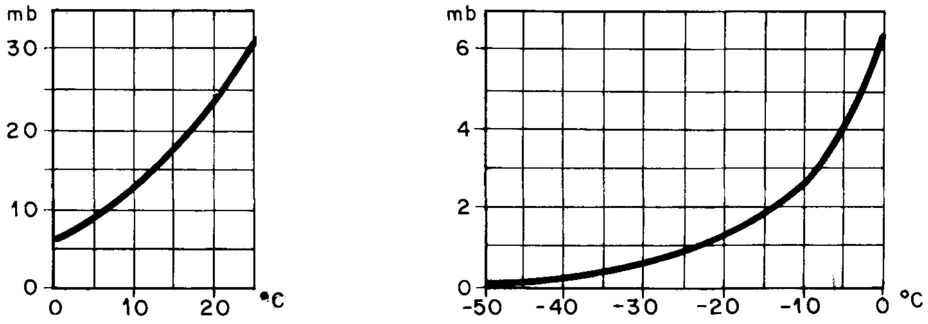


Fig. 1.17. Relation between temperature and saturation vapour pressure above and below 0°C.

The maximum amount of water vapour in the air depends directly on the temperature. The state in which water vapour can exist in equilibrium with a plane surface of liquid water at the same temperature is saturation. The relative humidity is the ratio of the actual vapour pressure to the saturation vapour pressure at the same temperature

$$U = \frac{e_w}{E_s} \cdot 100 \qquad (\%) \qquad (1.61)$$

U - relative humidity (%)

e_w - actual water vapour pressure (mb, 1 mb = 100 Pa)

E_s - saturation water vapour pressure (mb)

The air humidity is continuously changed by evapotranspiration, by the air flow and by temperature changes. The dew-point temperature is the temperature to which the air must be cooled at constant pressure to reach saturation with a given water content. Any further decrease in temperature thus causes condensation, occurring in different forms of the horizontal and vertical precipitation (Fig. 1.7) depending on the altitude, temperature and other conditions of the environment, e.g. morphology and temperature of the affected surface.

The daily minimum of the relative humidity occurs simultaneously with the daily maximum temperature at constant water vapour pressure. Changes in the water vapour content in the atmosphere cause changes in the relative amounts of the four principal gases (nitrogen N, oxygen O, hydrogen H and carbon dioxide CO_2) and minute traces of other gases, thus influencing its physical, chemical and biological characteristics.

The term drought (lack of rainfall), which negatively affects both the life of man and the development of plants, basically has two different connotations:

(a) climatic drought, caused by lack of precipitation or by its inadequate or irregular occurrence - particularly in the growing season or caused by high temperature, occassioning extreme evapotranspiration,

(b) local drought, which may also occur in areas with heavy rainfall during the growing season, occurring in regions with permeable soils, where excessive runoff impedes a sufficient water supply of plants.

The degree of the climatic drought depends especially on the average total rainfall or on its ratio to the mean annual temperature (Tab. 1.16).

TABLE 1.16

Climate	Mean annual precipitation total (mm)	Rainfall factor r(mm/°C.year)
Wet	> 500	> 60
Semi-arid	400- 500	40-60
Dry (steppe and semi-desert)	200- 400	< 40
Extremely dry (arid)	< 200	$\ll 40$

Categorization of climate according to the mean annual precipitation total and in relation to the rainfall factor.

Climate is the average cycle of weather characteristic for given area. The course of humidity and temperature can be considered as factors which determine the type of climate. The climate can be categorized according to the mean annual total precipitation and in relation to the rainfall factor according to Koeppen (1920) (Tab. 1.16, 1.17)

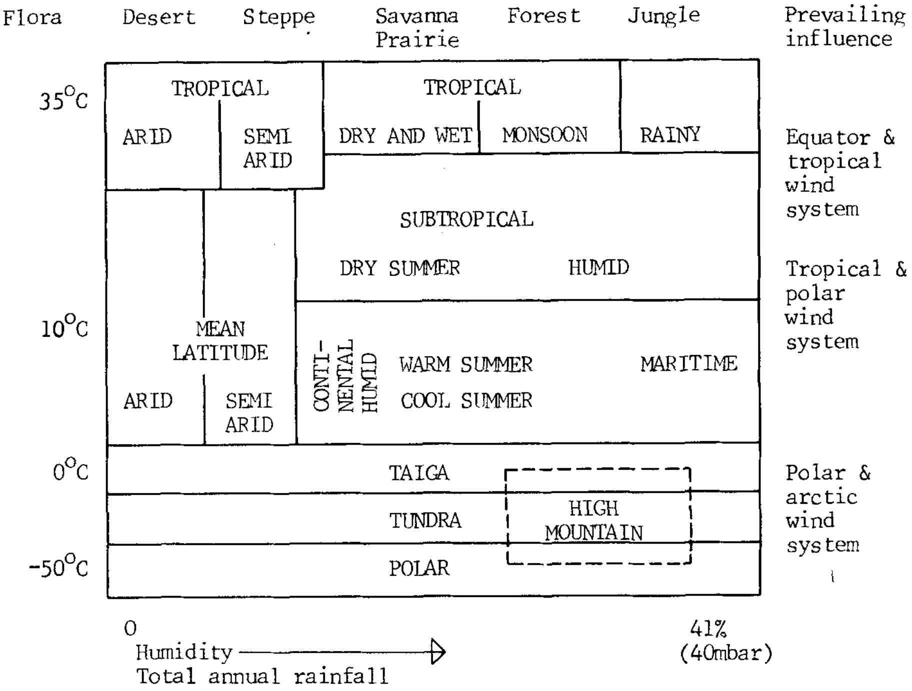
$$r = \frac{P}{T} \qquad (mm/\ ^{\circ}C\ .\ year) \qquad (1.62)$$

r - rainfall factor

P - annual average precipitation total (mm per year)

T - mean annual temperature (^{o}C)

TABLE 1.17



Categorization of climate according to humidity and mean annual temperature, characteristic flora and prevailing factors of influence. Completed according to the structure of Critchfield (1960).

The regulating functions of water vapour, ice, snow and water act in dependance on the supply of solar energy, deciding e.g. on the predominant air circulation system. They are interconnected with numerous factors depending on the location, especially the exposure and altitude of the region. The annual and daily amplitude of temperature, upon which the continentality of the climate depends, increases with the distance from oceans. Regional and local morphology, water bodies, soils and vegetative canopy influence the characteristics of the meso- and microclimate.

1.7 BIOGEOCHEMICAL CYCLE SYSTEM

The mechanical and thermal energy which water acquires during the hydrological cycle, and its chemical energy, is consumed by

(a) the mechanical and chemical destruction of the land surface during rainfall, runoff and weathering, i.e. by hydrolysis, hydration, dehydration, leaching, dissolving , disintegration, absorption etc.,

(b) transportation and corrosion of sediments,

(c) deformation of banks and riverbeds,

(d) overcoming of different resistances during flow and undulation, internal and external friction, water hammer,

(e) biological processes of plant and animal production,

Water is the widespread agent of erosion, entrainment, transportation and deposition of sediment. The system of these complex processes forms the most important abiotic cycle in the environment of the water cycle in the proper sense. It enables the circulation of an organic and organic compounds of the biogeochemical cycle of elements. This cycle is characterized by a continuous, periodic or stochastic

(a) destruction and transfer of compounds and mixtures of different elements,

(b) dissolving, disintegration and integration of gaseous, liquid and solid mineral and organic matter,

(c) synthesis of new compounds, also enabling the initiation of new cycles.

This cycle regenerates nutriments and thus enables the course of the life process. Only a small part of the stock of elements and their compounds, existing in the pedo-, litho-, hydro- and atmosphere, is activated by the biotic and abiotic processes and takes part in the biogeochemical cycle. Erosion is the process of activation. Eroded matter, including bioelements, i.e. elements whose organic compounds form living matter, are entrained, transported and deposed as sediments (Tab. 1.18).

TABLE 1.18

Biogeoelements

Macrobiogenetic		Microbiogenetic				Oligobiogenetic (trace)			
Carbon	C	Calcium	Ca	Silicon	Si	Fluorine	F	Vanadium	V
Oxygen	O	Iron	Fe	Alluminium	Al	Arsenic	As	Chromium	Cr
Hydrogen	H	Magnesium	Mg	Manganese	Mn	Lead	Pb	Molybden	Mo
Nitrogen	N	Sodium	Na	Iodine	J	Titan	Ti	Nickel	Ni
Phosphorus	P	Kalium	K	Zinc	Zn	Cesium	Cs	Cobalt	Co
Sulphur	S	Chlorine	Cl	Copper	Cu	Selenium	Se	Galium	Ga
				Boron	B	Stroncium	Sr	Vismut	Bi
						Rubidium	Ru	etc.	

Categorization of biogeoelements.

Sediments are transferred from the higher altitudes with less favourable topographical and climatological conditions for cultivation to lower elevations of fertile plains and more favourable climate. The boulders, the gravel and partially also the sand move near the bottom as bed load. Fine sand and some organic particles float as suspended load; silt, clay swim as wash load and trees, plants, leaves as floating debris (Fig. 1.4).

During low floods coarse sediments remain in the channel: inundation contains fine particles, whose composition and size is more suitable for the formation of living matter. Suspended and wash load, partly of organic origin, and floating debris regenerate the bioenergetic potential, i.e. the fertility of soils. During superfloods coarse materials may also enter the flood plain and affect the soil fertility in a negative way.

TABLE 1.19

Continent	Catchment area mil.km^2	Mean annual sediment transport		Decrease of the land surface during 1000 years mm
		t.km^{-2}	10^9 t	
Europe	9.3	34.8	0.32	23.2
Asia	26.9	591.0	15.91	394.0
Africa	19.9	27.0	0.54	18.0
North America	20.7	94.6	1.96	63.1
South America	19.4	61.8	1.20	41.2
Australia	5.2	44.4	0.23	29.6
Total/average	101.4	198.8	20.16	132.5

Annual sediment transport and decrease in the land surface according to Holeman (1968).

It is a basic prerequisite for intensive agricultural production that the production process should not be interrupted too often by floods and certainly not at harvest time. Intensive cultivation develops, therefore, in areas where the natural processes of inundation and regeneration of nutriments is already at least partially interrupted.

Wind erosion returns part of the sedimented matter from dry land to upper elevations. The greater part accumulates in seas and reservoirs. It is partially absorbed by the biomass. It also penetrates below the reach of the sun's rays, thus escaping from the biologic cycle.

Sediment represent a load on the geological formations of the Earth's crust and thus influence its balance. In such a way, they are one of the factors which have an impact on volcanic activities. New matter which occurs during volcanic activities on the Earth's surface enters the erosion process and thus closes the biogeochemical cycle.

Natural biogeochemical cycles, when not disturbed by human activities, are with some exceptions almost balanced. They reproduce 90 - 98% of the entering matter, thus maintaining the balance of quantity, structure and concentration, forming a stable basis for the biotic environment. All organisms conform to this state in the long term. The inequality of the biogeochemical cycle in the span of geological time leads to the migration and differentiation of species, to dispersion or concentration of elements and their compounds in different parts of the region and consequently to the formation of different abiotic environments and ecosystems.

An index of element migration, applied by Polynov (1936), is expressed by the ratio of the percentage quantity of the element in the hydrosphere to the percentage of its quantity in the lithosphere. Kovda (1975) proves on the basis of this index that chlorine Cl, sulphur S, iodine I, calcium Ca, natrium Na, magnesium Mg, fluorine F, strontium Sr, zinc Zn, uranium U and molybdenum Mb present high migration, while silicon Si, potassium K, phosphorus P, baryum Ba, manganese Mn, rubidium Rb, copper Cu, nickel Ni, kobalt Co, arsenic As, lithium Li and especially alluminium Al and iron Fe present low migration. Divergences occur locally, especially through changes in the conditions for oxidation or reduction and through the synthesis of integrated compounds.

The migration of biogenic elements is of crucial importance for the creation and growth of living matter. Of the macrobiogenetic elements only phosphorus creates an unbalanced cycle. Anorganic compounds of phosphorus are absorbed by plants. They serve as nutrition for animals, concentrate first of all in their bones and then enter the soil in their excrements as well as after extinction.

In the course of the erosion process 3-4 mil. tons of phosphates are transported annually to the sea. Only one thirtieth of this quantity returns to dry land as a product of fishing, or as guano etc. The natural processes supplement its active stock insufficiently. The circulation of phosphorus is activated by anthropogenetic measures: phosphates are a component of most fertilizers as well as of polluted waters. Blast-furnace slag also contains phosphorus. Nevertheless the stock of phosphorus is gradually diminishing, because this element is absorbed by plants (Fig. 1.18).

During the circulation of sulphur, bacteria, thiobacilli and desulphovibrions act in the soil, forming sulphates by oxidation and sulphides by reduction. The soil is enriched with sulphur by means of rainfall. Organic matter accepts sulphur from the soil and surrenders it again during decomposition. The erosion process transports part of the sulphur compounds to reservoirs and oceans, from which formations sulphur escapes to the atmosphere in the form of gaseous compounds (Fig. 1.19).

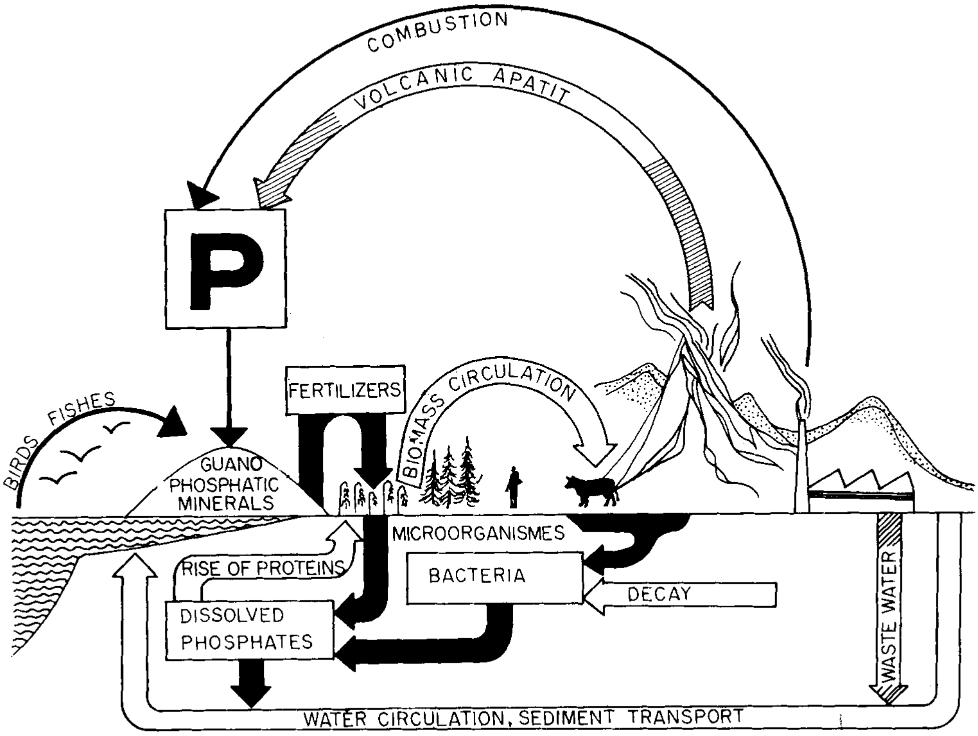


Fig. 1.18. Biogeochemical cycle of phosphorus as a subsystem of the hydrological cycle. The unbalanced cycle of phosphorus is controlled by volcanic and biological processes.

The amount of carbon dioxide CO_2 in the atmosphere is regulated 60% by oceans and 40% by the photosynthesis of plants. These processes also regenerate the stock of oxygen O. Carbonates are transported during erosion to the sea or enter organic matter from the soil. Carbon dioxide escapes from organic matter during respiration, burning, disintegration and from soil and water surfaces directly back to the atmosphere (Fig. 1.20).

The circulation of oxygen O, hydrogen H and nitrogen N, whose main stock is the atmosphere, is balanced. Forests, covering 9% of the Earth's surface, produce 48 mld. tons of oxygen annually, i.e. 47% of the total production. The rest is almost completely the share of oceans: the production of rivers and reservoirs is important from the point of view of meso- and microclimate only. Oxygen and hydrogen compounds, such as salts, oxides etc., forming soils, rocks and organic matter take part in the whole biogeochemical cycle.

The nitrogen N is a component of proteins, thus entering the waste from digestion, extinction and industrial processing. Saprogenic bacteria, acting during decay and ammonification, help to create ammonia NH_3. Under the action of nitrificating bacteria the ammonia changes into nitrites and than nitrates. Nitrates enter plant matter directly and, partly by denitrificating bacteria, are decomposed. Free nitrogen, the result of this process, is accepted by the atmosphere (Fig. 1.21). Goldschmidt (1954) states that 0.75 $g.m^{-2}$ of nitrogen enters into the circulation in zones with a mild climate and almost 3 $g.m^{-2}$ in zones with a humid, tropical climate.

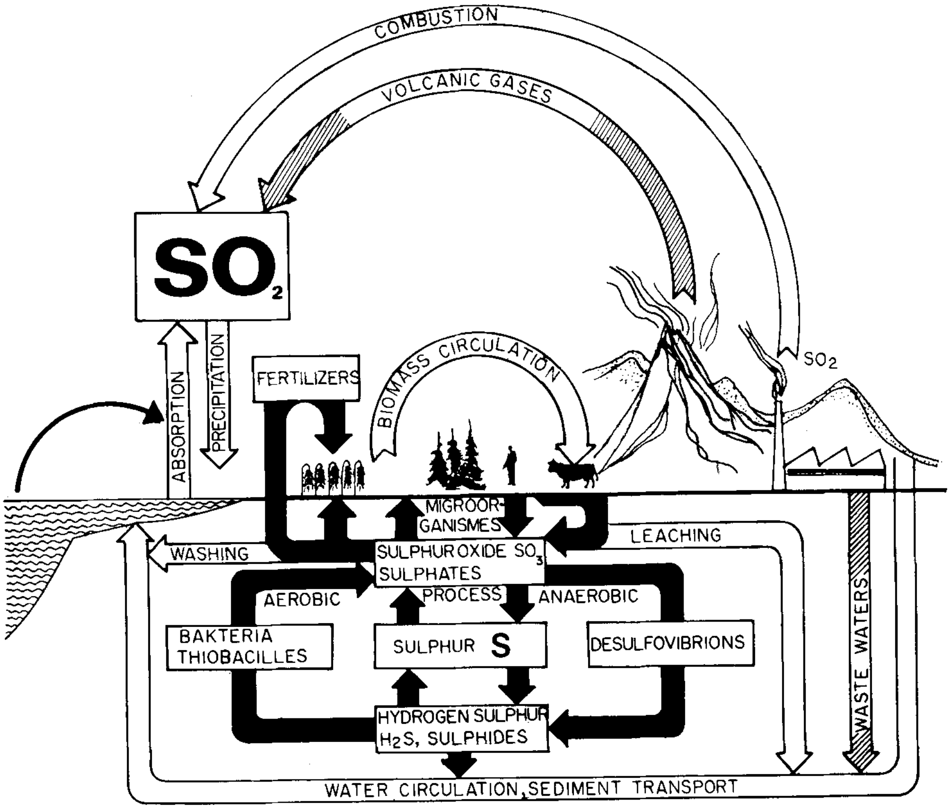


Fig. 1.19. Biogeochemical cycle of sulphur as a subsystem of the hydrological cycle. The cycle of sulphur is activated especially by microorganisms and volcanic activities.

1.7.1 Water Erosion as a Process Evoked by the Water Cycle

The chemical and mechanical energy of wind, ice, rainfall, runoff and sediments, acting during the hydrologic cycle in the system of the pedo- and lithosphere, results in the erosion, disintegration and wearing away of the land (Tab. 1.19). The interplay of acting forces depending on hydrometeorological factors, geomorphological conditions, namely of

- the density and composition of the vegetative cover,

- material and structure of the surface layer,

and of the duration of this process decides on its course and form. From this point of view, it is possible to distinguish

(a) sheet erosion - detachment and removal of the material from the entire land surface by wind, rainfall and overland flow, which may be either

- selective and change the soil texture together with the content of nutriments in the soil, or
- stratified, when the entire mould layer is washed away,

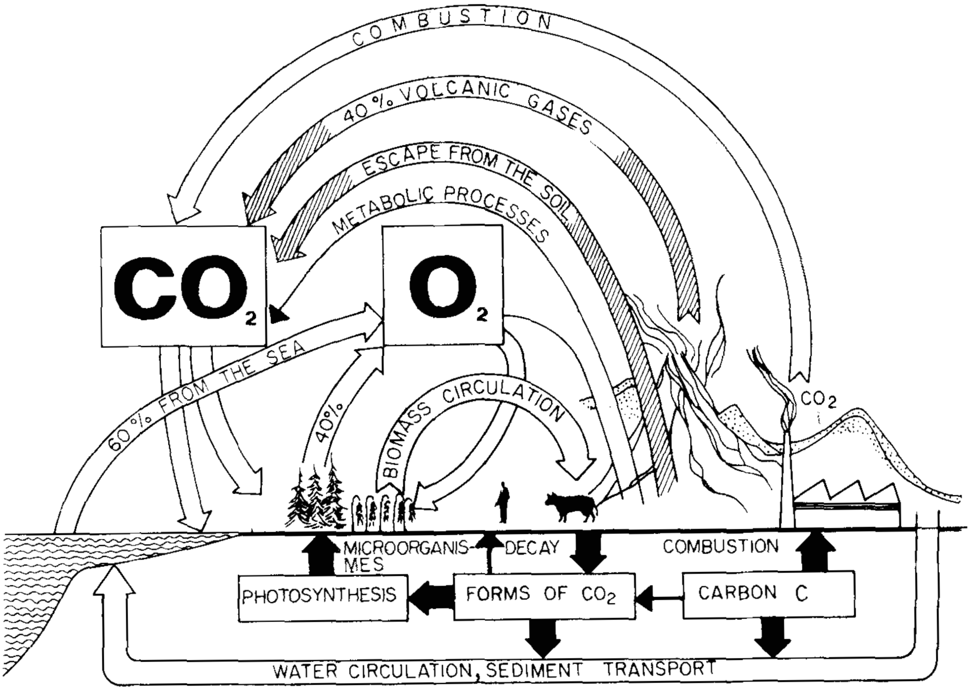


Fig. 1.20. Biogeochemical cycles of carbon and oxygen as subsystems of the hydrological cycle. The content of oxygen in the atmosphere is mainly controlled by oceans and by the photosynthesis of terrestrial flora.

(b) pre-channel erosion, exerted by forces of concentrated pre-channel flow, forming furrows, cuts, wash-outs, pot-holes and gullies,

(c) channel erosion, exerted by forces of concentrated water flow in stream beds, stream banks and flood-plains (Fig. 1.22).

NEAL (1938) connects the potential intensity of sheet erosion

$$G_s = K.\ I^{0.8}\ .\ i^{1.2} \qquad (t.ha^{-1}.min^{-1}) \qquad (1.63)$$

G_s - washed soil $(t.ha^{-1})$

I - gradient of the slope $(^{o})$

i - rainfall intensity (mm.min^{-1})

K - coefficient of local conditions

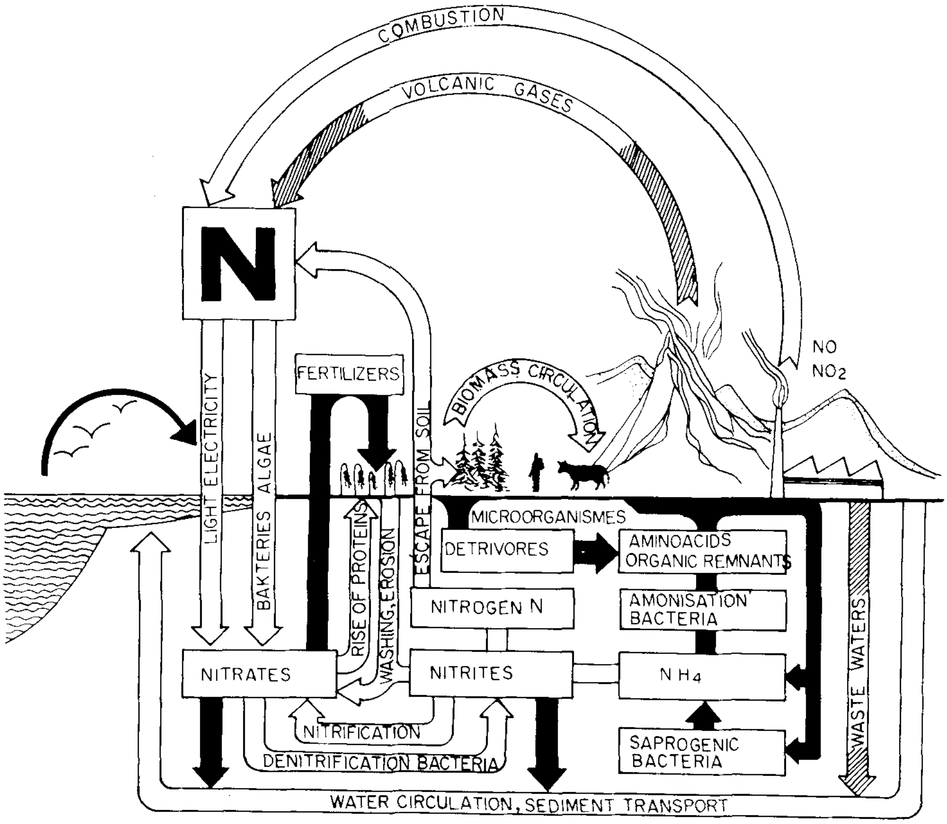


Fig. 1.21. Biogeochemical cycle of nitrogen as a subsystem of the hydrologic cycle. The main reserve of nitrogen is in the atmosphere.

From the chemical point of view, especially nitrogen N, potassium K, calcium Ca, magnesium Mg are washed away, while sulphur S, chlorine Cl and phosphorus P bond in the organic and anorganic soil particles. Washed soil particles may be intercepted by the grasslands, meadows and pastures, when eroded lands do not directly adjoin the stream channel. Plants in such grassland zones accept nitrogen N quite easily, potassium K, calcium Ca and phosphorus P are accepted only partially. Sulphur S, natrium Na and the rest of potassium K as well as of calcium Ca and phosphorus P are carried away by the surface runoff.

The energy of the overland flow increases with its depth and velocity. This flow is accelerated and its discharge grows with the length of course. The qualitative change of the sheet erosion into the pre-channel erosion depends especially on the combination of the soil surface characteristics, rainfall intensity, gradient of the slope and its length. The following formula can be derived for the critical length of the slope, i.e. the length where the less dangerous sheet erosion changes due to the concentrated flow and evoked forces into pre-channel erosion:

$$L = a \,.\, f\,(i,r,t) \qquad (m) \qquad (1.64)$$

L - critical length of the slope (m)

i - gradient of the slope (%)

a - coefficient of soil characteristics (0.5-4,5 : 1 - levelled fields, 1.7-2.3 - meadows)

r - rainfall and climate factor

t - time factor, depending on the duration of rainfalls of high intensity

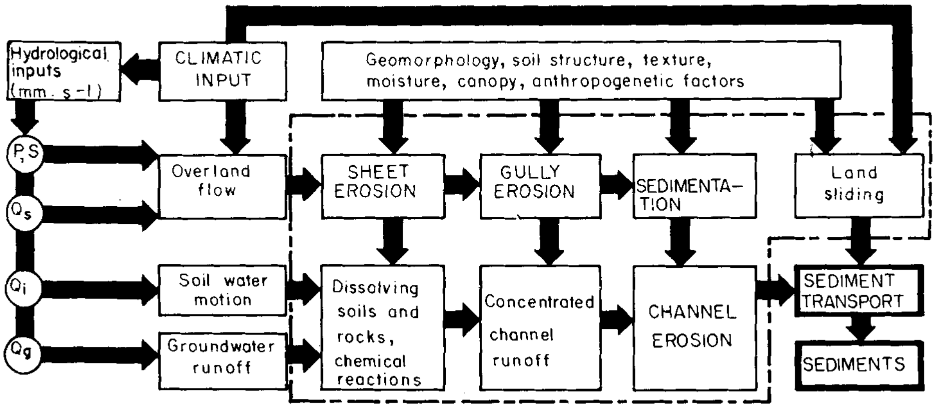


Fig. 1.22. Flow chart diagram of the erosion process. Rainfall and runoff, the output of the runoff system, are inputs of the erosion system. Output is: the sediment flow, settled sediments and the water quality.

Solid soil particles enter the movement of the concentrated overland flow at the moment when its speed achieves the value of the speed limit. This value depends on the weight, size and shape of the relevant particles, which can be expressed by a simple formula

$$v_c = s \,.\, \sqrt{d} \qquad (m.s^{-1}) \qquad (1.65)$$

v_c - speed limit of sediment motion $(m.s^{-1})$

s - coefficient of shape

d - size characteristics (Tab. 1.20)

TABLE 1.20

Material	Grain size (mm)	Speed limit ($m.s^{-1}$)	Material	Grain size (mm)	Speed limit ($m.s^{-1}$)	Coefficient of shape Grain parameter	value (s)
Sand	3	0.108	Cobble rounded	27	0.650	round perimeter	4.46
	6	0.189	Cobble angular, sharp	60	0.975	elliptic long semi-axis	4.43
	20	0.325	Boulders	100	1.400	angular longest edge	3.45

Values of speed limits v_c and shape coefficients s.

Levi (1948) states for nonhomogeneous sediments

$$v_c = 1.4 \cdot g \cdot d_s \cdot \lg \frac{h}{7 d_s} \cdot \left[\frac{d_{max}}{d_{min}}\right]^{\frac{1}{7}} \qquad (1.66)$$

d_{max}, d_{min}, d_s - maximum, minimum and medium size of sediment particles (m)

g - gravitational constant ($m.s^{-2}$)

h - depth of flow (m)

A decrease in the flow rate below the speed limit causes a sedimentation of the soil particles. A stable sediment size is characterized by a stable gradient of the stream channel slope, as can be proved on the basis of Chézy's formula:

$$c \cdot \sqrt{R \cdot i_b} = v_c$$

$$c^2 \cdot R \cdot i_b = s^2 \cdot d$$

$$i_b = \frac{s^2 \cdot d}{c^2 \cdot R} = c_b \cdot \frac{d}{R} \qquad (1.67)$$

i_b - balanced slope gradient (without erosion and sedimentation)

c_b - coefficient of the balanced slope

R - hydraulic radius (m)

Valentini (1893) stated c_b = 0,093 for the slope of the river bed up to 25% and

$$d = \sqrt[3]{\frac{V}{n}}$$

V - volume, n - number of cobbles

The balanced slope gradient for smaller grain sizes follows directly from Chézy's equation:

$$i_b = \frac{v_c}{c^2 \,.\, R} \qquad (1.68)$$

The current velocity is not uniform throughout the cross section and leads to the transport of particles of different size: For practical reasons the beginning of the motion of a considerable quantity of suspended matter is important i.e. the relevant critical discharge, whose occurrence is some 180 days per year under conditions of stable natural river beds.

Integrated processes of physical, chemical and biological character during wind, ice, water and gravity erosion, entrainment, transportation and deposition of sediments, forming the Earth's surface and enabling the circulation of biogeoelements, are greatly affected by human activities. The universal equation for predicting erosion rates can be expressed by the function

$$G_e = f\,(X_1, X_2, X_3, X_4, X_5, X_6, X_7, X_8) \qquad (1.69)$$

G_e - erosion rate (t.ha^{-1})

X_1 - climatic factor (rainfall and wind intensity, ice phenomena)

X_2 - hydrologic factor (concentration of the surface runoff)

X_3 - topographical factor (slope-length, steepness and exposure)

X_4 - soil factor (structure, texture, resulting in soil erodibility)

X_5 - geological factor (stability of the ground)

X_6 - vegetative cover factor (depending also on season)

X_7 - management practices factor (crop management, irrigation)

X_8 - anthropogenetic factor (effect of land development/disturbance and protection measures)

The analytic tractability of the function is complicated by the need to express the space-time characteristics of relevant factors, e.g. the coincidence of the rainfall occurrence and of different characteristics of canopy during relevant growth stages: harvesting, bed preparation and other management practices.

1.7.2 Water Quality as a Product of its Circulation

Water in nature is a multiconstituent compound whose substance is hydrogen dioxide H_2O containing dissolved and dispersed gaseous, liquid and solid cations, anions and nonionic constituents of anorganic and organic origin, as well as aquatic flora and fauna.

The unique chemical characteristics of the hydrogen dioxide H_2O, capillarity with other unique physical characteristics, are a consequence of the excentric position of oxygen with regard to the centroid of the water molecule. The very high dielectric constant of water permits it to decompose molecules of soluble compounds into ions and dissolve complicated matter, also forming soils and rocks.

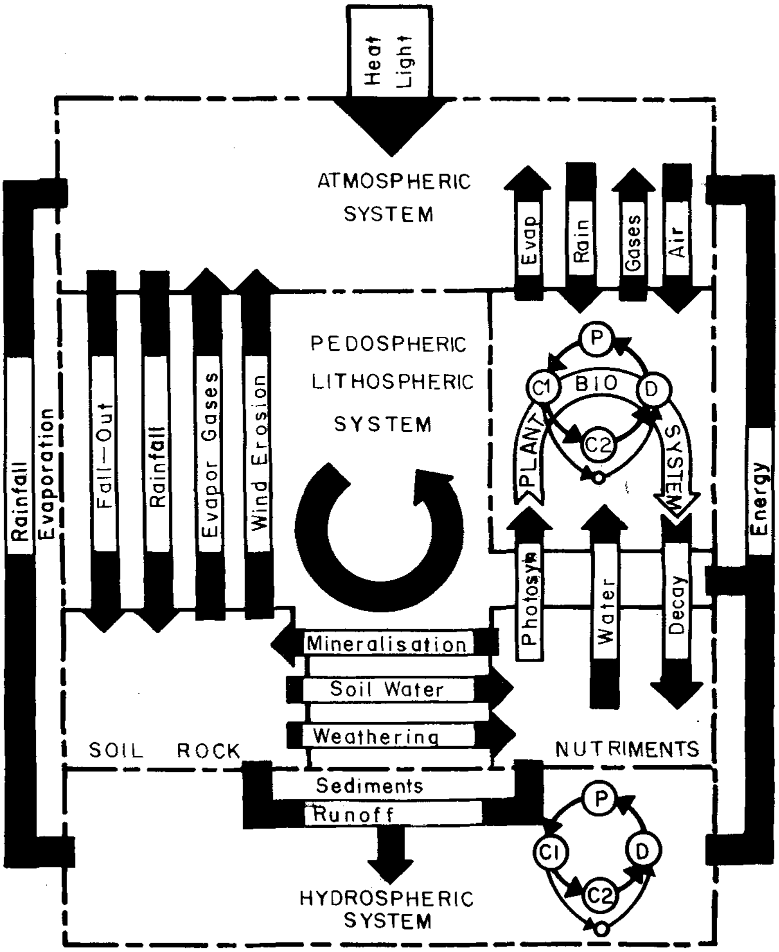


Fig. 1.23. Water and air as basic media of the geo- and biological processes as well as of the matter and energy transfer between continent and oceans.

Water quality consists of different properties, which are important when deciding about its possible utilization. These properties include temperature and colour, taste and odour, caused by the presence or absence of chemical substances and expressed by the alkalinity, acidity, hydrogen-ion concentration, content of carbonates (hardness), oxygen demand, sodium adsorption ratio, as well as by the presence of different ionic and nonionic constituents incl. microorganisms etc. Various physical, chemical and biological characteristics of water are formed by its contact with the environment during its circulation. Water quality changes most during its penetration through the soil and rock environment, but also during its contact with the atmosphere, vegetation, soil and rock during surface runoff and rainfall, further by mixing with water from deeper strata and by mixing and reacting with matter entering the system of atmosphere, hydro-, pedo- and lithosphere in the course of human activities (Fig. 1.23).

The complex of processes which create the characteristics, constituents and properties of water is called mineralization. The course of these processes depends also on the energy input, i.e. on the temperature, pressure and mechanical energy which crush the reacting particles, and on the concentration of acting components which come into mutual contact. Pores, cracks and cavities act as natural regulators of water quality: the rate of relevant processes also depends on the size of the acting surface, and on hydraulic parameters, especially on the flow rate, deciding about the duration of the contact of relevant constituents. Soil also functions as a natural filter, catching a substantial part of polluting matter and disposing it by subsequent biochemical processes (Tab. 1.21).

TABLE 1.21

Mineralization	Characteristics
Meteogenic	Mineralization in the atmosphere
Potamogenic (fluviogenic)	Mineralization in streams and rivers, dependent on the size, shape and geology of the river bed and on the velocity of the flow as well as on water temperature.
Limnogenic	Mineralization in lakes and other standing waters.
Litomorphic	Mineralization in groundwater, dependent on the petrographic and chemical characteristics and on the influence of the underground atmosphere on the infiltrated water and groundwater.
Bathymorphic	Mineralization in deep aquifers determined especially by temperature, pressure, migration, conservation and diagenesis of the groundwater in a way that is composition closely depends on the depth.

Categorization of the mineralization processes which determine water quality, in undisturbed natural conditions.

The following physical and chemical processes take part in the formation of natural water quality:

(a) transfer and diffusion of oxygen and other gases,

(b) destruction and crushing of matter and its diffusion in water,

(c) dissolving soluble and semi-soluble components of soil, rock and air,

(d) leaching of mineral and organic particles from the soil, rock and atmosphere,

(e) mixing, diluting and chemical reacting of waters of different qualities,

(f) separation of non-soluble precipitates and sedimentation,

(g) adsorption and desorption of soluted compounds on atmospheric, soil and rock particles and exchange of ions,

(h) radioactive decay of elements.

The intensity of the physical processes of dissolving and ion exchange, and also the chemical processes, especially those of dissociation and hydrolysis, depends on the saturation of the solution by the mineral components.

The saturation index

$$I_s = \frac{-[A]\ [B]}{S} \qquad (1.70)$$

$[A]\ [B]$ - concentration of ions of the nonsoluble matter

S - constant

This saturation index indicates the direction of the reaction. $I_s = 1$ marks the balanced state, $I_s < 1$ dissolving, $I_s > 1$ separation of the insoluble component. The rate of relevant physical and chemical processes is proportional to the concentration gradient. This can be expressed by the equation

$$\frac{dc}{dt} = k \cdot (c - c_b) \qquad (1.71)$$

$\frac{dc}{dt}$	- rate of the adsorption and desorption processes	$(mg.l^{-1}, s^{-1})$
k	- coefficient of the active surface	(s^{-1})
c	- concentration of the dissolving component	$(mg.l^{-1})$
c_b	- balanced concentration	$(mg.l^{-1})$
t	- time	(s)

Integration offers

$$\ln \frac{c - c_b}{c_o - c_b} = - k \cdot t$$

c_o - initial concentration $(mg.l^{-1})$

and consequently

$$\frac{c - c_b}{c_o - c_b} = e^{-kt} \qquad (1.72)$$

With regard to the relatively small value s of the balanced concentrations, c_b can be neglected

$$c = c_o \cdot e^{-kt}$$

$$c' = c'_o \cdot (1 - e^{-kt}) \qquad (1.73)$$

These equations prove that the rate of dissolving rock and soil particles, and also the rate of dissolving deposition of insoluble compounds in the rock environment of pores and cracks, is governed by exponential relations. The speed of these processes is higher in an environment where actual concentrations differ remarkably from the balanced concentration than in an environment where this difference is small (Tab. 1.22).

The balance of running dissolving and ion exchanging processes in the system soil/rock-water depends on the solubility of the rock, on the chemical composition of the water and its temperature, on the acting surface, depending on the porosity and size of the rock particles. The dependence between the solubility and the size of the soil particles can be expressed as

$$\ln \frac{c_k}{c} = \frac{a}{r} - \frac{b}{r^4} \qquad (1.74)$$

c_k - solubility of small crystals (size 10^{-4}m)

c_r - solubility of particles with the size r

a,b- constants

Diffusion rate at the constant value of the concentration gradient

$$\frac{dn}{dt} = - f \cdot D \cdot \frac{dc}{dc_o} \qquad (1.75)$$

$\frac{dn}{dt}$ = change of the number of mols passing through area f

$\frac{dc}{dc_o}$ = concentration gradient

D - diffusion coefficient

f - area

In most cases, the concentration gradient depends on time, which describes a partial differential equation, whose solution leads to the relation

$$c = c_s \cdot \left[1 - \frac{2}{\sqrt{\pi}} \cdot \int_0^{\frac{x}{2}\sqrt{DT}} e^{-y^2} . \, dy \right] \qquad (1.76)$$

c_s - concentration of the saturated solution

y - integration variable.

TABLE 1.22

a) Dissolving of gases	b) Dissolving of minerals
$O_2,N_2,CO_2,H_2S,HCl,NH_3,Ar$ Kr,Xe,Ne,Hc	$CaCO_3,Fe(OH)_3,NaCl,FeCO_3,$ $MnCO_3$ atd.
Mixing of waters: $q_1 .\ c_1 + q_2 . c_2 = 100\ .\ q_s$	
Chemical reactions:	
A) Oxygen O:	B) Calcium Ca
$O_2 + 4\,H^+ \rightleftharpoons 2\,H_2O$	$Ca^{2+} + CO_2 + H_2O \rightleftharpoons Ca^{2+} + 2\,HCO_3^-$
$H_2O_2 \rightleftharpoons O_2 + 2\,H$	$Ca^{2+} + CO_3^{2-} \rightleftharpoons CaCO_3$
$2\,H_2O_2 \rightleftharpoons H_2O + 2\,H^+$	
C) Iron Fe	D) Manganese Mn
$Fe^{2+} + H_2O \rightleftharpoons Fe\ (OH)^+ + H^+$	$Mn^{2+} + H_2O \rightleftharpoons Mn(OH)^+ + H^+$
$Fe(OH)^+ + H_2O \rightleftharpoons Fe(OH)2 + H^+$	$Mn(OH)^+ + H_2O \rightleftharpoons Mn(OH)_2 + H^+$
$Fe(OH)_2 + H_2O \rightleftharpoons Fe(OH)_3^- + H^+$	$Mn(OH)_2 + H_2O \rightleftharpoons Mn(OH)_3^- + H^+$
$Fe^{3+} + H_2O \rightleftharpoons Fe(OH)^{2+} + H^+$	E) Alluminium Al:
$Fe(OH)^{2+} + H_2O \rightleftharpoons Fe(OH)^{2+} + H^+$	$Al(OH)_3 + OH^- \rightleftharpoons AlO_2 + 2\,H_2O$
$Fe(OH)_2^+ + H_2O \rightleftharpoons Fe(OH)_3 + H^+$	$Al(OH)_3 + H_2O \rightleftharpoons Al(OH)_4^- + H^+$
$Fe(OH)_3 + H_2O \rightleftharpoons Fe(OH)_4^- + H^+$	$Al(OH)_4^- \rightleftharpoons AlO_2^- + 2\,H_2O$
G) Sulphur S:	F) Silicon Si:
$H_2S \rightleftharpoons S + 2\,H^+$	$CaAl_2Si_2O_8 + H_2O + 2\,H^+$
$H_2O + H_2SO_3 \rightleftharpoons SO_4^{2-} + 4\,H^+$	$\rightleftharpoons Al_2Si_2O_5(OH)_4 + Ca^{2+}$
$S^{2-} \rightleftharpoons 2\,S^-$	H) Ammonia NH_4:
$3\,H_2O + S \rightleftharpoons H_2SO_3 + 4\,H^+$	$NO_3^- + H_2O \rightleftharpoons 2\,OH^- + NO_2^-$
	$NO_3 + 10\,H^+ \rightleftharpoons NH_4 + 3\,H_2O$
	$NO_3^- + 2\,H^+ \rightleftharpoons HNO_2 + H_2O$

Selected characteristic physical and chemical processes which form water (especially groundwater) quality.

Chemical reactions mostly take place when water of different origin is mixed. Where these reactions are missing, the resulting water composition is

$$q_1 . c_1 + q_2 . c_2 = 100 . c_s \tag{1.77}$$

q_1, q_2 - quantities of mixed waters (%)

c_1, c_2 - concentration of non-reacting components in mixed waters ($g.l^{-1}$)

c_s - resulting concentration ($g.l^{-1}$)

Solid particles suspended in water are affected by chemical and physical, especially hydraulic forces. Under prevailing hydraulic forces, the intensity of the interception of solid particles in rock formations is low: their function may even wash out rock particles, i.e. cause suffusion.

Under the prevailing influence of chemical forces there occurs a gradual clogging of pores and cracks, resulting in a decreasing coefficient of hydraulic conductivity.

The course of the groundwater flow depends, therefore, on the coincidental physical and chemical reactions. Slightly polar substances are intercepted quite easily and in the proximity of the infiltration point. More polar substances are difficult to intercept. The interception of basic, alkaline components depends to a remarkable extent on their chemical composition. Iwasaki (1974) states that the decrease in the concentration percentage is proportional to the actual concentration

$$\frac{c}{L} = (A + K . s) . c \tag{1.78}$$

c - concentration percentage (%)

L - the thickness of the layer (m)

A,K - constants

s - the volume of suspended particles intercepted by the unit volume of the layer

The volume of intercepted particles

$$\frac{s}{t} = -v_f . \frac{s}{L} \tag{1.79}$$

v_f - apparent velocity of the groundwater flow (m per day)

The above-mentioned physical and chemical processes usually determine the quality of natural waters in the underground, on the Earth's surface and above it. These processes are often accompanied by biological, or bacteriological processes. These usually depend not only on the thermal energy input or output, but also on luminous energy.

From the biological point of view, the complex of local physical, chemical and other conditions is named biotope. The development of a biocoenosis, a community of organisms, depends on this biotope. Each type of biotope contains not only characteristic organisms, but also organisms which are usual in other biotopes and which have been brought in by chance.

The resulting water quality can be classified

(a) formally

(b) genetically

(c) in accordance with the possibilities of its utilization,

(d) in accordance with its influence on biological factors.

Systematic categorization, enabling an explicit genetic classification on the basis of formal physical and chemical factors, has not yet been developed. Different classifications are, therefore, used for the classification of surface waters and groundwaters.

Groundwater can, in accordance with its chemical composition, be classified as

(a) plain - containing less than 1 g of dissolved substances in 1 liter,

(b) brackish - containing 1 to 30 g of dissolved substances in 1 liter,

(c) brine - containing more than 30 g of dissolved substances in 1 liter of water.

Waters which contain more than 1 g of carbon dioxide CO_2 in 1 liter are denoted as acidic. Natural mineral waters contain either more than 1 g of dissolved substances or of carbon dioxide. Waters whose healing properties have been scientifically proved may be described as medicinal, or curative. The classification and terminology of mineral waters can be derived from partial classification according to the content of dissolved gases, quantity of dissolved solids, main ion components, biologically and pharmaceutically important components, chemical reaction, radioactivity, osmotic pressure and temperature. Thermal waters whose temperature generally exceeds $25^{\circ}C$ can be classified as

- hypothermal	$25\text{-}35^{\circ}C$
- isothermal	$35\text{-}42^{\circ}C$
- hyperthermal	$> 42^{\circ}C$.

The classification of groundwaters on the basis of their chemical composition is governed either by the prevailing ions or by the prevailing ion combinations. Jetel (1975) connects both these principles in the following way and differentiates:

- classes S, C, N, Cl - according to the representative anion SO_4, HCO_3, N, Cl,
- groups according to the representative cation Na, K, Mg, Ca.

Contents (r) of other ions are to be added to the similar main ions: Mn^{2+} and Fe^{2+} to Ca^{2+}, NO_3 to Cl^- and alkalines to Na^+, CO_3^{2-} to HCO_3^-.

- species according to combinations of the main ions, marked by Roman figures (Tab. 1.23).

The symbol of groundwater quality forms marks of the main anions, cations, species and the value of the total mineralization, e.g. Cl_{II}^{Na} - 0,014 $g.l^{-1}$.

TABLE 1.23

Index	Hypothetical combinations of ions			Characteristic relationship
1	NaCl	Na_2SO_4	$NaHCO_3$ $Mg(HCO_3)_2$ $Ca(HCO_3)_2$	$rHCO_3 > r(Ca+Mg)$ $rNA > r(Cl+SO_4)$
II	NaCl	Na_2SO_4 $MgSO_4$	$Mg(HCO_3)_2$ $Ca(HCO_3)_2$	$rHCO_3 < r(Ca+Mg) < r(HCO_3+SO_4)$ i.e. $r(Cl+SO_4) > rNa > rCl$
	NaCl	Na_2SO_4 $MgSO_4$ $CaSO_4$	$Ca(HCO_3)_2$	
IIIa	NaCl $MgCl_2$	$MgSO_4$	$Mg(HCO_3)_2$ $Ca(HCO_3)_2$	$rCa < r(HCO_3+SO_4) < r(Ca+Mg)$ i.e. $r(Na+Mg) > rCl > rNa$
	NaCl $MgCl_2$	$MgSO_4$ $CaSO_4$	$Ca(HCO_3)_2$	
IIIb	NaCl MgCl $CaCl_2$	$CaSO_4$	$Ca(HCO_3)_2$	$rCa > r(HCO_3+SO_4)$ i.e. $rCl > r(Na+Mg)$
IV	NaCl	Na_2SO_4 $MgSO_4$ $CaSO_4$ H_2SO_4		$rHCO_3 = 0$
	NaCl MgCl	$MgSO_4$ $CaSO_4$ H_2SO_4		
	NaCl $MgCl_2$ $CaCl_2$	$CaSO_4$		

Groundwater categorization depending on ion combinations according to Alekin (1970). IV. species categorized according to Florea (1970).

The quality of surface waters depends to a great extent on hydrological, meteorological and antropogenetic factors and changes considerably with time. The chemical composition of these water is not the prevailing agent of their quality.

Their mineralization is remarkably lower than that of groundwaters. The quality of surface water, therefore, is to be defined on the basis of further characteristic indicators: oxygen regime, microbiological and other indicators (Tab. 1.24) and also with regard to its possible utilization.

Oxygen demand is the ability of substances to utilize the dissolved oxygen for their stabilization. Chemical oxygen demand (COD) is a measure (mg $O_2 . l^{-1}$) of the materials present in water which may be readily oxidized, in order to ascertain the amount of organic and reducing material. Biochemical oxygen demand (BOD) is the amount of oxygen used by aquatic microorganisms in their metabolic processes. The evaluation of the $\frac{BOD}{COD}$ ratio offers the basic information concerning the content of biologically degradable and resistant constituents.

The amount of dissolved oxygen ($mg.l^{-1}$) depends on the water quality. The difference between the maximum possible concentration of oxygen (depending on temperature - Tab. 1,24) and the actual concentration is termed oxygen deficit. Their ratio is marked as saturation ratio.

Very important indicators of the basic chemical composition include the value of total dissolved solids TDS ($mg.l^{-1}$) and the value of suspended matter: The concentration of ions is presently followed in multiples of the content of matter in a system, whose number of molecules is equal to the number of atoms in 12.10^{-6}kg of carbon isotope ^{12}C at 1 liter ($mmol.l^{-1}$). The ions Cl^- , SO_4^{2-}, Na^+ , Mg^{2+} are followed in particular.

Constituents which may significantly affect the application of water for beneficial purposes include cations, namely sodium Na^+, potassium K^+ , iron Fe^+ , and Mn^+ , calcium Ca^+ and magnesium Mg^+ , further anions, namely nitrate NO_3^- and ammonia NH_4^- , chloride Cl^- , and also bicarbonates HCO_3^- , carbonate CO_3^- and hydroxide OH^+ , are generally considered in the light of their influence on alkalinity and hardness. Nonionic constituents include especially detergents, oily substances, phenols, cyanides and silica (silicon dioxide SiO_2).

The sum of the calcium and magnesium, expressed as an equivalent amount of calcium carbonate, was in the past followed as water hardness and is presently replaced by a separate following up of both these cations.

An important integrated property of water is the hydrogen-ion concentration pH, the negative logarithm to the base 10 of the hydrogen-ion concentration. A balance between dissociated hydrogen and hydroxyl ions denotes a pH value of 7.0, but this value has no special significance as an expression of alkalinity and acidity.

TABLE 1.24

Class Characteristics		Ia very clean	Ib clean	II slightly polluted	III intensively polluted	IV deteriorated
a) Indicators of the oxygen regime						
Dissolved oxygen	$mg.l^{-1}$	>7	>6	> 5	>3	<3
BOD_5	$mg.l^{-1}$	<2	< 5	<10	<15	>15
Oxidizability by permanganate	$mg.l^{-1}$	< 5	<10	<15	<25	>25
Saprobity		oligo-	β-meso	β-α-meso	α- meso	poly -
b) Chemical indicators						
Dissolved matter	$mg.l^{-1}$	<300	<500	<800	<1200	>1200
Suspended matter	$mg.l^{-1}$	< 20	< 20	< 30	< 50	> 50
c) Special indicators						
pH		>6.5 <8.5	6.5-8.5	6.0-8.5	5.5-9.0	<5.5 >9.0
Temperature 1)		<22	<23	<24	<26	>26
d) Indicators of the microbiological pollution						
Coli index	$pc.l^{-1}$	$<10^3$	$<10^4$	$<10^5$	$<10^6$	$>10^6$

Categorization of surface waters according to the basic indicators of water pollution. 1) in mild continental climate.

A pH value of less than 7.0 denotes an excess of hydrogen ions. Such water is called acid, when the value pH is below 4.5. Acidity is caused especially by free mineral and carbonic acids.

A pH value greater than 7.0 indicates an excess of hydrogen ions, the alkalinity, i.e. ability of this water to neutralize acids. Alkalinity is caused especially by hydroxide, carbonate and bicarbonate.

Sensoric properties such as temperature, colour, turbidity, odour and taste are very important indicators of water quality, influencing its utilization for drinking purposes. They depend on the chemical and biological composition of water.

A simple indicator of water quality is the electrical conductivity, which depends on the concentration of salts and refers to the conductivity of a one centimeter cube of water at 25°C.

On the basis of a set of graduated indicators, surface waters can be divided into several classes (Tab. 1.24). For a better appreciation several less favourable values should be collected uniformly throughout the year. The classification itself has to be carried out using arithmetic means after the exclusion of values influenced by heavy rainfalls.

On the basis of the utilization of water for drinking purposes three classes can be distinguished

(a) water requiring mechanical treatment and disinfection only, or not requiring treatment at all,

(b) water requiring chemical treatment,

(c) water not suitable enough or unsuitable for municipal water supply (high content of nitrogen N, iron Fe or manganese Mn, high mineralization, high content of calcium Ca and magnesium Mg, fluorine F, as well as chemically, radiologically, biologically or bacteriologically highly polluted waters.

1.8 HYDROLOGIC CYCLE AS REGULATOR OF BIOLOGICAL PROCESSES

The biosphere is an open system of anorganic and organic, living and withered matter on the Earth, transforming solar energy by means of biological processes.

Complicated water solutions of compounds of biogenetic elements formed by living environment before the gradual development of life (Fig. 1.24). The presence of bioelements in water as well as the presence of water in nutriments and organic matter has a basic importance for metabolic processes and the growth of living matter (Tab. 1.25).

Living matter, or biomass, consists of the three main components: proteins, fats and sacharids. During the decay of organisms, absorption and disintegration of nutriments, water is separated from all the mentioned components (Fig. 1.24). The circulation of matter can be followed as chains of nutrition in connection with the energy transfer. Organisms can be classified according to their relation to this circulation as producers, consumers and destructors.

Organisms which produce organic matter by consuming anorganic compounds (microorganisms, flora etc.) are termed as producers. Organisms which are not able to perform this function and consume organic matter, formed by producers and other consumers, are termed as consumers (animals etc.). Destructors are organisms which decompose organic matter, thus producing anorganic components (microorganisms etc.) (Tab. 1.26).

In ecosystems, subsystems of the biosphere, these biotic communities are interconnected with the biotic and abiotic environment, thus forming closed and independently functioning systems, whose biotic structure, cycle of nutriments and energy flow can be exactly determined (Fig. 1.25).

The energy in ecosystems basically has three forms:

- kinetic energy - whose manifestation is external motion,

- potential energy - accumulating in relation to the environment,
- internal energy - manifestated by internal movement and internal relations.

TABLE 1.25

Function	Description
Solution dissolving agent	Water dissolves partially or fully the prevailing part of anorganic and organic compounds, forming foodstuff and nutriments.
Chemical activator	Water fissure molecules, whose motion enables the approach and mutual reaction of molecules of different compounds.
Heat regulator	The high specific heat of water helps to maintain a stable temperature of organisms, without any excessive fluctuation. Its optimum value is kept by evaporation of water, which requires much energy because of the high latent heat of vaporization.
Regulator of pressure	The propagation of pressure in water is uniform in all directions, enabling the cellular pressure to be maintained and regulated.
Distribution agent	Water distributes nutriments to and takes away wastes from different parts of the living organisms. It also distributes thermal energy, transporting it from the place of its occurrence (muscles, liver) to cooler parts (skin).
Economizing agent for water organisms	The unit mass of these water organisms is the same as that of water, thus removing the effect of gravity and economizing their energy requirements.

Functions of water in the biomass.

The transformation of energy during its flow through the ecosystem takes place in accordance with physical laws, its manifestation is the motion and change of temperature. Apart from the law of conservation of energy, the law of degradation of energy also makes itself felt: during any process part of the energy changes into heat. According to Darwin-Lotke's rule, during a quick change of energy, heat losses form one half of the energy total. Due to these energy losses, any process in ecosystems is one-way. Heat losses are transferred to the environment.

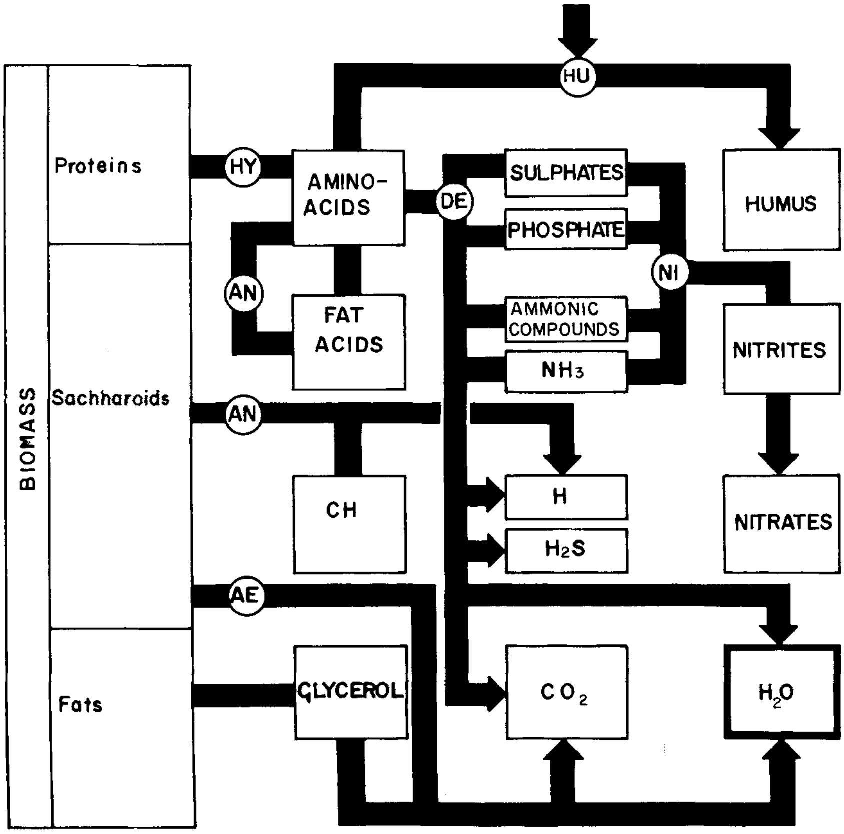


Fig. 1.24. Block diagram representing the decay of the basic components of the biomass. Processes: AE - aerobic, AN - anaerobic, HY - hydratation, HU - humification, NI - nitrification, DE - decarboxylation, dezamination, hydrolysis.

The ecosystem is a community, including all organisms of a certain space, and abiotic elements, acting in the climatic regime, whose relation with the environment is stable and determined, by the input and output of energy and matter.

The formation of regional ecosystems depends on topographical, soil and climatological conditions with the prevailing influence of solar radiation (temperature) and rainfall (Tab. 1.27). In natural ecosystems of forests, steppes, deserts, water formations etc. the input of matter is not significantly affected by human activities (pollution) and solar radiation is the sole source of energy.

In the framework of these regional ecosystems, depending on the macroclimate, local ecosystems exist, influenced especially by the local topographic, soil and moisture conditions and by the conditions of the meso- and microclimate. With regard to the wetness of their habitats, plants can be classified in four groups (Tab. 1.28).

TABLE 1.26

Organic groups (species)	Nutriment substance	Production	Process	Simplified formula
Producers (green plants, sulphur bacteria)	anorganic compounds of biogeo-elements	oxygen biomass	Photosynthesis	$6\ CO_2 + 6\ H_2O + 1612\ kJ = C_6H_{12}O_6 + 6\ O_2$
			Chemosynthesis	$6\ CO_2 + 6\ H_2S + \text{energy} = C_6H_{12}O_6 + 6\ S$
			Biological uptake	$265\ CH_2O + 16\ NH_3 + PO_4 + 146{,}5\ O_2 + \text{energy} = C_{106}H_{180}O_{45}N_{16}P + 159\ CO_2 + 199\ H_2O$
Consumers (herbivore, carnivore omnivore)	organic matter: proteins, sachharoids fat	waste organic matter CO_2 H_2O	decomposition of sachharoids	$C_6H_{12}O_6 + 6\ O_2 = 6\ CO_2 + 6\ H_2O + 161{,}1\ kJ$
Detrivore (fungi, bacteria)	Waste organic matter	CO_2, H_2O NH_3, SO_4 PO_3, salts	decomposition	

Categorization of nutrient relationships in terrestrial or aquatic ecosystem. Nutrients continually circle within the boundaries of the ecosystem. Fluxes across the ecosystem's boundaries link it with the environment, i.e. nutrient, soil and atmospheric compartment.

Four zones of plant communities can be distinguished along the water courses and reservoirs as a result of the impact of water level fluctuation (Fig. 1.26, Tab. 1.29).

Living processes and changes in ecosystems are managed by the transfer of matter, especially by means of water and air, and energy. Under the influence of human activities stable natural ecosystems change into (often unstable) anthropogenetic ecosystems

(a) agricultural ecosystems, where the dominant source of energy is solar radiation and the climate determines the type of cultivated agricultural plants, but the flow of matter and energy is significantly influenced hy human activities,

(b) industrial ecosystems, where in addition to the supplemented input of matter, the energy input is considerably supplemented by man-made energy,

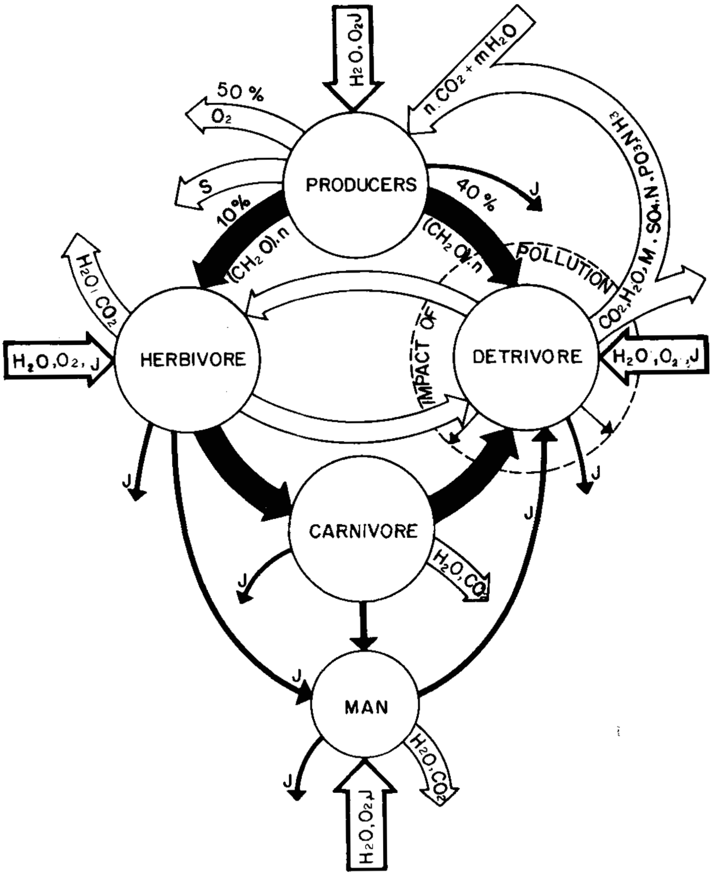


Fig. 1.25. Circulation of water and nutriments in ecosystems: the transfer of matter including water, proteins, sachharoids, fat etc. is represented by thick black arrows, energy input and output by thin arrows. The impact of pollution (dotted arrow) changes the input of nutriments, improving living conditions for detrivores, thus changing the ecological succession to the disadvantage of carnivores first. J - energy

TABLE 1.27

Temperature Rainfall	Minimum	Optimum	Maximum
Maximum	-	steppe	desert
Optimum	tundra	mixed and coniferous forest	tropical monsoon forest
Maximum	-	salty marshes	tropical jungle

The dependence of the basic natural ecosystems on temperature and rainfall.

TABLE 1.28

Category	Characteristics	Selected species
Hydrophytes	grow in water swamps and bogs	water plants, phytoplankton, pondweed, rice, cattail
Hygrophytes	need moisture, not resistant against drought	ferns, moss
Mesophytes	medium water requirements survive both short periods of wetness and drought	grasses, wood
Xerophytes	low water requirements drought-resistant	cacti, haloxylon

Categorization of plants with regard to the relative wetness of their habitats.

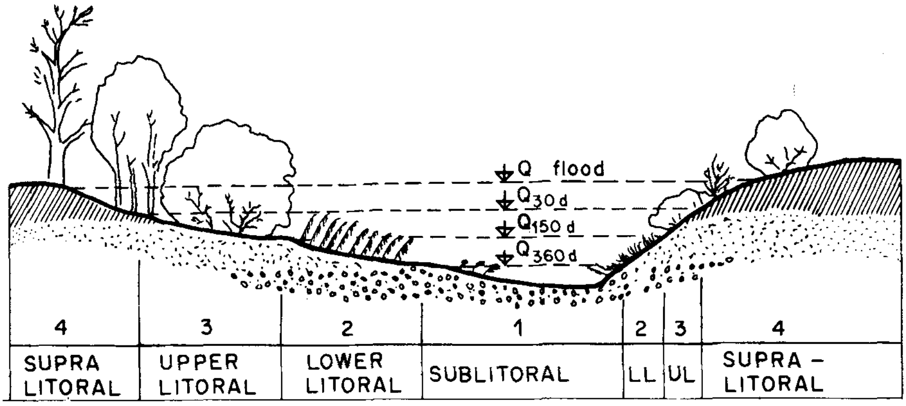


Fig. 1.26. The development of the flora in the sublitoral (1), litoral (lower - 2, upper - 3) and a supralitoral (4) zone along a stream or a reservoir shore depends on the frequency of flooding and on the groundwater table. Soft species occur especially in the litoral zone, hardwood species in the supralitoral zone.

(c) degradated ecosystems, where the additional input of matter (water, air and/or soil pollution) changes the input of energy and results in the disintegration of stable natural ecosystems.

TABLE 1.29

Zone	Characteristics or limitation	Characteristic	
		wood	other species
Supra-litoral	flood only exceptionally	Hard: oak, maple, elm, ash-tree	grasses,
	water table Q_{30d}		
Upper-litoral	occasionally flooded	Soft: willow, poplar, alder	bramble, nettle, meadow-sweet, rush, willow-herb
	water table Q_{150d}		
Lower litoral	frequent water table fluctuation	Soft: willow, alder, hedge,	brooklime, bulrush, monkey flower, water celery, water mint, water forget-me-not, reed, reed-grass
	water table Q_{360d}		
Sub-litoral	permanently flooded	no trees	bog-pond weed, bulbous rush, floating clubrush, water-lilly, water crow foot, strapweed

Classification of shore zones according to the frequency of their flooding and characteristic representatives of relevant plant ecosystems.[1]

1.8.1 Interrelationship of Aquatic Ecosystems and Water Quality

An aquatic ecosystem, i.e. a system of organic communities in a certain water formation, where this community lives in a stable relation with its abiotic environment, represents an example of a closed chain of nutrition.

In this enviornment, Liebig's law of minimum is of particular significance: The missing energy or bioelement decides on the course of the production process. It is impossible to prove the presence of such limiting bioelements (usually phosphorus P, potassium K, nitrogen N etc.) in water, because it is changed into organic matter immediately after its entry into the system.

The production process is, in addition to this, also governed by the presence of matter stimulating the growth, primarily of vitamins: tiamin, biotin, cobaltamine etc. The consequence of a high content of nutriments and stimulating matter in water is eutrophication, manifestated by its biological overproduction.

Uhlmann (1975) expresses the circulation of matter in the aquatic ecosystem by the differential equation. The change in the concentration of biomass can be formulated as follows:

$$\frac{dZ}{dt} = {}_2k_{21}.P.Z + {}_2k_{24}.B.Z - {}_1k_{02}.Z - {}_1k_{42}.Z - {}_2k_{32}.F.Z \qquad (1.80)$$

Z - zooplankton

P - phytoplankton

B - bacteria

F - fishes

${}_1k$ - coefficient of rate for reaction of the 1.range

${}_2k$ - coefficient of rate for the reaction of the 2.range

The index on the right of the coefficients marks the receiver (1st number) and the supplier (2nd number).

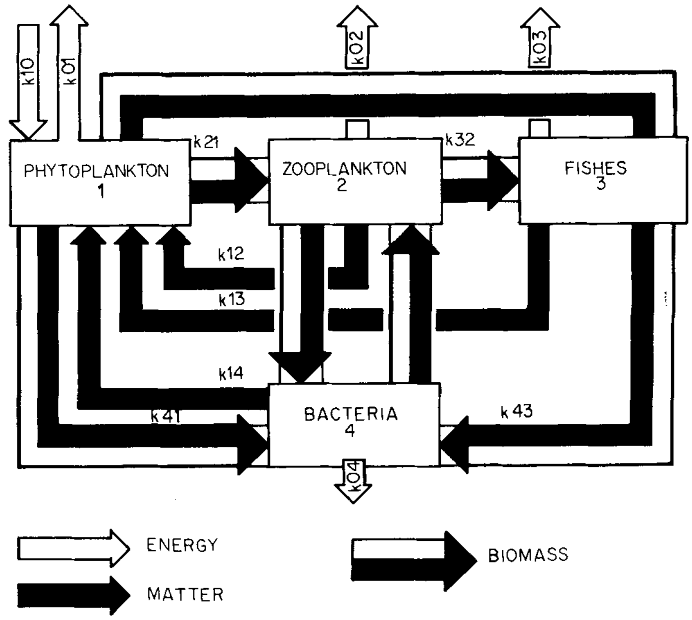


Fig. 1.27. Couplings of the aquatic ecosystem according to Uhlmann (1975): phytoplankton (1), zooplankton (2), fishes (3), bacteria (4). Index ij marks the transfer of matter or energy from the part j to the part i of the ecosystem, 0 - the abiotic environment.

The biomass in flowing waters is influenced by the inflow, as can be expressed by the differential equation

$$\frac{dX}{dt} = \mu . X + Q . X' - (i_1 + i_2) . X \qquad (1.81)$$

X - biomass	(g)
μ - daily increment ratio	(d^{-1})

Q - water inflow $(\mathrm{l.d}^{-1})$

X' - specific inflow of the biomass $(\mathrm{g.ll}^{-1})$

i_1 - ratio of biomass outflow (d^{-1})

i_2 - ratio of biomass utilization by consumers and destructors (d^{-1})

In newly constructed river channels and after the flood

$$\frac{dX}{dt} \gg 0. \tag{1.82}$$

After this initial phase, the value of the biomass increment decreases to the value of the biological balance, characterized by

$$\frac{dX}{dt} = 0 . \tag{1.83}$$

The water quality with regard to its biological properties can be classified in terms of its saprobity or trophicity. The saprobity is the biological state of water, determined on the basis of the presence or absence of biocoenosis, that is characteristic for a certain degree of biochemical decay, i.e. in relation to the degree of pollution (Tab. 1.30).

The saprobity characterizes the changing properties of the water environment during a certain period, thus differing from the physical and chemical indicators, which characterize the actual state only. For this reason, an explicit relation between the saprobity and relevant physical and chemical indicators does not exist, though the classification system of saprobity is closely interrelated with the biological oxygen demand (BOD) of water (Fig. 1.28).

The trophicity is the ability of water to nourish water organisms. Nauman (1932) classified the trophicity on the basis of the surplus, average or undersized values of the basic physical and chemical preconditions of the development of different biocoenosis. Seven basic trophical types of stagnant and flowing waters follow as a consequence of combining these criteria (Tab. 1.31).

Some changes and combinations are chemically impossible, such as alkalitrophicity with siderotrophicity, acidotrophicity and dystrophicity, or acidotrophicity with eutrophicity. A practical classification is based on the mean annual values.

The views on the interrelationship of the saprobity and the trophicity have not yet been unified. Kolkwitz (1935), Sládeček (1961) and others state that the eutrophization and self-purification processes are only two directions of one natural process, relating in this way the degree of saprobity with the degree of trophicity. The scale of trophicity is then the ratio of the production of living matter to respiration, liberating the organic energy, bonded by living matter.

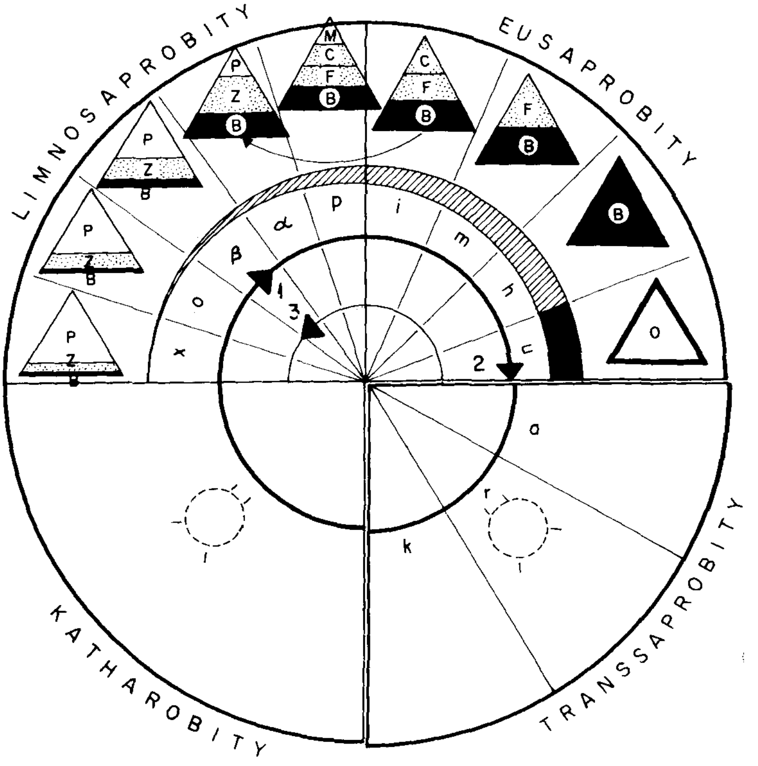


Fig. 1.28. Circular representation of water quality by degree of saprobity and relevant structure of aquatic ecosystems according to Sládeček (1972): Limnosaprobity: (x-xeno, o-oligo, β-betamezo, α-alphameso, p-polysaprobity), Eusaprobity (i-iso, m-meta, h-hyper, u-ultrasaprobity), Transaprobity (a-anti, r-radio, k-kryptosaprobity). Detrivore marked black: B-bacteria. Consumers-herbi-, carni-and omnivore marked dotted: Z-zooplankton, C-ciliata, F-zooflagellata. Producers- green plants and autotrophic bacteria marked white: P-phytoplankton, M-mixotrophic flagellata. O-no life. Arrows show the direction of eutrophization (1) and further pollution (2), decay and self-purification (3). Dsahed segment represents the increase in material input and output (and also the increase in the biological oxygen demand).

TABLE 1.30

Symbol Saprobity			Process	Symbol	Degree of saprobity	Pollution (degree)	BOD$_5$ mg.l^{-1}	Plankton (in ml^{-1})	Chlorophyle (mg.m^{-3})	Primary production (mg.m^{-3} .d^{-1})	(g.m^3 .year^{-1}	Decomposition	Salts	Input output	Trophicity
K Katarobity	clean water		abiotic no life			without pollution	0	1	0	0	0				atrophicity
L Limno-saprobity	polluted water	self-purification/eutrophization	completed oxydation	x	xenosaprobity	very light	1	100	1	50	10				ultra-oligotrophicity
			completed reduction	o	oligosaprobity	light	2,5	1000	5 (20)	100 (1500)	30 (500)				oligotrophicity
			starting oxydation	β	mesosaprobity	medium	5	10^5	300 (1500)	500 (12000)	150 (4000)				β-eutrophicity
				α	mesosaprobity	medium	10	10^6	1000	1500	300				α-eutrophicity
			predominant reduction	p	polysaprobity	heavy	50	10^7	10000	12000	4000				polytrophicity
E Eusaprobity	biologically degradable matter		biological	i	isosaporbity	(ciliatic)	400	1000	3	100	30		growth		isotrophicity
			and	m	metasaprobity	(hydrogen sulfidic)	700	0	+	+	+				metatrophicity
			chemical (see	h	hypersaprobity	(bacteriologic)	2000	0	0	0	0				hypertrophicity
			column 8)	u	ultrasaprobity	(abiotic)	12.10^4	0	0	0	0				ultratrophicity
T Trans-saprobity	biologically non-degradable matter	abiotic	chemical, abiotic, anorganic, no life	a	antisaprobity	toxic									
				r	radiosaprobity	radioactive									
				k	kryptosaprobity	physical									

Categorization of water quality according to its saprobity and trophicity. Relationships of the metabolism, self-purification processes and the trophicity. Values concerning chlorophyle and primary production are related to the result of the eutrophization process.

TABLE 1.31

Physical and chemical indicators			pH	Poly-typus	Meso-typus	Oligo-typus
Summer temperature just below the water table		°C		> 20	15- 20	< 15
Calcium oxide	CaO	mg.l^{-1}		>100	25-100	3- 25
Nitrogen	N	mg.l^{-1}		> 40	1- 0	0
Phosphorus	P	mg.l^{-1}		> 25	1.5	0
Humus acids		mg.l^{-1}		80-100	25- 50	25

Biotypus	Characteristics Occurrence	pH			
alcalitrophic	poor plankton carst	> 7	Ca	-	N,P,Fe
siderotrophic	limonitic bed		Fe		
argilotrophic	high turbidity loamy and silt sediments		humus matter		
eutrophic	high content of nutriments rich phytoplankton mud, marshes bottom hydrogen sulphide	> 7	N,P	N,P	
oligotrophic	clear water, poor phytoplankton, poor flora in the neighbourhood new reservoirs, mountain lakes	$\doteq 7$			N,P,Ca
acidotrophic	turfic bed	< 5,5			Ca
dystrophic	browny water, poor phyto-plankton occasionally rich zooplankton marsh without odour	< 7	humus matter		Ca,P,N
hypertrophic	heavily polluted water, degradated aquatic life				

Categorization of water reservoirs, their trophic- and biotypes, physical and chemical indicators according to Nauman (1932), completed by Thienemann (1942) and Šrámek/Hušek (1948).

1.9 RUNOFF PROCESS AS REGULATOR OF THE LIVING ENVIRONMENT

Basic characteristics influencing the life and its environment gradually change with a characteristic gradient in the course of the longitudinal profile of the river. They also have an important impact on the utilization of water for different beneficial uses as well as on the ecosystems and the utilization of the neighbouring area.

Some of these characteristics, namely geomorphological and hydraulic, and also selected hydrological characteristics, are quite independent of the season. The course of the rest depends on the annual season and on the climate, especially in the case of rivers running through different climatological zones (Fig. 1.29).

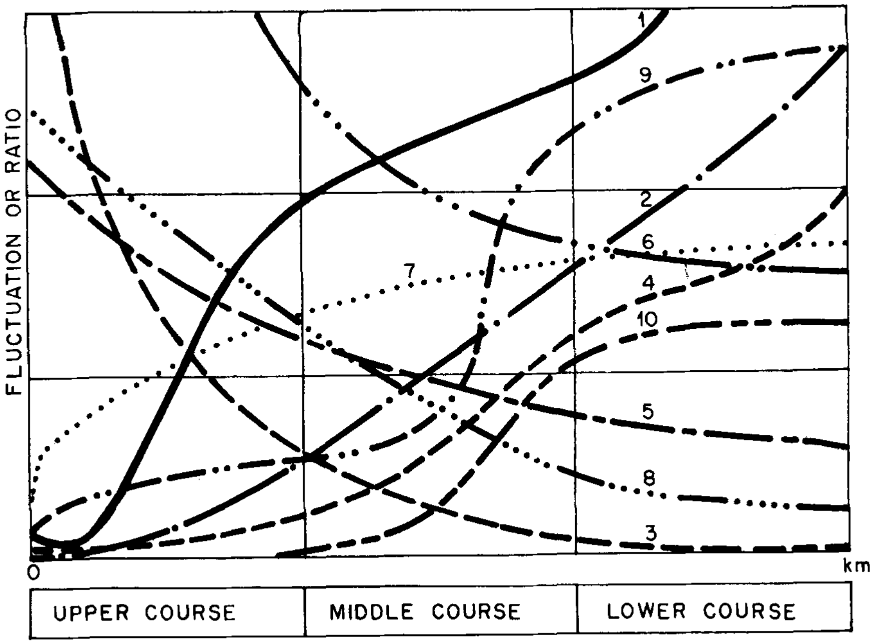


Fig. 1.29. The development of basic characteristics of a water course represented in its longitudinal profile, offering important information for the concept of agricultural, industrial and urban development: 1 - valley width, 2 - average discharge, 3 - river channel slope, 4 - river channel width, 5 - velocity of flow, 6 - fluctuation of discharge, 7 - water temperature, 8 - aeration, 9 - biochemical and chemical oxygen demand, 10 - sedimentation rate.

The suitability of a riverside area for different development purposes can be derived from the course of the mentioned curves by analysing the relevant interrelationships. Some relations appear to be contradictory: Lower and middle course areas are more suitable for the location of housing and industrial estates due to the necessary space. From the water supply point of view, upper courses are more suitable having more favourable conditions for reservoir construction.

Upper courses offer a high head and thus more favourable conditions for water power generation - with reservoirs as well for peak power production. Lower courses with high, sustained discharges offer conditions for the construction of run-of-river plants with lower heads. They offer only restricted possibilities for the construction of reservoirs, but favourable conditions for inland navigation.

Protection against floods can be achieved more economically by a proper location of the relevant estates out of the flood plain at the upper courses if relevant valleys are sufficiently wide, and by river training in the middle and lower courses.

The longitudinal proportion of the water course in an open area creates a natural traffic artery: not only its channel, but also the riversides on both banks. Fluctuations in the water table limit the beneficial utilization of the flooding area for communication, housing and industrial estates, which have to be protected from overflooding. Nevertheless, this zone usually has exceptional climatological and aesthetic conditions, extremely suitable for recreation. It is, therefore, useful to reserve this zone, especially in cities, for pedestrain communications or industrial quarters by a zone of plants or by a communication line (Fig. 1.30).

Communication, recreational, industrial and housing areas are mutually interrelated. The location of a housing estate in the second zone is aesthetically and functionally favourable when it corresponds to the topography, and provided that flood protection measures are realized.

The function of water courses for housing and industrial estates is twofold: they are used as sources of water supply and as recipients of waste disposal. These two functions are contradictory. Besides, the utilization of water courses for waste disposal is contradictory with their urban and aesthetic functions.

The aesthetic enjoyment of water as an intangible natural factor is manifestated in three dimensions

- in the water scenery, created by the synthesis of the water formation and its surroundings namely river banks,
- in the visual unit - i.e. the part of the landscape which is visually accessible from one place,
- in the landscape unit, i.e. formed by a geographically complex area.

The aesthetic properties of water are formed by its prominence, especially by its

- continuity and comprehensiveness,
- diversity,
- vivacity and variability.

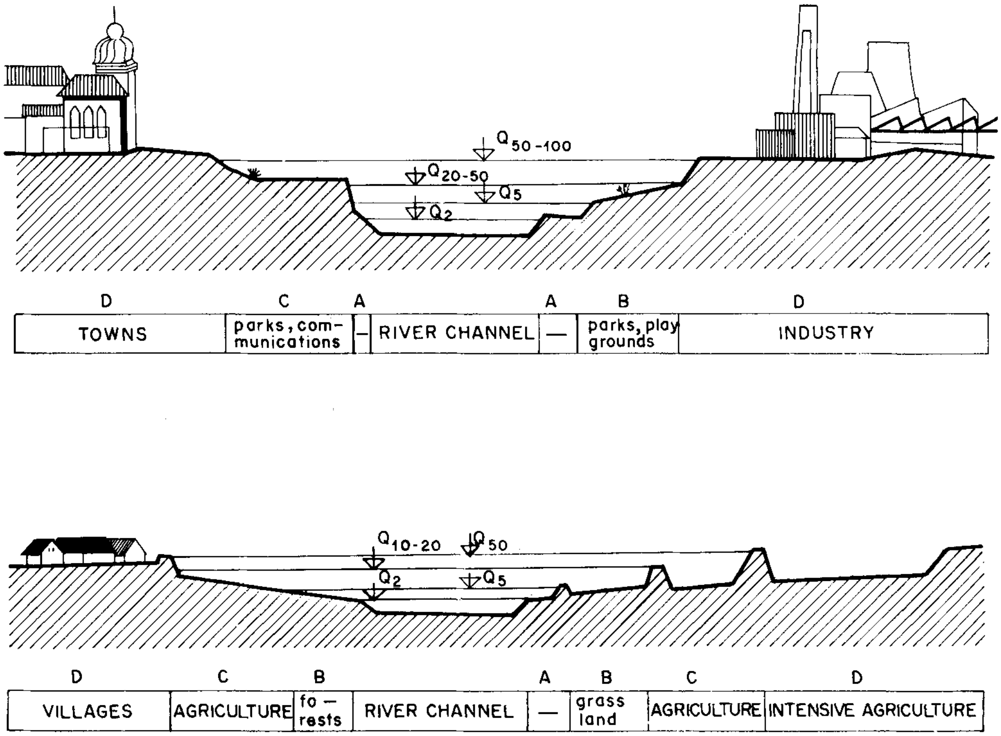


Fig. 1.30. Regulation of flood plain use derived from frequency of flooding and differing in the town and outside the town: A - floodway, B,C - partially protected zones, D - protected zone.

These properties depend on qualitative and quantitative factors. The aesthetic expression of a reservoir or a stretch of stream follows from its comprehensive continuous form, limited by the diversified banks, whose edges form limits between the water and riverside ecosystems. In a built-up area three categories can be distinguished on the basis of the dimension ratio of the housing, industrial, recreational etc. area to the width of the water level:

(a) dimensions of the water table are not remarkable with regard to the dimensions of the relevant area,

(b) dimensions of the water level do not differ remarkably from the dimensions of the relevant estate,

(c) dimensions of the water table exceed the dimensions of the relevant estate.

The resulting aesthetic enjoyment depends in any category on the quality of the architectural structure, but the medium category offers the most favourable conditions for an aesthetic and balanced effect of the water table in an urban setting. Architectural complexes in these conditions evoke the strongest aesthetic emotions.