THE HYDROLOGICAL CYCLE AND HUMAN IMPACT ON IT

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Summary

The hydrological cycle is usually called a recurring consequence of different forms of movement of water and changes of its physical state on a given area of the Earth. The role of different processes in the hydrological cycle and their description depends on the chosen spatial-temporal scales. The terrestrial hydrological cycle is of special interest as the mechanism of formation of water resources on a given area of the land. The main processes of this cycle include: precipitation; formation of snow cover; snow metamorphosis and formation of ice; melting of snow and ice; interception of precipitation by vegetation cover and storage in land surface depressions; infiltration of water into soil and vertical transfer of soil moisture; evapotranspiration; recharge of groundwater and ground flow; river runoff generation; and movement of water in river channel systems. The global hydrological cycle is produced by water exchange between the atmosphere, the land, and the oceans, and its main components are precipitation on the land and the oceans, evaporation from the land and the oceans, and runoff from the land to the oceans. Current scientific understanding of main processes qualitative peculiarities, and models of components of the terrestrial and global hydrological cycle are considered. The peculiarities of the modeling of the hydrological cycle of a river basin is demonstrated, taking into account the lack of measurable characteristics of environment. Estimations of influence of irrigation, land treatment, deforestation, and other human activities on the terrestrial hydrological cycle are presented. The role of the terrestrial hydrological cycle in the global climate system and global change is examined. The possible hydrological consequences of human-induced climate change are also discussed.

1. Introduction

The hydrological cycle is usually called a recurring consequence of different forms of movement of water and changes of its physical state in the nature on a given area of the

Earth (a river or lake basin, a continent, or the entire Earth). The movement of water in the hydrological cycle extends through the four parts of the total Earth system—atmosphere, hydrosphere, lithosphere, and biosphere—and strongly depends on the local peculiarities of these systems. The terrestrial hydrological cycle is of a special interest as the mechanism of formation of water resources on a given area of the land. The global hydrological cycle is also often considered, taking into account its role in the global climate and other geophysical processes. It is obvious that the role of different processes in the hydrological cycle and their description have to depend on the chosen spatial- temporal scales. The main components of the terrestrial hydrological cycle and the global hydrological cycle are presented in Figures 1 and 2.

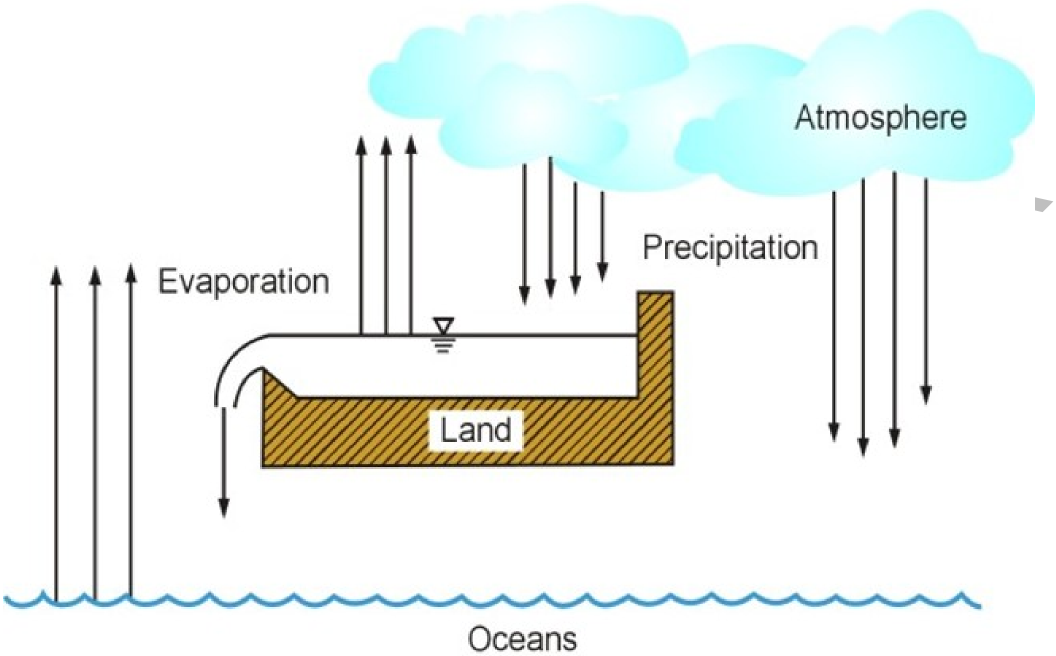

Figure 1 Global hydrological cycle are presented .

The generation of precipitation is commonly considered as the beginning of the terrestrial hydrological cycle. The precipitation may be in the form of rainfall or snow. The falling snow forms the snow cover where the snow may change its properties and may partially transform into ice. The rain or melt water may be intercepted by vegetation cover or detained by land surface depressions, may infiltrate into the soil, or may run over land surface into streams. The infiltrated water may store in the soil as soil moisture or may percolate to deeper layers to be stored as groundwater.

During cold periods a portion of infiltrated water may freeze in the soil. A part of water intercepted by vegetation, accumulated in the land surface depressions, and stored in the soil, may return back to the atmosphere as a result of evaporation. Plants take up a significant portion of the soil moisture from the root zone and evaporate most of this water through their leaves.

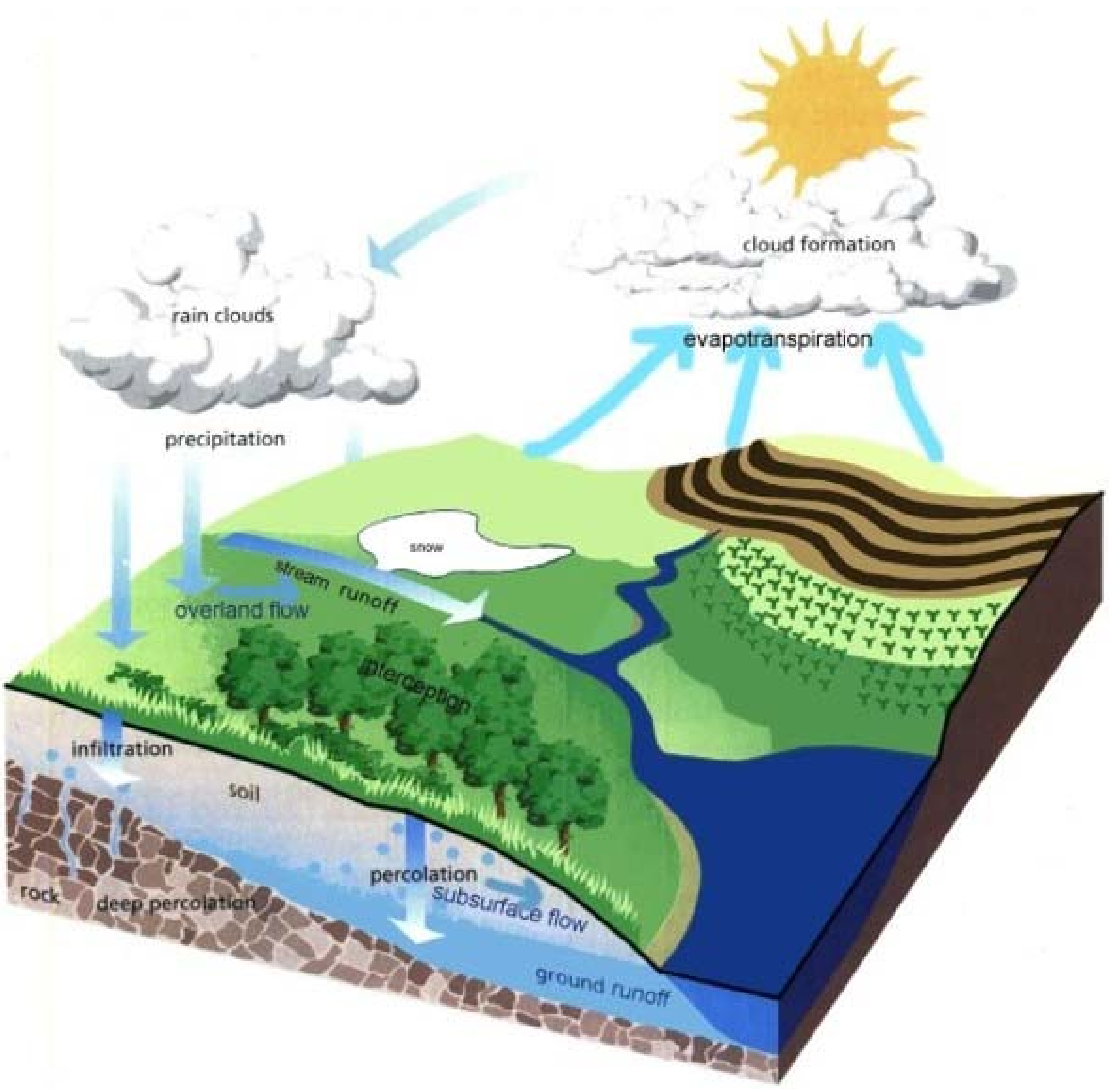

Figure 2. Terrestrial hydrological cycle.

Beside water which travels to the streams over the land surface, the stream runoff also includes water which moves to the streams through the upper soil horizons, flows out from deep layers as springs, and seeps directly in the river channels. The water collected in the river channel systems flows to lakes, seas, and oceans.

When we consider the global hydrological cycle, the principal process is water exchange between the atmosphere, the land, and the oceans. In this case, the main components are the precipitation on the land and the oceans, the evaporation from the land and the oceans, and the runoff from the land to the oceans. The movement of water in the hydrological cycle is linked with erosion and transport of sediments and chemicals. The erosional and depositional effects of streams, waves, and ice have produced a diversity of Earth's landscapes that make the Earth's surface unlike that of any other planets.

2. The Terrestrial Hydrological Cycle

The key component of the terrestrial hydrological cycle is generation of river runoff and movement of water in the river networks. The main land area units where this process occurs is the river watersheds. The sizes of these areas vary from tens of to 6900 square km (the Amazon River catchment area). Within these areas, distinct spatial differences, in topography, geology, vegetation, soil properties, land use, and meteorological conditions may be well-expressed even on small scales. The land surface heterogeneity may be essentially strengthened by human activities, that can cause a significant modification of the characteristics of the natural landscapes. Therefore, to describe the terrestrial hydrological cycle it is important not only to single out the main processes, but also to take into account the relevant topographic, geological, vegetation, and soil parameters that control runoff generation conditions and give an opportunity to represent the land surface heterogeneity.

2.1. Precipitation

Precipitation is the principle source of the Earth's water supply and may occur in liquid (rain) and solid (snow) forms. The production of the precipitation results from condensation of small water vapor droplets around available nuclei, or from ice crystal process in the clouds. Water droplets are increased in size by means of collision and coalescence until they attain approximately 2 mm in diameter; under action of gravity they then begin to descend to the Earth's surface forming the rainfall. Ice crystals may also collide and stick to one another, forming snowflakes. These snowflakes can reach the ground in the form of snow or rain, depending on the temperature of the lower atmosphere. For the condensation of water vapor or the creation of ice crystals, it is necessary for the moist air to cool to a sufficient extent and generate lift. Precipitation can be classified into four main types according to the air lifting mechanism: (1) frontal precipitation, where the lifting is due to relative movement of two large air masses; (2) precipitation caused by horizontal convergence; (3) convective precipitation; and (4) orographic precipitation. Each type rarely occurs alone in nature, but some may dominate under certain conditions.

Frontal or cyclonic precipitation occurs at convergence of air masses of various character and at different temperatures. A warm front is formed when warm air rises over the cold air at a relatively gentle slope of 1:100 to 1:400. The precipitation zone extends 300–500 km ahead of the warm front. A cold front is formed when cold air moves under a warm air mass forcing the latter upward. A steeper sloping interface (1:25 to 1:100) is observed. The precipitation zone is limited in this case to about 80 km ahead of the front. The horizontal convergence of air into a low-pressure point results in vertical displacement of air, which may lead to condensation and precipitation. Such meteorological processes commonly occur on or near the tropics as northern and southern components of the trade winds and easterlies. The cold air that commonly prevails over warm oceans in the lower latitudes during the latter part of summer, causes tropical storms during which enormous wet air masses pulled in the lower layers rise in the upper atmosphere. The resulting rains fall mostly near the trajectory of the tropical storm center.

Horizontal convergence may also occur as western and eastern sides of two adjacent low-pressure cyclones meet. Frontal and horizontal convergence commonly generates precipitation of moderate intensity. Convective precipitation is caused by local differential heating of air masses, leading to air instabilities and upward movement of air. Instability showers often occur when cold air moves over a warm surface. Air-mass showers is the name of convective rains that are not associated with a pressure system. These showers commonly have relatively low intensity and small areal coverage.

In many regions, a significant part of precipitation is caused by thunderstorms. These convective storms have high intensity and short duration. Thunderstorms develop in three stages. During the first stage, which lasts 10–15 minutes, cumulus cloud formation is observed. Simultaneously, upward air flows at velocities of up to 60–70 km per hour and a significant horizontal inflow of air into convective cells occur. The vertical air movement may reach heights of 7–8 km. The second stage lasts 15–30 minutes and is characterized by strong lifting air movement at velocities to 110–120 km per hour and high rainfall intensity. At heights of 1.5–2.0 km, descending air movement begins. During the dissipating stage, descending air movement predominates until the convective cells disappear.

Orographic cooling occurs when air masses are forced to rise over an obstruction, like a mountain ridge. The result is condensation and rain on the windward side of the mountain, with contrasting dryness on the lee side of the mountain. The amount of precipitation at the orographic cooling is roughly proportional to the windspeed up the slope and to the amount of moisture in the air. Mountains are not so efficient as cyclonic systems in removing the water from a given air mass, because the rising of the moist air caused by mountains is usually less than in cyclonic systems. However, orography is a constant factor in the cause of precipitation at the same place. Regions with orographic effects exhibit relatively high precipitation accumulation, as well as increased frequency of events (for example, some mountain regions of Mediterranean area, the region of the Cascades in the northwestern United States, and some coastal regions of Japan).

Time-spatial distribution of rainfall, especially of storm rainfall, is important for many hydrological events. Storms generally exhibit one or more centers of maximum depth. The difference between the area-averaged depth and the storm-center value increases with increasing area and decreases with increasing total rainfall depth. For storm rainfall in many regions, stable depth–area–duration relations exist. In many cases, it is also possible to construct the dependencies between rainfall frequency, its duration and its average intensity.

2.2. Snow Cover and Ice

Permanent snow cover is formed on about 20 percent of the Northern hemisphere and about 15 percent of the Southern hemisphere. A significant part of the land is covered by snow several times during the cold period. Changing the heat balance of the land, the snow cover has a considerable effect on the climate. The presence of snow cover on a drainage basin also greatly influences runoff generation. In many parts of the world, river runoff consists mainly of water yielded by the melting of snow. The snowmelt spring runoff of most large plain rivers of Russia and Canada exceeds half of annual runoff; at the same time, the portion of snowmelt runoff from mountain areas in the arid regions can be significantly larger.

Snowfall over an area is more uniform than rainfall, however; snow accumulation is largely a function of elevation, slope, exposure, and vegetative cover. Snow spatial redistribution is strongly affected by the interaction of wind and topography as well as by interaction of wind and vegetation. Gullies and surface depressions are filled up by snow first of all and can accumulate a considerable portion of the total river basin snow resources (in some parts of Russia, the snow in rills and gullies consists of about 30 percent of total river basin snow resources). In forests, much of the intercepted snow is blown off and accumulates on thesoil surface. The snow retention coefficients (the ratios of snow catch in the surface in question to the accumulation in an otherwise virgin soil) vary from 0.4 for open ice surface and 0.9 for arable land, to 1.2 for hilly district and 3.2 for edges of forests. During blowing and transport of snow significant evaporation(sublimation) may occur (the evaporation losses may reach 40–50 percent of annual snowfall). The snow water equivalent (the depth of water which would result from the melting of the snow) in forest areas is usually 10–40 percent more than in the open areas (in some cases, a general increase of precipitation in the forest is possible). Snow accumulation generally increases with elevation because of the combined effect of the prevailing lower temperatures and the increased frequency of precipitation events caused by orographic effects.

The small-scale variations of snow cover, caused by spatial change of terrain, vegetation, and local meteorological conditions, are superimposed on large-scale variations associated with physiographic and climatic zonality. This leads to very large spatial variability of snow cover characteristics, and they are often considered as random values. The coefficients of spatial variation of the snow water equivalent range from 0.15–0.20 in the forest zone to 0.30–0.60 in the steppe zone. To describe spatial variability of the snow water equivalent one commonly applies the lognormal or gamma statistical distributions.

After snowfall, the snowpack undergoes essential transformation (metamorphosis) caused by compaction, action of the thermal gradients, and change in the crystal structure resulting from interactions of ice, liquid water, and water vapor. Because of migration of water vapor and the freezing together of the small particles of ice, the average ice particle size increases and to the end of winter a snowpack commonly consists of uniform coarse crystals (the process of the formation of coarse snow crystals is called riping). The metamorphosis of snow produced a significant change of density and other physical properties of snow. Snow at the time of fall may have a density as low as 0.01 to as high as 0.15 gcm^{-3}; snowfall in the form of dry snow may vary in density between 0.07 and 0.15 gcm^{-3}; average wind-toughened snow has a density about 0.28–0.30 gcm^{-3}. Ripe snow has a uniform density of 0.4–0.5 gcm^{-3}. The greatest density that can be attained by shifting the snow grains around is about 0.55 gcm^{-3}. Further densification, which can occur under action of deformation, refreezing, and recrystallization, produces a compact, dense material called firn. At a density of between 0.82 and 0.84 gcm^{-3}, the air spaces disappear and the material becomes impermeable to air and water. . This material can be defined as ice. The old ice has a density about 0.90 gcm^{-3}; the theoretical density of pure ice is 0.92 gcm^{-3}Accumulation on land of ice resulting from recrystallization of snow or other forms of precipitation leads to the formation of glaciers. Typical peculiarities of the glacier are the presence of an area where snow or ice accumulates in excess of melting, and another area where the wastage of snow or ice exceeds the accumulation, as well as a slow transfer of mass from the first area to the second. Glaciers exist in a wide variety of forms. They range in size from ice masses occupying tens of square meters to the great continental ice sheets of Antarctica and Greenland. On the Earth's land there are 140 glaciers with areas of more than 1000 square km; at the same time, only on the area of the past Soviet Union are there about 30 000 glaciers of size less 0.1 square km.

A significant amount of ice can accumulate in the ground. If the climate is very cold, a layer of frozen ground may be formed which persists from year to year. The surface layer of this ground (the active layer) normally thaws during the summer and refreezes during the winter, but the ground below remains frozen and impermeable. Such ground is called permafrost and occupies about a quarter of the Earth's land. In areas mantled with peat or a dense mat of living vegetation, the active layer is generally thin and permafrost occurs close to the surface. In areas of bare gravel or exposed bedrock, the active layer may be quite thick. Permafrost is more widely and continuously distributed in lowlands than it is in the mountains in spite of lower temperatures prevailing in the mountains. Lakes, reservoirs, and large ponds produce a warming effect on the ground increasing the depths where lies permafrost.

The water frozen on the land surface and in the ground may form icings which cover considerable areas. In the northeast part of Russia, icings occupy 7–10 percent of area and accumulate 200–300 mm of water.

Being a porous medium, the snowpack has much in common with the soil. In the dry snow, liquid water retains mostly by film tension and capillary forces. The porosity of snow varies from 0.80–0.87 (for new snow) to 0.50–0.70 (for old coarse-grained snow). The liquid water-holding capability of snow (the maximum value of liquid water content beyond which water will drain by gravity action) is about 0.13–0.15. The movement of water through the snowpack begins when the snowpack is saturated by liquid water more than to these values. In the period of snowmelt, a part of the liquid water may refreeze.

The rate of snowpack melt is determined by the incoming heat. The energy budget of the snowpack includes: the net shortwave and longwave radiation; the turbulent exchange of heat in the atmospheric layer above the snow surface (sensible heat); the latent heat consumed in evaporation and sublimation; the heat delivered to snowpack by precipitation; the heat exchange at the land surface; and the change in heat storage including the heat released by freezing of liquid water content. The net shortwave radiation is the most dominent energy component during snowmelt. In the process of metamorphosis and riping, the snowpack decreases its reflected capability (albedo) and absorbs the most part of shortwave radiation during snowmelt. The new snow has the albedo 0.75–0.90, and after riping the albedo can reach 0.35–0.40. The empirical dependence can usually be constructed between the albedo and the snow density as a characteristic of snow riping. A close relationship commonly also exists between albedo and the accumulated daily maximum temperature after the last snowfall. The sensible heat is the second important energy budget member. Sometimes, the precipitation heat can be a considerable contribution to positive snowpack energy balance. However, in most cases effects of rainfall on the riping snow and a decrease of albedo are more important.

The most simple and informative index of the snowmelt rate is the air temperature. The relation between these values can be presented as

$$M = a(T_a - T_b) \tag{1}$$

where M is the snowmelt in millimeters per day, T_a is the air temperature in degrees Centigrade, T_b is an air temperature below which no melt occurs (it is commonly 0–2°C), and a is an empirical coefficient (degree-day factor) which can be interpreted as the snowmelt per day at change of air temperature per degree. The degree-day factor varies depending on climatic and physiographic conditions, but in many cases variations are possible to classify according to the latitude, topography, and vegetation. Because forest cover has a significant effect on many of the variables affecting snow cover energy exchange, there is an essential difference in degree-day factors for forest and open areas. The typical degree-day factors for mid-latitude open areas are usually 4–5mm/day °C; for deciduous forest the figure is 3–4 mm/day °C; for dense coniferous forest 1.5–2.0 mm/day °C.

Differences in aspect are also important. At open mountain areas the degree-day factors reach 5–6 mm/day °C. Melt factors in Arctic areas tend to be smaller than those at lower latitudes with similar physiographic conditions, mainly due to lower radiation intensities and relatively little wind during the melt season. Windy areas typically have higher melt factors than areas where calm conditions prevail. In many cases, the degree-day factors increase during the progress of snowmelt as a result of the decrease of the snow albedo, soil warming, and increasing solar radiation. For example, the degree-day factor averaged for Finland is 1.45 mm/day °C at the beginning of the snowmelt period, and 4.75 mm/day °C at the end of the snowmelt period. The maximum values of the degree-day factor reach 80–90 mm/day °C.

The main difference between the melting of snow and ice results from the low albedo of ice. Typical mid-latitude degree-day factors for ice melting are 5–10 mm/day °C. In investigating mass balance of the glaciers, it is more suitable to measure the ablation (that refers to all processes by which solid material is removed from the glacier) instead of the melt. Because evaporation from the glacier surface is small, in temperate climates the values of the glacier melt and of the ablation are close. In some high Arctic regions, appreciable snow and ice are removed by wind erosion. Most ablation occurs on the surface of a glacier. During the ablation season the surface level of a glacier drops, not entirely due to ablation, but partly due to compacting, or densification, of the snow layers beneath. Thus, in order to measure ablation, one must measure the thickness and the density of a surface snow layer at each time of measurement.

2.3. Interception and Depression Storage

Before reaching the land surface, a part of the precipitation may be intercepted by vegetation and/or other types of surface cover. A portion of intercepted rainfall evaporates and the other portion may flow down on vegetation stems. Rainfall interception varies with species composition, age, and density of vegetation cover. A dense conifer stand usually intercepts to 25-–30 percent of the rainfall at the stem flow of 5–7 percent. The net rainfall interception by hardwood stands is about 15 percent for the period with leaves and about 7 percent for the period without leaves. According to the detailed measurements carried out in the Central Amazonia, the net interception in the tropical rainforest is approximately 10 percent of rainfall. The rainfall interception losses for dense grasses and herbs is as great as for deciduous trees. Interception can be also be significant in large urban areas. The urban landscape includes flat rooftops, potholes, parking lots, cracks, and other rough surfaces that can intercept and hold a significant amount of water.

Interception of snowfall by vegetation may lead to direct sublimation of snow and significant redistribution of snow by wind. The interception of falling snow by conifer forests often reaches 30–35 percent. Snow interception in hardwood stands is about 7–10 percent.

To calculate net rainfall reaching the land surface through the canopy, the Rutter model of interception is often used. In this model the canopy is considered to have a surface storage of capacity S_c, which is filled by rainfall P, and emptied by evaporation and drainage Q. This capacity may be interpreted as the minimum depth of water required to wet all canopy surfaces. When the depth of water C on the canopy equals or exceeds S_c, the evaporation from the canopy is assumed to occur at the rate E_p. When C is less than S_c, the rate is assumed to be $E_p C S_c^{-1}$. The rate of change of storage is then calculated as:

$$\frac{\partial C}{\partial t} = Q - ke^{b(C-S_C)} \qquad (2)$$

where

$Q = a(P - E_p C/S_c)$ when $C < S_c$

$Q = a(P - E_p)$ when $C > S_c$

k, a, and b are empirical constants which depend on the type of vegetation; t is time.

The value of storage of capacity S_c can change over time, depending on seasonal growth of the vegetation cover.

The water reaching the land surface begins to fill up first of allthe depressions of land surface, and simultaneously moves vertically down under action of gravitational and soil suction forces. The sizes and depths of these depressions vary within a large range depending on relief, slope, vegetation, human activity (especially land use and land treatment). The portion of precipitation trapped in the surface depressions can be crudely evaluated as the runoff losses minus the infiltration during the largest rainfall. On the moderate and gentle uncultivated slopes, the initial runoff losses on the filling up of the surface depressions are, on the average, 1.0–1.5cm; however, some kinds of land treatment can lead to almost full cease of the runoff. In some regions (for example, the tundra), the surface depressions and closed areas can occupy a significant part of river basins. Assuming the portion of area with a depth of depressions exponentially decreases with growth of this depth, we can express the total volume of water stored in depressions as

$$S = Sm[1 - \exp(-R / S_m)] \qquad (3)$$

where R is the accumulated excess of rainfall over the infiltration and S_m is the maximum depression storage capacity of the drainage basin, which depends on the physiographic conditions andlanduse.

2.4. Infiltration of Water into the Soil and Vertical Movement of Soil Moisture

Infiltration is the flow of water through the soil surface. The rate and volume of the infiltration depend on the conditions on the soil surface, soil properties (texture, structure, and chemical peculiarities), and soil moisture content.

At the bare soil, a surface crust may develop under action of raindrops. The impact of raindrops breaks down soil crumbs and aggregates, and the particles of silt and clay penetrate previously existing pores, clogging them, and greatly reducing infiltration. The vegetation protects the soil from rainfall action and increases entrance permeability resulting from root activity and increment of the organic content. Root systems perforate the soil, keeping it unconsolidated and porous. The organic matter promotes a crumb structure and improves permeability magnitude. During short high-rate rainstorms, most of the rain quickly travels through macropores to the lower layers of the soil, and only a small fraction of the rain is absorbed by the soil matrix. During low-rate storms, a greater fraction of the rain is adsorbed by soil matrix and soil swells reducing the width of the macropores. The swelling soil after drying may also form a surface crust.

Soil texture is determined by the size distribution of individual particles in the soil (the percentages of clay, silt, sand, and coarse fragments more than 2 mm). Soil structure depends on morphological properties of soil particles, and clay, silt, and sand types. It is characterized by bulk density, pore-sized distribution, and construction of vertical profile of soil. The pore sizes and pore-size distribution are greatly affected by the content of soil organic matter, which determines both the sizes of soil aggregates and their stability in water Soil texture and structure are closely related to soil porosity and capillary suction forces. Natural cracks, worm holes, or tillage marks create soil macroporosity. The increase of soil porosity leads to the increase of soil permeability, but also to the decrease of capillary suction. Chemical properties of the soil affect the integrity and stability of the soil aggregates, processes of colloidal swelling of the soil, and the suction pressure of the soil matrix.

Water may exist in the soil as liquid water, vapor, or ice. A part of liquid water (hygroscopic and capillary water) is held by molecular forces of the soil matrix. Hygroscopic water exists in the thin films around soil particles at negative (suction) pressures of 31 to 10 000 bars, and may freeze at temperatures below 0°C. Capillary water is held at a negative pressure of 0.33 to 31 bars, filling gaps between the particles. As the soil moisture increases, the gravitational forces become strong enough to counteract the negative pressures (this occurs at pressures between 0 and 0.33 bars). The maximum amount of water which soil can hold against gravity is called field capacity. Water in excess of field capacity percolates down the soil column, ultimately reaching the soil layer with a small permeability where this movement stops. The plant root system can extract the water if the negative soil moisture pressures are less 15 bars (the content of soil moisture at this pressure is called the permanent wilting point).

Soil water content has a significant effect on the physical and chemical characteristics of the soil and conditions of water flow in soil pores. Soils with an appreciable amount of silt or clay are subject during wetting to the disintegration of the crumbs or aggregates, which in their dry state may provide relatively large pores. These soils also normally contain more or less colloidal material, which in most cases swells appreciably when wet. The pores of sands are relatively stable.

When the all pores of the soil are saturated by water, the water flux depends mainly on gravitational forces. The dynamics of flow in the saturated soil can be described by Darcy's law. This law states that the velocity of flow through a porous medium is directly proportional to the gradient of the piezometric head $h = p\rho g + z$, where p is the pressure head, ρ is the density of water, g is gravitational acceleration, and z is the elevation of the point under consideration from an arbitrary datum. If q is the vertical velocity of soil moisture movement, then Darcy's law can be expressed as

$$q = -K\frac{dh}{dz} \qquad (4)$$

where K is the coefficient which is called the hydraulic conductivity saturated soil (or the coefficient of filtration).

The dynamics of flow in the unsaturated soil for isotermical conditions at some additional unessential assumptions can be described by the relation, analogous to Darcy's law

$$q = -K(\theta)\frac{dh}{dz} \qquad (5)$$

where, the pressure head $h = \psi(\theta) + z$ includes the capillary suction $\psi(\theta)$ that is as well as the proportionality coefficient $K(\theta)$ (the hydraulic conductivity of unsaturated soil) a function of soil moisture, θ is the ratio of the volume occupied by liquid water to the total volume of soil pores (volumetric soil moisture content). Substitution of the last relation into the mass-conservation equation results in a dependence that is called the Richards equation

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z}\left[D(\theta)\,\frac{\partial \theta}{\partial z}\right] - \frac{\partial K}{\partial z} - r(z,t) \qquad (6)$$

where $D(\theta) = K(\theta)\,\partial\psi/\partial\theta$ is called the diffusivity of soil moisture, and r isthe uptake of water by plant roots.

A lot of different empirical relationships which relate the hydraulic conductivity of unsaturated soil, the diffusivity of soil moisture to the volumetric soil moisture content, and the commonly measured soil moisture constants (porosity, hydraulic conductivity of saturated soil, field capacity, permanent wilting point, and others) has been developed. The uptake of water by plant roots is a complicated biological process, however, which in many cases can be represented as a empirical function of difference of soil and root capillary–osmotic suction, the hydraulic conductivity of unsaturated soil, and the root density.

The moisture movement in unsaturated soil often shows a clear hysteresis effect: the relationships between the soil matrix suction and the soil moisture, as well as between the hydraulic conductivity and the soil moisture, are not the same during wetting and drying events. Capillary forces speed up the filling of small pores during wetting, but delay their emptying during drying. Moreover, the tortuosity of channels, where the water fluxes occur, essentially depends on the previous history of soil moisture conditions.

Freezing of the soil decreases its porosity and capillary forces. If the soil is frozen while saturated, it may become completely impermeable. The hydraulic conductivity of dry soil commonly changes insignificantly after freezing; however, in cold periods a considerable variation of the soil moisture may occur under action of the temperature gradients. There is often a significant flux of soil moisture from the unfrozen zone to the front of freezing. This flux may lead to swelling upper layers of soil and to decreasing soil permeability. In the snowmelt period, the hydraulic conductivity of the dry frozen soil may also decrease as a result of freezing the melt water inside of the soil matrix, where the melt water mixes with the overcooled hygroscopic and capillary water.

The infiltration of melt water into frozen soil can stop because of formation of a impermeable layer at a soil depth. In order to compute heat and moisture transfer in frozen and thawed soil, the following system of equations can be used:

$$\frac{\partial\theta}{\partial t} + \frac{\rho_i}{\rho_w}\frac{\partial I}{\partial t} = \frac{\partial}{\partial z}\left(K\frac{\partial\psi}{\partial z} - K\right) \qquad (7)$$

$$C_{ef}\,\frac{\partial T}{\partial t} = \frac{\partial}{\partial z}\left(\lambda\frac{\partial T}{\partial z}\right) + \rho_w C_w\left(K\frac{\partial\psi}{\partial z} - K\right)\frac{\partial T}{\partial z} + \rho_i L\frac{\partial I}{\partial t} \qquad (8)$$

where

$$C_{ef} = \rho_w C_w\theta + \rho_i C_i I + \rho_g C_g (1 - P_s) \qquad (9)$$

θ and I are the volumetric contents of liquid water and ice in the soil, respectively; T is the temperature of the soil, P_s is the soil porosity, K is the soil hydraulic conductivity, L is the latent heat of fusion of ice, and λ is the soil thermal conductivity; ρ denotes density, C is heat capacity, w, I, and g are indices of water, ice, and soil matrix. To take into account the change in soil capillary pressure and the hydraulic conductivity of frozen soil, different empirical formulae can be used.

Depending on the soil moisture conditions and soil–rock properties, the subsurface space where the infiltrated water is stored may be divided vertically into two zones based on the relative proportion of pore space that is occupied by water: an unsaturated zone, or aeration zone, in which the pores contain gases (chiefly air and water vapor); and a saturated zone in which all soil pores are filled with water. During recharge periods, water moves under the force of gravity downward through the unsaturated zone to the saturated zone. The upper limit of the saturated zone (the water table) is the depth at which the water is at atmospheric pressure. The unsaturated zone is often divided into the soil water zone, extending downward from the land surface as far as plant roots penetrate; the capillary fringe, where water rises by capillary forces above the saturated layers; and the intermediate zone, where downward percolation presumably occurs, at least intermittently, toward the saturated zone. The depth of the soil water zone is variable and dependent on soil type and vegetation; the capillary fringe may be from practically zero in coarse material to tenths of meters for fine clays; the intermediate zone may be hundreds of meters thick or be completely absent. Water in the capillary fringe exists at pressures less than atmospheric pressure. All pores may be saturated near the base of this capillary fringe, and the number of pore spaces that are filled with water decreases in an upward direction. In areas of shallow water table, the capillary fringe may extend upward to the root zone or plants and even to the land surface, thus permitting discharge of water by evaporation.

2.5. Evapotranspiration

The evaporation of the water into the atmosphere begins when the temperature of the evaporating surface is such that some molecules of the liquid water have attained enough kinetic energy to eject from the water or the land, and to penetrate into the air forming water vapor. Some of these molecules may return from the air and condense, but the number of escaping molecules will be larger than the returning ones until the number of molecules in the air reaches a value which is the maximum possible amount for a given temperature of the air and the water vapor becomes saturated. If more molecules enter the surface than leave it, condensation occurs. The motion of the molecules escaping from the evaporating surface and returning from the air produces a difference of pressure which is determined by the rate of evaporation. The losses of kinetic energy needed for transfer of liquid water into vapor lowers the temperature of evaporating surface (the latent heat of vaporization of water at 0°C is 596 calg^{-1}). To support evaporation, a steady income of energy has to occur. An energy balance for a given area, particularly for a defined water body can be written as

$$Q_e = Q_s - Q_r + Q_a - Q_{ar} - Q_{bs} + Q_v - Q_h - Q_c \tag{10}$$

where Q_e is the energy used in evaporation (latent heat), Q_s is the incident solar radiation, Q_r is the reflected solar radiation, Q_a is the incoming longwave atmospheric radiation, Q_{ar} is the reflected longwave radiation, Q_{bs} is the longwave radiation emitted by evaporating body because of its temperature, Q_v is the net energy advected by moving water, Q_h is the heat removed from the system into the air as sensible heat, and Q_c is the change in system energy. The main source of such an inflow of energy during evaporation of water into the atmosphere is solar radiation and the heat of the atmosphere. Evaporation can be estimated by direct or indirect measurement. Direct methods are mostly dominated by point sampling or integrated measurements over small areas, mostly with evaporation pans or lysimeters (vessels or containers placed below the land surface to intercept and collect water moving downward through soil). Indirectly, evaporation can be measured by performing a water balance of a given area.

The rate of evaporation from the water surface into the atmosphere depends on the difference between the pressure of saturated vapor at the temperature of the water and the vapor pressure of the air above the water surface. The last value is determined by the content of the water vapor in the air and the air temperature, which depend in turn on the atmospheric circulation and the turbulent transport in the atmospheric boundary layer. Experimental research has shown that the rate of evaporation from water bodies is not related to the size of the areas of these bodies, if these areas are less than approximately 20–30 square km. The rate from the larger areas slowly decreases with the growth of water body areas, reaching the values of the evaporation rate from the seas and oceans. During the warm period of the year, the rate of evaporation from the water bodies does not necessarily depend on the water body depth if it is more 2–3 meters (there is a insignificant growth of the evaporation rate to these depths). During the cooling of the surface water in the fall and the heating the water body in the spring, the role of the water body depth in evaporation increases. In deep water, the relatively low temperature of the surface water in the spring (less than 4°C, the temperature of maximum water density) will decrease evaporation, while the higher temperature in the fall leads to an increase in evaporation.

The rate of evaporation from water surface E_w can be calculated as

$$E_w = C_e\, V_r\, \rho_a (q_s^* - q_a) \qquad (11)$$

where C_e is a bulk coefficient for water vapor, V_r is the mean wind speed at some reference level z_r above the water surface, ρ_a is the density of the air, q_a is the mean specific humanity at a reference level z_a, and q_s^* is the saturation specific humidity at the mean temperature at the water surface. The value of C_e can, in principle, be calculated on the basis of the characteristics of the atmospheric boundary layer, but given the unavailability of data this coefficient is possible to find more exactly by fitting.

Because the water surface temperature is often unknown, to calculate the rate of evaporation the Penman combined mass-transfer and energy balance method is frequently applied. This requires information only on wind speed, temperature, humidity at one level, and the available energy flux at the water surface. In this method

$$E_w = \frac{\Delta}{L_e(\Delta+\gamma)} Q_{ne} + \frac{\gamma}{\Delta+\gamma} C_e V_r \rho_a (q_a^* - q_a) \qquad (12)$$

where $\Delta = dq^*/dT$ at the temperature of the air T_a, q_a^* is the air saturation specific humidity at the same temperature, $\gamma = c_p/L_e$ is the psychrometric constant, c_p is the specific heat of air at constant pressure, L_e is the latent heat of vaporization, and Q_{ne} is the net radiation minus the heat flux to the ground.

The rate of evaporation from the land is determined not only by meteorological conditions but also by the amount and rate of water supply to the evaporating surface. Further, water molecules have to overcome greater resistance to escape from a surface of soil or plant than from a free-water surface. Evaporation of water by plants (this process is called by transpiration) occurs mainly through intercellular openings in the leaves (stomata), which open to allow in the carbon dioxide necessary for photosynthesis and respiration.

To reach the surface of the soil and plant cover, soil moisture must move from the lower depths to the surface. If evaporation from the land is to be a continuous process, movement will have to take place through considerable distances in unsaturated soils. The movement of water upward to the evaporating surface in the soil occurs under action of molecular forces of the soil matrix. In the case of enough dry soil, the movement of water vapor under action of the soil temperature gradients may also play an essential role. The wet soil often maintains a practically constant rate of evaporation at a certain range of moisture content, until alow moisture content (approximately the permanent wilting point) is reached. According to some experimental investigations, evaporation from the soil surface continues as long as the upper surface layer—about 10 cm for clays and about 20 cm for sands—remains moist.

The movement of the water to the stomata is caused by stomatal capillary suction and osmotic pressure resulting from the difference in moisture between sap at the roots and surrounding soil. The volume of water evaporation by plants is much larger than the volume of consumption in the formation of vegetative material, and water coming from the roots reaches the stomata almost entirely. Transpiration is a complicated process, depending on both biological and environmental factors. The most important biological factors are type, stage, and growth of the plants, leaf and root structure, and density and behavior of stomata. If the water supply to the leaves is greater than the evaporative capacity of the atmosphere, transpiration is at its climate-controlled potential rate. A plant may help the transpiration process by root development, change soil moisture gradients, and regulate stomatal openings.

In many cases, vegetation cover may also affect the temperature and the moisture content of the atmospheric boundary layer. The link between transpiration, photosynthesis, and the exchange of carbon dioxide makes the transpiration an important factor in long-term interactions of vegetation and climate. Many different attempts at f modeling the processes of energy and water transfer in the soil–vegetation–atmosphere system have shown that the construction of such models require more than 50 parameters reflecting the local soil, vegetation, and atmosphere conditions. To determine most these parameters, it is necessary to carry out special measurements that are only available at the present time for small areas.

Analogously to (10), the transpiration rate in the simplified form can be presented as

$$E_T = \rho_a \cdot A_f [q_f^*(T_f) - q_{ac}]/r_{af} \quad (13)$$

where A_f is the ratio between the area of leaves and the land ares shadowed by leaves (the leaf area index), q_f^* is the saturated special humidity of air in stomatal openings at the temperature T_f, q_{aC} is the special humidity in the canopy air space, and r_{af} is a coefficient which characterizises the aerodynamic and stomatal resistance to water transfer.

The sum total of evaporation and transpiration is usually called evapotranspiration. When the vegetation cover is dense, the transpiration commonly is larger than the evaporation from the soil. According to experimental data collected in the central part of Russia, transpiration contributes 45 percent of evapotranspiration in conifer forest, and 50 percent of evapotranspiration. in deciduous forest, while evaporation from soil is only 30 percent of evapotranspiration in conifer forest, and 35 percent in deciduous forest (about 25 percent of precipitation in conifer forest and 15 percent in deciduous forest is evaporated as a result of interception). In the Amazonian rain forest, on the average, 50 percent of the incoming rainfall is reevaporated, about 25 percent through the interception process and almost the all remainder by transpiration. Transpiration is the predominant cause of losses of soil moisture in arid and semi-arid regions.

It is often necessary to differ the actual evapotranspiration and the potential evapotranspiration (the climatically controlled rate of evapotranspiration that occurs when the supply of water to the land surface and water resource in the root zone are unlimited). The potential evapotranspiration is more easily measured and calculated (approximately, it can be estimated as evaporation from the shallow water surface). To calculate the rate of actual evapotranspiration ET_a, it is possible to use the relation

$$ET_a = \beta \, ET_p \quad (14)$$

where ETp is a potential evapotranspiration rate and β is a reduction or moisture availability factor. In many cases, it is possible to take

$\beta = 1$ for $w > w_0$

$\beta = w/w_0$ for $w < w_0$

where w and w_0 are the soil moisture contents in the 1 m soil layer.

The most known model used for estimation of actual evapotranspiration is the Penman–Monteith equation:

$$ET_a = \frac{\Delta}{L_e\left[\Delta + \gamma\left(1 + r_c / r_a\right)\right]} Q_{ne} + \frac{\rho_a \gamma (q_a^* - q_a)}{r_a\left[\Delta + \gamma\left(1 + r_c / r_a\right)\right]} \qquad (15)$$

where r_a is the aerodynamic resistance to water vapor transport, and r_c is the stomatal resistance to water transport in dry conditions. Coefficient r_a is determined as functions of the roughness of land and vegetation cover as well as the characteristics of the atmospheric boundary layer. Coefficient r_c varies as a function of soil moisture as well as vegetation type and equals to zero for a wet canopy.

Evaporation from snow requires a three-phase change of state from a solid to liquid to gas (this process is called sublimation). The latent heat of sublimation is much higher than the latent heat of melting (80cal g^{-1}), so that the latter is a preferred process. In order for evaporation to occur the saturated vapor pressures at the snowmelt temperatures must be low enough, and the air above the snow surface must be sufficiently dry. If there is a favorable vapor pressure gradient, evaporation of snow may occur even at absence of heat income. The necessary heat may be taken from the snow itself by cooling it. On the average, the evaporation rate from snow cover is approximately 0.3 mm per day before snowmelt, and 0.4–0.5 mm per day after the beginning of snowmelt. In spite of the fact that the daily rates of evaporation from snow cover seldom reach more than 1–2 mm, the accumulated evaporation during sunny and dry winter–spring periods may significantly decrease the snow resources before the snow melt (for example, in the forest zone of Russia the evaporation from snow cover leads to decreases of the snow water equivalent before the spring snowmelt by 15–20 percent).

The role of condensation on land surface in the hydrological cycle has been investigated insufficiently. However, it is known that in some mountain regions condensation of liquid water from fogs in forest may increase the annual precipitation up 7–10 percent. In arid areas, night condensation of air moisture may essentially increase moisture content in the upper layer of the soil.

2.6. Groundwater and Groundwater Flow

Water in the saturated zone of soil–rock systems is commonly called groundwater, and it represents the largest liquid water store of the terrestrial hydrological cycle. Water may penetrate into soil–rock systems in the process of vertical movement from the unsaturated zone; as a result of filtration from river channels, lakes, and reservoirs; and also as a consequence of artificial recharge Groundwater reservoirs in permeable geological formations that can release a considerable amount of water with relative ease are called aquifers.

If, after drilling a fully-penetrating well through a geological formation, the groundwater rises to the piezometric level (which is equal to the elevation above a datum plus the pressure in the aquifer), this formation is called a confined, or artesian, aquifer. An unconfined (phreatic) aquifer has a free water surface. This free water surface may be directly connected to a stream or other surface waters. The water in phreatic aquifers comes from direct rainfall recharge over the aquifer, from connections to surface waters, and/or from other aquifers. The confining beds separating the aquifers may be completely impermeable (aquifuge), or “leaky” (aquiclude). Whether a rock or soil formation is an aquifer, aquifuge, or aquiclude depends largely on its geologic origins and history.

Confined aquifers recharge through areas where the soil system is exposed to the surface, or through aquicludes. Many confined aquifers contain “fossil waters” deposited in past geologic times. The best aquifers are generally sediment deposits of alluvial or glacial origin. Some sandstone and sedimentary rocks may have very little permeability through pore openings (for example, dolomite and limestone), and their water-bearing capacity and transmission depend mainly on the degree of fracturing resultant from weathering, and the degree of solution of cementing material. The formation of fractures, crevices, or caves in highly-weathered and dissolved limestone (karst processes) often leads to development of underground river systems.

Movement of groundwater occurs under action of hydrostatic head with velocities ranged from several meters per day to only several meters per year. The change of moving water mass in a confined aquifer can only be attributed to changes of porosity caused by compression of the soil–rock matrix. Substanding Darcy’s law for the groundwater fluxes in all cartesian coordinates x, y, and z in the mass balance expression, and taking into account the change of water storage in the pores, we receive

$$\frac{\partial}{\partial x}\left(K_x\frac{\partial h}{\partial x}\right)+\frac{\partial}{\partial y}\left(K_y\frac{\partial h}{\partial y}\right)+\frac{\partial}{\partial z}\left(K_z\frac{\partial h}{\partial z}\right)=S_0\frac{\partial h}{\partial t} \qquad (16)$$

where K_x, K_y, and K_z are saturated hydraulic conductivities in the x, y, and z directions respectively, and S_0 is the specific storativity (the volume of water released from storage per unit volume of aquifer per unit change in pressure head), which has units of inverse distance and is usually taken as a fitting parameter.

Under steady-state conditions, equation (16) becomes

$$\frac{\partial}{\partial x}\left(K_x\frac{\partial h}{\partial x}\right)+\frac{\partial}{\partial y}\left(K_y\frac{\partial h}{\partial y}\right)+\frac{\partial}{\partial z}\left(K_z\frac{\partial h}{\partial z}\right)=0 \qquad (17)$$

For an isotropic and homogeneous medium, the hydraulic conductivities can be taken out of the derivatives and divided out; thus equation (17) is transformed into the well-known Laplace equation:

$$\frac{\partial^2 h}{\partial x^2}+\frac{\partial^2 h}{\partial y^2}+\frac{\partial^2 h}{\partial z^2}=0 \qquad (18)$$

Description of the flow in an unconfined aquifer is complicated by the presence of a free surface which changes in time and in plane, and may include a recharge of the water from the unsaturated zone If it is possible to assume that there is only horizontal flow, and that the slope of the phreatic surface is small in comparison to total aquifer depth, for this aim the Boussinesq equation is commonly applied:

$$S_0 \frac{\partial h}{\partial t} = \frac{\partial}{\partial x}\left(K_x H \frac{\partial h}{\partial x} \right) + \frac{\partial}{\partial y}\left(K_y H \frac{\partial h}{\partial y} \right) + R_g \qquad (19)$$

where *h(t)* is the phreatic surface level, *H(t)* is the saturated thickness, and R_g is the instantaneous vertical recharge into the saturated zone.
If the groundwater is located close to the land surface, it can intensively interact with the surface water. The rise of groundwater level to the land surface may lead to a sharp increase of overland and subsurface flow. Groundwater discharges on the river slopes or in the basin depressions form springs and creeks. Where a river channel is in contact with an unconfined aquifer, groundwater may flow from the aquifer into the river channel, or vice versa, depending upon where the water level is lower. During a flood period, groundwater levels may be significantly raised near a channel by inflow from rivers. This process is known as bank storage. The reduction of the maximum discharge during floods caused by bank storage can reach 10–15 percent. After the rise of river stage during flood, a long period of groundwater recession may be observed. In many cases, pumping of groundwater leads to a decrease of surface runoff. If the groundwater located deep enough, recharge of the unconfined aquifer may occur from the river drainage network without hydraulic interaction. Such a type of recharge is often observed in dry regions or on permafrost.

Groundwater discharging into a river system forms the base runoff that is the main sustainable portion of total runoff for many plain rivers. The contribution of groundwater to the total river runoff may vary from a negligible fraction (for instance, for mountainous rivers) to 100 percent (for some karst river basins); however, there is a clearly-expressed physiographic zonality in the distribution of this contribution. For example, in the northern regions of the European part of Russia, where the water table is shallow and river drainage is not well-developed, the portion of the groundwater runoff is 10–30 percent; in the middle part of Russia, where there is shallow groundwater and well-developed river drainage, the portion of the ground runoff reaches 40–50 percent; and in the southern part, where the groundwater is deep, this portion is 15–30 percent. The portion of groundwater which discharges directly into large lakes, seas, and oceans is about 2 percent; however, in some regions there is a significant submarine intrusion of seawater into coastal aquifers.

2.7. River Runoff and Generation Mechanisms

River runoff is that part of the precipitation which is collected from a drainage basin or watershed and flows into the river system. From the hydrologic point of view, the runoff from a drainage basin may be considered as a product of the hydrologic cycle and a result of a compound interaction of meteorological and physiographic factors. Physiographic factors can be classified into two main groups: basin factors (size, shape, and topography of drainage area, geology, properties of soils, presence of lakes and swamps, vegetation cover, and land use); and channel factors (slope, hydraulic properties of the channels, channel storage capacity, sediments, and stream bed material). Frequently two basins of nearly the same size may behave entirely differently in runoff phenomena. The essential differences occur between large and small basins. For example, most large basins have significant channel storage effects that smooth the variations of water inflow caused by meteorological factors or change of conditions on the basin area. Small basins are very sensitive both to climatic factors and change in land use.

The variety of runoff generation conditions is reflected in the temporal-spatial change of runoff coefficients (runoff–precipitation ratios). Depending on meteorological and physiographic conditions, these coefficients may vary from 0 to almost 1. In the deserts of the tropical and temperate zones almost all precipitation evaporates. Small runoff coefficients (0.05–0.15) are also typical for the steppe and dry savanna zones. In the zone of hard-leaf forests, the runoff coefficients are of the order of 0.1–0.2. However, in the zone of permanently-humid forests, the runoff coefficients reach 0.40–0.45. High runoff coefficients are characteristic also for the tundra and rainforest zones (0.5–0.6). The runoff coefficients from glaciers are usually close to 0.8–0.9.

Runoff commonly shows a well-expressed seasonal variability. Runoff of a typical river basin in the temperate climate region has one or several periods with a significant rise in runoff discharges (such rises are usually called floods), and one or several periods of low flow. In the humid and tropic climates, seasonal variability is comparatively less; in arid regions, there are ephemeral rivers where the runoff is nonexistent during periods without precipitation, although it may appear at different times.

The variability of runoff can be estimated in terms of the day-to-day fluctuation of the river discharges or stages. A graph showing river discharge with respect to time is known as a hydrograph. Hydrographs can be regarded as integral expressions of the physiographic and climatic characteristics that govern the relations between water inflow and runoff of a particular drainage basin. The shape of a hydrograph reflects the difference in runoff components and their paths of movement. A typical, single-peaked, simple flood hydrograph consists of three parts: the approach segment; the rising (or concentration) segment; and the recession (falling or lowering) segment. The lower portion of the recession segment is a groundwater recession (or depletion) curve, which shows the decreasing rate of groundwater inflow. The peak of a rainfall flood hydrograph represents the highest concentration of the runoff from a drainage basin. It occurs usually at a certain time after the rain has ended, and this time depends on the size and the shape of the drainage basin as well as the spatial distribution of the rainfall. The multiple peaks of a hydrograph may occur in any basin as the result of multiple storms developing close to each other. If a hydrograph shows double or triple peaks fairly regularly, this may be due to nonsynchronization of the runoff contributions from several tributaries to the main stream. The recession segment represents withdrawal of water from storage after all inflow to the channel has ceased.

According to the area of genesis, river runoff components can be divided into surface runoff, subsurface runoff, and groundwater runoff. The surface runoff is that part of runoff which is produced on the land surface and flows over the land surface and through river drainage system to reach the basin outlet. The part of the surface runoff which does not reach stream channels is called overland flow. The subsurface runoff, also known as interflow, is that part of the precipitation which infiltrates into the soil and moves horizontally through the soil and the ground above the main groundwater level. A part of the subsurface runoff may enter the stream quickly, while the remaining part may take a long time before appearing in the stream channels. The groundwater runoff is that part of the groundwater which discharges in the river drainage system.

The proportions of surface, subsurface, and groundwater components in the total runoff strongly vary in space and time and are defined by the physical mechanisms of river runoff generation. Field research of runoff genesis on experimental and representative river basins can commonly only provide data which is sufficient for discovering the main features of these mechanisms for small plots. At the same river basin, several distinct runoff generation mechanisms may exist. To establish the leading runoff generation mechanisms on large basins, it is usually necessary to use long series of meteorological data and runoff measurements together.

It is easy to establish from a simple analysis of flood hydrographs that the river runoff includes three components which have differences in timing: 1) quick flow, consisting of water which reaches the river channel network promptly after rainfall or snowmelt and has velocities of several centimeters per second; 2) flow, consisting of water which reaches the river channel network at velocities of order 0.1–1.0 centimeters per second^{-1}; and 3) slow flow, where velocities can be several orders less than the velocities of quick flow. It is commonly assumed that quick flow is mainly overland flow and the slow flow is mainly ground flow. Hypotheses on the paths of the second mentioned component of runoff may be very different, but in most cases, taking into account the velocities, it is possible to determine that it is dominantly subsurface flow.

To explain the mechanism of flow generation, the renowned American hydrologist Robert E. Horton assumed that the overland flow was generated on all (or a significant part) of the watershed area as sheet flow, and only when a excess of rainfall (or snowmelt) over infiltration was formed. In the initial period of rain, all water may infiltrate into the soil, but the infiltration rate decreases as a function of time because of increases in the soil moisture content at the soil surface. At some point in time (it is called the ponding time), the infiltration rate drops below the rainfall rate. The accumulated water covers all the drainage area by a thin layer and then begins to flow along the slope to the rills and gullies. Thus, the necessary conditions for the generation of overland flow by the Horton mechanism are: (1) a rainfall rate greater than the hydraulic conductivity of the soil; and (2) a rainfall duration longer than the required ponding time for a given initial moisture profile. Field research of rainfall runoff generation confirms that such a mechanism is often observed during highly intensive showers on arid and semi-arid watersheds, which lack enough vegetation cover to retain moisture. At a suitable combination of soils and topography, high rainfall rates lead to splash erosion and transport of soil particles by water fluxes. The transported sediments are deposited on the land surface and can significantly decrease soil permeability or form an impermeable crust. Horton overland flow may also occur during snowmelt on the plain watersheds when the permeability of frozen soil is low.

However, the analysis of runoff coefficients and field observations shows that the rainfall Horton overland flow occurs in the temperate climate zone very seldom. Sheet flow is usually observed only on partial areas where the soil profile is saturated before the start of rainfall. In this case, water accumulates on the land surface due to the soil's inability to absorb any more moisture (regardless of the difference between rainfall intensity and infiltration rate), and such a type of overland flow is called saturation overland flow. Typically, saturation overland flow occurs when long-duration rains cover the areas where the initial water table is shallow and it can quickly rise to the land surface, or when an impermeable layer is relatively close to the land surface. Such areas are commonly located in valley bottoms, along streams, and near wetlands, but various subsurface conditions can also cause the formation of saturated zones in topographically-high parts of a basin. The area of saturation depends on the season and it can expand and contract during a storm and may differ from storm to storm. Thus, the source area of saturation overland runoff can significantly vary. Basins generating variable source runoff often display the same type of relationship to rainfall and watershed conditions as are recognized for Hortonian overland flow.

Because depths of sheet overland flow are commonly less than 1–2cm, and ranges of velocity variation in time and space are also small, the velocities of overland flow are a one-to-one function of flow depth, watershed slope, and roughness of land surface. As a result, to describe overland flow for a one-dimensional case the following equations can be used:

$$
\begin{aligned}
& b_s \frac{\partial h_s}{\partial t} + \frac{\partial q_s}{\partial x} = R_e b_s \\
& q_s = \frac{1}{n_s} i_s^{-1/2} h_s^{5/3} b_s
\end{aligned}
\qquad (20)
$$

where b$_s$, h_s, q_s, i_s, and n_s are respectively the depth, discharge, width, slope, and Manning roughness coefficient for overland flow, R_e, is the rainfall excess, t is time, and x is the space coordinate.

Subsurface runoff may generate in the unsaturated zone of the soil above the layers with the temporary low permeability, or in the temporary saturated soil layers. In mountainous regions, subsurface runoff is often observed in the rough soil mantles lying above the ground with small hydraulic conductivity. Subsurface storm runoff may also occur through macropores resulting from animal or vegetation action, and in fractures and joints between soil strata. These paths may be enlarged by erosion and sediment transport, and piping drainage systems may be formed. Depending on their origins and the stability of their walls, pipes may vary in diameter from less than 10 mm to more than 1 m. In the unsaturated zone, pipe networks carry water in turbulent flows at velocities which match those for open channels, sometimes over distances of several hundred meters. Pipes also provide bypass routes for water in the saturated zone, essentially increasing seepage velocity. This mechanism of subsurface runoff generation may occur when the capillary fringe in regions of shallow groundwater (usually near streams) becomes quickly saturated, resulting in water flow into the stream. The subsurface water may rise to the land surface and form overland flow, which provides a mechanism for the rapid discharge of subsurface water to stream channels.

The one-dimensional equations for subsurface flow can be presented as

$$(\theta_{mp} - \theta_s) b_s \frac{\partial h_g}{\partial t} + \frac{\partial q_g}{\partial x} = R_g b$$
$$q_g = K_g i_s h_g b_s \qquad (21)$$

where h_g, and q_g are the depth and discharge of subsurface flow, R_g is the recharge of subsurface water, θ_s is the soil capillary porosity, θ_{mp} is the maximum porosity including macropores, and K_g is the coefficient characterizing the horizontal hydraulic conductivity of the soil. The main assumptions here are the following: 1) subsurface flow follows the same slope and has the same width of flow strip as the overland flow; 2) the saturated layer h_g is formed above the base of the subsurface ground layer under consideration; 3) the capillary water (i.e. water at a moisture content less than θ_s) does not take part in the horizontal movement.

In spite of an enormous variety of climatic and physiographic conditions which the runoff may generate, in most cases it is possible to establish the general peculiarities of runoff generation for large physiographic zones and types of landscapes (for example, tundra, forest, steppe, arid and rainforest zones, and urban, agricultural, or forest lands). Special mechanisms of runoff generation are typical for mountain, swamp, permafrost, and glacier watersheds.

Mountainous regions cover more than 20 percent of the land and provide the main source of available water resources in many arid and semi-arid areas. Mountain watersheds commonly have a well-expressed vertical zonality in climatic and physiographic conditions. The complex structure of mountain topography, and its interaction in blocking and uplifting large air masses, results in widespread and intensive precipitation on windward slopes with great seasonal variation. A considerable increase of precipitation with altitude can generally be observed, but the value of this increase varies depending on climatic zones and exposition of mountain ranges. Mountain topography strongly affects the spatial distribution of water and energy, and generates heterogeneity at all scales. Large variations of albedo, soil, and water storage conditions in relation to the surface conditions (rocks, snow, vegetation, altitude, exposure, etc.) cause local variations in the structure of the atmospheric boundary layer and heat fluxes. At mountain heights greater than 1 km, meteorological processes are influenced by the state of stratification of the atmosphere. Runoff of many mountain rivers is of mixed rainfall and snow melt origin. A characteristic feature of mountain rivers is extreme seasonal variation. Most runoff is produced quickly as overland flow, or subsurface flow in shallow rough ground layers. Immediately after rain or snow melt, destructive floods transporting significant amounts of hard material and sediments can occur. In dry periods, subsurface flow often results in increased soil moisture in lower slopes and valleys, giving better-developed vegetation than that on the upper slopes. The ground flow is generally small. The runoff coefficients are high (0.4–0.6) and vary within a narrow range.

Persistent swamps occupy only about 2 percent of the land surface, but in some regions of the world (for example, central parts of South America and the northwest area of the European part of Russia) swamp watersheds contribute a significant portion of runoff. A characteristic feature of swamp watersheds is that the water table is situated closely to the land surface, so that the runoff varies only marginally during the warm period. However, swamp watersheds respond quickly to large rainfalls. Evapotranspiration from swamps is usually considerably higher than from dry neighboring areas and leads to a decrease of annual runoff.

In the permafrost river basins, most hydrological activities occur in the active layer. Because of the relative impermeability of frozen ground, runoff losses are determined by evaporation and water storage in depressions, peat mats, and large-pored soils. The value of free basin storage capacity depends on the antecedent hydrometeorological conditions of the current year, or foregoing years. The year-to-year change in basin water storage can reach 10–15 percent of the annual precipitation. During snowmelt, the main mechanism of runoff generation is overland flow. A part of the melt water can freeze in the snow, in the peat mats, or in the ground during the nightly drop in air temperature, and because of low ground temperatures generally. The water frozen in the surface basin storage and in the active layer of the ground can generate a significant portion of river runoff during the entire warm period. There are river basins where floods have resulted from the melting of ice after cessation of snowmelt. Subsurface flow starts after the beginning of the melt of ice in the ground, increases gradually in line with increases in the depth of thawed ground, and may become the main mechanism of rainfall runoff generation. The groundwater component of river runoff in the permafrost regions is usually small.

A glacier can be considered as a watershed whose characteristics change during the course of a year. In early spring the surface of glacier begins to thaw. Melt water and rain are effective agents of heat transfer and quickly thaw holes in the lower layer. Gradually, the area between the holes also becomes thawed, and the snowpack reaches a uniform temperature at the melting point. The thawed zone gradually moves to higher altitudes. In late spring, the glacier is covered entirely by a thick snowpack. Melt water and rainfall must travel through the snowpack by slow percolation (unsaturated flow), until reaching meltwater channels in the solid ice below. In summer, some bare ice is exposed and here there may be surface drainage. In autumn, a dense snow layer covers only part of the glacier and bare ice is exposed over the rest of the glacier. Melt water and liquid precipitation travel very quickly from the surface to the outflow stream. In winter, snow accumulates and the surface layer freezes. A small amount of water deep within the glacier slowly drains out during the winter. The lack of a direct relation between precipitation and runoff from a glacier is evident for all seasons except for late summer. The diurnal fluctuation of ice and snow melt usually corresponds to the diurnal fluctuation of discharge from the glacier, and reflects the peculiarities of the shape of the glacier.

2.8. The River Network and Movement of Water in River Channels

The construction of a river network and river channels is determined by climate, topography, and the geological structure of a river basin. At the same time, river networks and river channels exhibit amazing general regularity and organization. Each river network has a treelike structure with the outlet stream as the main trunk and tributaries that bifurcate into smaller and smaller rivers. It has been established that there are stable empirical laws controlling the construction of the river network. Stream links can be classified as follows: 1) channels that originate at a source are defined to be first-order streams; 2) when two streams of order ω join, a stream of order (ω+ 1) is created; 3) when two streams of different order merge, the channel segment immediately downstream is taken to be the continuation of the higher order stream. These laws can be summarized as:

Law of stream numbers $$\frac{N_{\omega-1}}{N_\omega} = R_B \quad (22)$$

Law of stream lengths $$\frac{\overline{L_\omega}}{\overline{L_{\omega-1}}} = R_L \quad (23)$$

Law of stream areas $$\frac{\overline{A_\omega}}{\overline{A_{\omega-1}}} = R_A, \quad (24)$$

where N_ω, L_ω, and A_ω are, respectively, the number of streams of a given order ω, their mean length, and the mean area contributing to streams of this order; R_B is called the bifurcation ratio and normally ranges from 3–5; R_L ranges from 1.5–3.5; R_A varies from 3–6. The river length is related to the river area by a power function where the exponent is about 0.6. The fact that this exponent is not 0.5 (river length is not proportional to the square root of area) results from the fact that river basins become morenarrow as they enlarge. Geometrical characteristics of river channels are determined by climatic and topographic conditions; however, the velocity, width, and depth of flow can be presented as power functions of discharge whose exponents change in narrow ranges.

During movement in the river system the runoff may significantly increase due to the lateral inflow, or vary in time due to change of runoff generation on the watershed. As a result, for the most part of the year rivers have an unsteady flow varying in space and time. The unsteady character of river flow can especially be observed during floods, or at some river reaches where there are river flow control constructions.

To describe the unsteady flow of river channels, the Saint–Venant equations are usually applied. The basic and general assumptions underlying the development and the applicability of these equations are: 1) flow is gradually varied, or the vertical velocities are considered small in comparison with the longitudinal velocities;; 2) the velocity distribution along a vertical, in unsteady flow, is the same as in steady flow;3) the friction resistance in unsteady flow is the same as in steady flow. The Saint–Venant equations can be written in the following form:

$$\frac{\partial A}{\partial t} + V\frac{\partial A}{\partial x} + A\frac{\partial V}{\partial x} = 0 \quad (25)$$

$$\frac{\partial V}{\partial t} + V\frac{\partial V}{\partial x} + \frac{g}{A}\frac{\partial(\bar{y}A)}{\partial x} + \frac{Vq}{A} = g(S - S_f) \quad (26)$$

where A is the area of channel cross section, V is the water velocity, q is the lateral inflow per unit channel length, h is the depth of flow, g is acceleration of gravity, and S is the channel slope. The term S_f is the friction slope which can be determined as $S_f = V^2P/C^2 A$ (this relationship is known as Chesy's law for the turbulent flow), C is Chesy's roughness constant, P is the wetted perimeter of channel. The first equation is the continuity (or mass-conservation) equation, and the second is the momentum equation. The first term in the left-hand side of the second equation $\partial V/\partial t$ corresponds to local inertial acceleration or velocity change. The second term $V(\partial V/\partial x)$ is the inertial acceleration, corresponding to velocity change in space. The third term is momentum change induced by pressure differentials related to flow depth changes. The fourth term is the momentum change caused by the incoming mass of lateral inflow. The terms on the right are the gravity force and the friction force related to channel slope and roughness. The inertia terms and the term representing the influence of lateral inflow are usually negligible in comparison with those of bottom slope and friction for an uncontrolled river regime.

For many rivers with large and moderate slopes where dams are absent, the term related to flow depth changes is also small. In this case $S = S_f$, and there is a one-to-one relation between the river discharge $Q = AV$ and the area of channel cross-section (or river stage). Thus, the unsteady river flow equations can be written as

$$\frac{\partial A}{\partial t} + \frac{\partial Q}{\partial x} = q \qquad (27)$$

$$Q = \alpha A^m$$

where α and m are constant coefficients (the dependencies $\alpha = CS^{0.5}$, $m = 3/2$ are commonly used). This approximation of unsteady flow description is called the kinematic wave model. The system (27) can be also presented as

$$\frac{1}{c}\frac{\partial Q}{\partial t} + \frac{\partial Q}{\partial x} = q \qquad (28)$$

where

$$c = \left.\frac{dQ}{dA}\right|_{x=x_c} \qquad (29)$$

is the celerity of movement of the discharge Q in direction x in any x_c (this is known as the celerity of kinematic wave). Following from the relation between the river discharge and the area of channel cross-section, the celerity of kinematic wave increases with increase of the discharge, and, as a result, the forward part of the kinematic wave steepens and the back part attenuates. The maximum discharge of kinematic waves does not change. Kinematic wave behavior is very close to the behavior of most natural floods.

In cases where the term related to flow depth changes is relatively large (the slope of river channel bed may be mild, or there may be backwater effects related to flood control or a significant inflow), the Saint–Venant equations can be simplified to the so-called diffusion analogy model:

$$\frac{1}{C}\frac{\partial Q}{\partial t}+\frac{\partial Q}{\partial x}=\frac{D}{C}\left(\frac{\partial^2 Q}{\partial x^2}-\frac{\partial q}{\partial x}\right)+q \qquad (30)$$

where C has the same meaning as in the kinematic waves, and D is a coefficient of diffusion coefficient that introduced the attenuation of the flood waves. The diffusion coefficient is derived from the equation

$$D=\left(B\frac{\partial S}{\partial Q}\right)^{-1} \qquad (31)$$

where B is the width of the channel.

2.9. Modeling the Hydrological Cycle of a River Basin

By constructing mathematical models of the hydrological cycle it is possible to chrystallize our conceptual understanding of individual processes of water turnover in nature, and obtain means for diagnosis and prediction of these processes. Complex hydrological processes usually occur over large areas, and the special heterogeneity of these events, together with a deficit of information on the characteristics of their environment, make the development of models of the hydrological cycle one of the most difficult problems of geophysics. The optimal structure for models describing the individual processes of the hydrological cycle depends not only on the required refinement of these processes, but to a significant extent on the availability of data needed in order to determine the parameters (coefficients) of the models. In many cases, information needed for choice of the structure of the model and determining its parameters are absent, and there sare only experimental field data received at the input and the output of the hydrological system under consideration. Less frequently, the structure of the model is chosen mainly on the basis of a priori knowledge of the process or system behavior.

For most practical tasks, the structure of hydrological models and their attendant parameters is determined on the basis of both a priori (theoretical) and experimental information. As a result, according to the proportion of a priori and experimental information for model construction, the models of processes of the hydrological cycle can be conditionally divided into three types: 1) black box model,s where only measurements at the input and the output of the hydrological systems are used; 2) gray box models (in hydrology, the term “conceptual models” is used more often), where measurements at the input and the output of the hydrological systems provide the main information, but conceptual understanding of processes is also applied; 3) physically-based models, in which the choice of the structure is based on a priori information on processes of the hydrological cycles (mainly, on fundamental laws of hydrophysics and hydrodynamics), , and the parameters are determined mainly on the basis of direct measurements of characteristics of the hydrological systems.

However, because of strong spatial variability of these characteristics, present-day measurement methodologies cannot provide the accuracy that is necessary for most hydrological problems. There are also always some inconsistencies between real natural processes and their representation in the models. As a result, to improve the accuracy of the model, some parameters have to be fitted. Such fitting (it is also called calibration) is commonly carried out by applying the comparisons of calculated and measured components of the hydrological cycle. Because runoff is the most exactly-measurable component of the hydrological cycle, in most cases runoff hydrographs at the outlets of river basins are used for fitting, and the construction of models of the hydrological cycle is commonly carried out for river basins with the available measurements of runoff. For fitting the parameters, the available measurements of evapotranspiration, soil moisture, and snow characteristics are also often used.

To decrease the number of fitting parameters, the hydrological systems in the models of the first and second types are usually assumed to be lumped (characteristics of systems and their conditions do not change in space), and, as a result, the models of hydrological processes are also frequently classified as either lumped or distributed. Lumped conceptual models, which are mostly applied in hydrological practice, contain aggregated empirical parameters that have a complicated physical interpretation and a large range of variation. In contrast, distributed physically-based models include parameters with clear physical meanings, and if there are no direct measurements of these parameters in a given river basin, they can often be gained from laboratory or field investigations of hydrological processes in similar physiographic conditions. A good correspondence between the model structure and the prototype may be supposed to facilitate fitting the model parameters and increase the predictive capability of the model. In many cases, even the a priori information on the possible range of parameter values can considerably decrease uncertainty in the estimation.

Another important advantage of physically-based models is the opportunity to use simulation to explore different assumptions and physical hypotheses about the particular basin and mechanisms of the hydrological processes. Such sensitivity analyses, coupled with observed data, allow for the “deciphering” of dominant mechanisms, and simplify the choice of model structure. The development of databases that include information with high resolution in space (in the form of digitized maps, digital elevation models, and remote sensing data from satellites, airplanes, ground-based weather radars, and more modern Geographic Information Systems (GIS)) that can help in handling this information, essentially widens the opportunities of construction of distributed physically-based models of the hydrological cycle. The use of general physical laws and meanings of parameters in physically-based models also simplifies the coupling of models of the hydrological cycle with models of meteorological and other geophysical phenomena.

2.10. Human Influence on the Terrestrial Hydrological Cycle

Human activities that change the land cover of river basins and are aimed at regulating the water fluxes in nature can considerably change the hydrological cycle of the separate river basins, and even of large regions. A striking example of such change is the present-day situation in the Aral Sea basin, where intensive irrigation has resulted in almost full cessation of the water inflow from the Syr-Darya River and the Amu-Darya River, as well as the drastic drop in the Aral Sea level. Other well-documented examples include the increased drought risks in the Mediterranean and the Sahel, following removal of vegetation by forest clearing and overexploitation respectively. There are also some indications that the considerable changes in scale and frequency of flooding in the Ganges basin may be explained by deforestation in the local mountainous region.

Due to human activities, the natural hydrological cycle of most river basins is becoming more and more transformed and regimented. The main stream flow regulation methods are construction of dams, levees, barrages, and dikes, which provide water accumulation, decreasing flood flow, and increasing low flow. The major effects of reservoir construction on the hydrological cycle (excepting runoff control) are an increase of evaporation and a rise of groundwater table. In dry regions, evaporation losses from the reservoir water surface may be so large that they seriously compromise any potential gains. At the same time, in the conditions of moderate climate, the reservoir losses on evaporation are relatively small. For instance, evaporation from the reservoirs in the Volga River basin (where there are about 300 reservoirs the storage capacity over 1 000 000 cubic meters) constitute less than 3–5 percent of the Volga River runoff. The rise in groundwater level along the reservoir periphery and in surrounding areas changes the runoff generation mechanism on these areas. Gradual change of the river flow regime can occur as a consequence of decreasing the river's ability to transport sediments, especially in upper parts and in reservoirs. Reduction of sediment input at the dam site reduces the river channel slope and the bed sheer stress, resulting in dropping flow velocities and the development of river meandering.

The impact of irrigation on the hydrological cycle is especially revealing in the arid regions, but it is also considerable in regions with moderate climate where irrigation is of supplementary character. Diversion of water for irrigation purposes from surface or groundwater resources modifies the natural hydrological processes. It is common for runoff and evaporation from irrigated areas to increase significantly. Irrigation in river basins where there is no additional method of supply often leads to runoff reduction in the outlet site. In many dry regions, a considerable rise in the groundwater table can occur because of water filtration from reservoirs, leakage from water distributing systems, and faulty irrigation technology. Such a rise may cause waterlogging of plants and development of soil salinization.

To remove excess water from waterlogged soils, drainage is applied in many regions of the world. The primary effect of drainage is the lowering of the groundwater table and the extension of the layer with unsaturated soil. As a result, evapotranspiration may considerably drop (in some cases, by more than 50 percent). The improvement of hydraulic conditions due to drainage increases flow velocities. In the first years after construction of a drainage system, the annual runoff can increase by 20–30 percent. Especially large runoff rises can be observed during the low runoff months in winter and summer. Acceleration of flow also leads to a significant increase in flood peaks. After 10–15 years the impact of drainage on runoff decreases.

Because the quality of groundwater is mostly far better than that of surface water, and its temperature is relatively constant, large volumes of groundwater are extracted for domestic and industrial use in different regions of the world. If groundwater is extracted from confined aquifers below impermeable layers, the groundwater table is not, or is only slightly, affected. However, at some river basins the groundwater table often drops steeply, and this may reduce the surface runoff and the lower level of the small rivers. In many coastal areas, the extraction of groundwater leads to the seawater intrusion.

Together with direct change of the hydrological regime of the river basins by means of stream channel control, irrigation, drainage, and groundwater abstractions, changing the land use of river basins can exert significant influence on the hydrological cycle. Consequences of land use change may be revealed gradually, and be masked by climate variations, but an essential transformation of hydrological regime can occur. The most significant distortions of the hydrological cycle are observed in urbanized areas. The replacement of natural land cover by the urban impermeable surface causes great reductions in infiltration and evapotranspiration. The rainfall runoff from urbanized areas is mainly generated as overland flow and reaches the river drainage system very quickly. Accordingly, the rainfall flood volumes may increase by several times, and the peaks of the hydrographs may increase by 10–15 times. At the same time, snow transport may result in a decrease of snowmelt runoff. Due to reduction of infiltration and groundwater abstractions for urban water supply, falls in the groundwater table are also observed in urban neighborhoods.

The effects of agricultural and forestry practices on the hydrological cycle are less apparent, and depend, to a significant extent, on the physiographic and climatic conditions. It is evident that ploughing, especially contour ploughing, usually breaks up overland flow and increases infiltration. Some special types of ploughing may increase the depression and detention storage on gentle slopes from about 8–10mm (in the natural conditions) to 30–40mm. Tillage and the activity of plant root systems modify the structure of the upper soil layer and change not only the vertical permeability, but also the water retention capacity. Extension of vegetation cover and the leavesarea increases the interception of precipitation and evapotranspiration. Control of overland flow by dense permanent grasses on steeper slopes can reduce storm runoff from small watersheds by 20–25 percent. However, the relative influence of all these changes on annual and flood runoff is determined by the river basin and climate characteristics. In dry regions (for example, in the steppe zone), the 15–20 percent change of the annual runoff caused by agricultural practice has been fixed, and in different years these changes reached 30–40 percent. At the same time, in the wet regions, especially in the forest and northern forest steppe zone, the impact of agricultural practices on runoff may be neglected.

The main clearly-expressed effects of deforestation on the hydrological cycle of a river basin are the increases in transpiration and interception of precipitation, which in turn result in a decrease of the volume of total runoff. Deforestation reduces infiltration and improves the conditions for overland flow. As a consequence, flood runoff and peak discharges may significantly increase. At the same time, the higher infiltration of forest soils increases the opportunity for recharge groundwater, and the flow of small rivers tends to be more sustained, especially in the case of the generation of snowmelt runoff, when forests further sustain flow by delaying the snowmelt. A rise in the groundwater table and an increase of ground runoff may also raise the low flow of medium- and large-sized rivers. Such effects often result in the conclusion that forests increase runoff. However, careful observations on representative and experimental basins do not commonly confirm such conclusions. For example, results of fifteen individual watershed-scale experiments, involving various rates of forest cutting, carried out during 50 years at the Coweeta Hydrologic Laboratory in Southern Appalachia, indicated that deforestation increased and afforestation decreased annual and monthly runoff, but the magnitude of the responses was highly variable. It was established that streamflow response to forest cutting was inversely proportional to solar energy input (as an evapotranspiration index). The alteration of the monthly runoff is in close agreement with changes in evapotranspiration. At the lowest flow, the monthly runoff was about 100 percent greater from the clearcut forest than the uncut forest. Clearcutting has little effect on flow during winter and early spring.

Long-term observations have also shown the strong dependence of runoff volume on the type of vegetal cover. Conversion of hardwood to pine reduced the annual runoff by 25 cm and produced significant reductions of monthly runoff. At the same time, forest cutting has led to a considerable increase in flood peaks. Similar results have been also received on the basis of analyzing data obtained in other physiographic conditions. Research carried out in the forest zone of the European part of Russia has shown that the influence of the forest on evapotranspiration and runoff significantly depends on the age of the forest. Cutting of old forest may not alter evaporation, and the increasing accumulation of snow may even lead to some growth of spring runoff. In many regions, deforestation has resulted in a significant increase in disastrous floods and has also caused severe soil erosion.

3. The Global Hydrological Cycle

Most of the Earth's water (about 1400 million cubic kilometers or 96.5 percent of the totalresources) resides in the oceans. Continental water makes up about 3.5 percent of the Earth's water; however about three-quarters of this amount (29 million cubic kilometers) is present as polar ice caps and glaciers, and 5.3 million cubic kilometers as deep groundwater. Thus, only the remaining fraction can take part in the water exchange between the oceans, the atmosphere, and the continents. This remaining part includes shallow groundwater and soil moisture (about 4.3 million cubic kilometers); water in lakes, reservoirs, and swamps (0.125 million cubic kilometers); water storage in river channels (2.1 thousand cubic kilometers); biosphere water (about 0.6 thousand cubic kilometers). The amount in the atmosphere is only 0.013 million cubic kilometers (0.001 percent).

The dominant mechanisms of the global and terrestrial hydrological cycles are the water exchange between the atmosphere and the Earth's surface, and between the oceans and the continents. The main energy sources of this exchange are solar radiation and gravitational force. Every year the Earth's surface evaporates about 580 thousand cubic kilometers of water. The energy necessary to convert this amount of water into vapor corresponds to 36 percent of the solar radiation absorbed by the whole Earth. About 85 percent of the total global evaporation is evaporation from the ocean surface and the rest is the land surface evaporation. The average annual evaporation from the ocean surface unit area is 1400 mm; at the same time, this value for the land surface is only 485 mm (approximately three times less). On average, the evaporating water remains in the atmosphere for only about 8–10 days, and, as a result of condensation of air moisture, most of the evaporated water falls on the Earth's surface as rain or snow.

The total annual precipitation on the Earth's surface is approximately equal to the total global evaporation. However, the total annual evaporation from the ocean surface is larger than the precipitation. The surplus of the evaporated water (about 47 000 cubic kilometers) is transported by air currents from the oceans to the continents. About 40 percent of the precipitation falling over continents returns to the oceans as river runoff (approximately 450 000 cubic kilometers), or as direct groundwater discharge to the oceans (about 20 000 cubic kilometers). The other part reevaporates and falls back to the land. On average, the water transported from the oceans is recycled and precipitated 2.7 times over land before it runs back to the oceans.

Let us consider the spatial-temporal distribution of the components of global hydrological cycle.

3.1. Precipitation

Regional distribution of precipitation depends on the general circulation of both the atmosphere and the oceans, as well as on the position and form of continents. Average annual precipitation is generally heaviest in tropical regions, where the trade winds converge in the low-pressure belt, and decrease poleward with decreasing average annual air temperature. However, there are essential deviations from this main trend. The areas near 30 degrees latitude have relatively small precipitation because in this region the air, haven risen near the equator, sinks and is counter-productive to precipitation generation. At the same time, the rising air in the middle latitude gives precipitation increase as a result of more frequent cyclonic activity. Polar regions have very small precipitation because of the influence of sinking air and the very cold atmosphere, which contains relatively little water vapor. An important factor in the continental distribution of precipitation is the distance from a moisture source. In many regions the spatial distribution of precipitation is affected by mountain systems.

The maximum precipitation over the world ocean is observed in the equatorial zone (about 1600 mm per year). Along the equator lie the areas where abundant rains fall during the whole year. Near the coasts of Indonesia and Burma, the precipitation reaches more than 4000 mm per year. At the same time, the tropical zones of the Northern and Southern hemispheres, which are under the influence of trade winds, are characterized by low amounts of precipitation. In the regions adjoining the Sahara Desert and the Arabian Peninsula ,the precipitation is less than 50 mm per year.

Between 20 and 30 degrees of northern latitude, the average amount of precipitation is about 700 mm.

In the annual zone the amount of precipitation reaches more than 1200 mm on the average. In the sub-Arctic and sub-Antarctic zones, precipitation decreases, reaching 300–400 mm in some regions. In the Arctic zone, precipitation continues to decrease, reaching 150–200 mm on the average.

The annual precipitation over the Earth’s land is about 800 mm. The least precipitation occurs in Antarctica (about 178 mm), and in Australia (456 mm). The most humid area is South America, where the annual precipitation is about 1600 mm. The annual precipitation in Europe, Asia, Africa, and North America has close proximity (790, 740, 740, and 756 mm, respectively).

Temporal variation of precipitation on large spatial scales is determined by the seasonal periodicity of large-scale atmospheric circulation, and by the distribution of warm and cold sea currents. Trade winds and monsoons play an especially important role in this variation. For example, the zone of equatorial monsoons is characterized by well-pronounced wet and dry seasons. In temperate zones, the temporal variation of precipitation essentially depends on the cyclones that form over the oceans. Short-time variation of precipitation is mainly a result of front passages or cyclonic storms, as well as diurnal changes in solar radiance. The coefficient of variation of annual precipitation is comparatively small (on the average, 0.2–0.3); the coefficients of serial variation of the first order are about 0.10–0.15.

3.2. Evaporation

Evaporation from the surface of the oceans significantly exceeds evaporation from the land and is the principal source of water in the atmosphere. Spatial distribution of the evaporation from oceans depends on distribution of solar radiation, atmospheric circulation, and the thermal regime of oceans. In conformity with latitudinal changes in solar radiation, annual evaporation from the ocean surface decreases towards higher latitudes. However, this change is disrupted by the influence of atmospheric circulation and sea currents. The maximum evaporation from the oceans occurs in the trade wind zones between 10 and 20 degrees North, and between 10 and 20 degrees South, where the dry air flows take place. In the Atlantic Ocean, the evaporation in these latitudes reaches 1950 mm and 1700 mm, respectively; in the Indian Ocean, it reaches 2000 mm and 2100 mm; and in the Pacific Ocean, it reaches 2000 mm and 1950 mm, respectively. The Arctic Ocean is an area of very low evaporation (in the central part of this region the evaporation is less 100 mm per year).

Variation of evaporation from the ocean surface within the year is determined by changes of air and ocean surface temperature. Evaporation is highest during the autumn–winter period, when the water surface is warmer than the air. In summer, the water surface becomes colder than the air and evaporation is low. In some areas during the summer months, condensation of water vapor can occur. Seasonal variation of evaporation is largest in the temperate latitudes.

The spatial distribution of evaporation from the land depends not only on atmospheric circulation but also on natural landscapes, especially types of vegetation and soil. This can explain a significant diversity in the mean evaporation from different continents. The evaporation from the islands of Oceania amounts to 1060 mm per year; at the same time, the evaporation from Antarctica is close to zero. Evaporation figures from South America, Africa, Europe, North America, and Asia are 910, 587, 507, 418, and 416 mm per year, respectively.

3.3. Runoff from the Continents

The river runoff from only 80 percent of the land reaches the oceans. Precipitation on approximately 30 million square kilometers, located mainly in the arid and semi-arid regions, evaporates entirely or forms runoff that flows in internal watersheds. This runoff is about 1010 cubic kilometers per year (only 2 percent of entire global runoff), but internal watersheds can occupy large areas (for example, the drainage basins of the Caspian and Aral seas). The runoff coefficient for the whole land is 0.39. Due to very low evaporation, the runoff coefficient of Antarctica is close to 1. The runoff coefficient of North and South America is about 0.43. The runoff coefficient of Asia is also close to this value at 0.40. The runoff coefficient of Europe is about 0.30. The figure for Africa is 0.20. The greatest runoff values (3000–4000 mm per year) are observed in the eastern areas of New Guinea, on the Malaysian Archipelago, on the Pacific coast of the southern extremity of South America, and on the western shores of Europe (Norway).

The largest contribution to the total value of runoff is made by Africa (31 percent). South America gives 25 percent of the total runoff. The contributions of the other continents are as follows: North America, 17 percent; Asia, 10 percent; Europe, 7 percent; Antarctica, up to 5 percent; Australia and Oceania, up to 4 percent.

The coefficients of variation of annual runoff vary in a significant range, depending on the climate, runoff origin, and runoff depth. The typical coefficients of variation of annual runoff of the rivers with rainfall feeding in dry climate regions are 0.8–1.0; for rivers of mostly snowmelt feeding in dry climate regions these coefficients are about 0.5–0.8; the large plain rivers with mixed feeding have values of 0.25–0.40; the rivers which flow from large lakes have the typical coefficients of variation of annual runoff of 0.15–0.25. The typical coefficient of serial variation of the first order is about 0.2–0.3.

3.4. The Role of the Terrestrial Hydrological Cycle in the Global Climate System and Global Change

The terrestrial hydrological cycle is an important component of the global climate and biospheric system. The processes within the terrestrial biosphere, the atmosphere, and the hydrological cycle are intrinsically coupled and associated with a continuous exchange of water, energy, and materials. Water vapor in the atmosphere absorbs the incoming solar radiation and holds the long wave radiation from the Earth’s surface. A number of studies have already shown a significant influence of soil moisture change and evapotranspiration on the continental distribution of precipitation and air temperature. This means that these land surface processes must be taken into account in