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.
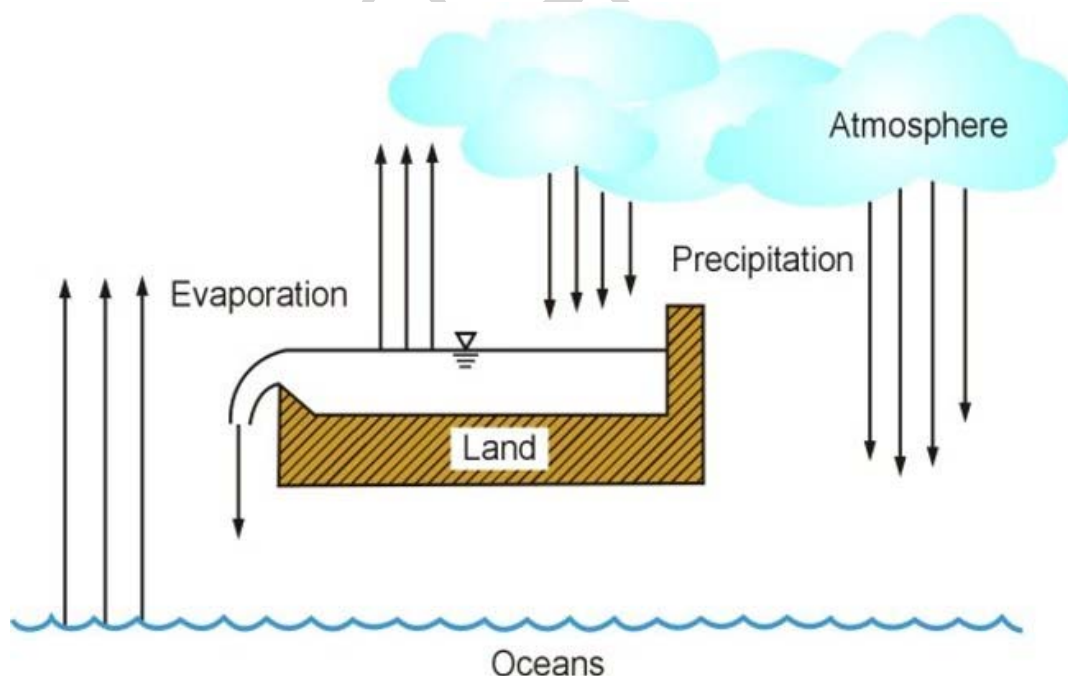

Figure 1. Global hydrological cycle are presented .

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.

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Biographical Sketch

Lev S. Kuchment has held the position of Head of the Laboratory of the Hydrological Cycle at the Water Problems Institute of the Russian (formerly USSR) Academy of Sciences, Moscow, since 1977. He was born in the Khmelnizky region of the Ukraine in 1937, attending the Hydrometeorological Institute at Odessa between 1954–1959, and completing his postgraduate education at the Hydrometeorological Centre of USSR, Moscow, where he gained a Ph.D. in . in Physics and Mathematics"Modeling Unsteady Flow in River Systems." Later, he gained a D.Sc. in Physics and Maths ("Modeling Runoff Formation Processes"), and was made Professor of Hydrology and Water Resources in 1985. Professor Kuchment is the author of 140 scientific publications, including seven books. He is an expert in the modeling of hydrological processes and hydrological forecasting and design. His interests have included: snow cover formation; heat and moisture transfer in soil; evapotranspiration; overland, subsurface and groundwater flow; interaction of surface water and groundwater; unsteady flow in river channel systems; water quality formation; the hydrological cycle as a whole; and dynamic–stochastic models of runoff formation. He has also developed a number of models and techniques for short-term and long-term forecasting of rainfall and snowmelt floods. A number of his papers are devoted to the inverse problems of hydraulics. He has also dealt with estimation of human-induced changes of the hydrological cycle and possible hydrological impacts of global climate change. In recent years, his main research fields have been modeling the global hydrological cycle and risk assessment of catastrophic floods. Professor Kuchment has also held the following positions: vice-president of the International Commission for Resources Systems of the International Association of Hydrological Sciences (IAHS); full member of the Russian Water Management Academy; Chairman of Commission of Water Resources of Russian Geophysical Committee; Member of Working Group (IAHS/WMO for GEWEX); associate editor, *Journal of Hydrology*.