THE HYDROLOGICAL CYCLE: SYSTEM APPROACH IN HYDROLOGY.

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INTRODUCTION: -

The hydrologic cycle is a conceptual model that describes the storage and movement of water between the biosphere, atmosphere, lithosphere, and the hydrosphere (see Figure 1,2,3). Water on this planet can be stored in any one of the following reservoirs: atmosphere, oceans, lakes, rivers, soils, glaciers, snowfields, and groundwater.

Water moves from one reservoir to another by way of processes like evaporation, condensation, precipitation, deposition, runoff, infiltration, sublimation, transpiration, melting, and groundwater flow. The oceans supply most of the evaporated water found in the atmosphere. Of this evaporated water, only 91% of it is returned to the ocean basins by way of precipitation. The remaining 9% is transported to areas over landmasses where climatological factors induce the formation of precipitation. The resulting imbalance between rates of evaporation and precipitation over land and ocean is corrected by runoff and groundwater flow to the oceans.

Water is continually cycled between its various reservoirs. This cycling occurs through the processes of evaporation, condensation, precipitation, deposition, runoff, infiltration, sublimation, transpiration, melting, and groundwater flow. Table 1.3 describes the approximate residence times of water in the major reservoirs. On average water is renewed in rivers once every 16 days. Water in the atmosphere is completely replaced once every 8 days. Slower rates of replacement occur in large lakes, glaciers, ocean bodies and groundwater. Replacement in these reservoirs can take from hundreds to thousands of years. One of the oldest groundwater is found beneath the Sahara Desert in Egypt (Nubian aquifer), where the water ages range from about 200 ka to 1 million years old, depending on location [18].

Some of these resources (especially groundwater) are being used by humans at rates that far exceed their renewal times. This type of resource use is making this type of water effectively nonrenewable.

Every year the turnover of water on Earth involves 577,000 km^3 of water. This is water that evaporates from the oceanic surface (502,800 km^3) and from land (74,200 km^3). The same amount of water falls as atmospheric precipitation, 458,000 km^3 on the ocean and 119,000 km^3 on land. The difference between precipitation and evaporation from the land surface (119,000 - 74,200 = 44,800 km^3/year) represents the total runoff of the Earth's rivers (42,700 km^3/year) and direct groundwater runoff

Fig. 1 Schematic of the hydrological cycle.
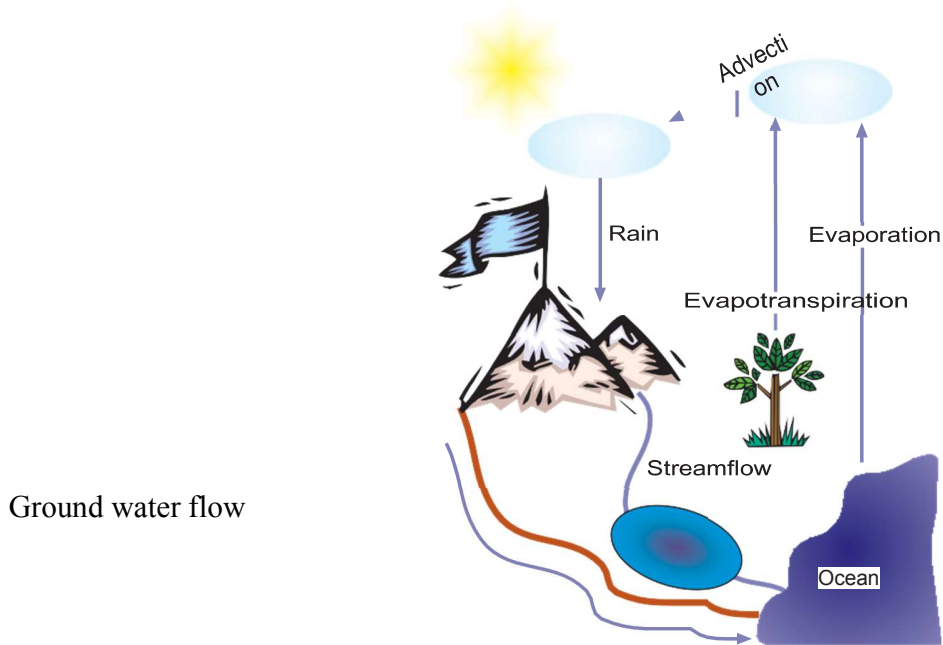
to the ocean (2100 km^3/year). These are the principal sources of fresh water to support life necessities and man's economic activities.

The hydrologic cycle is a model of the movement of water through the Earth system. It also is a pathway through which energy is transferred between the surface of the earth and the atmosphere. The unequal distribution of water leaves its imprint on the surface of the earth through the mosaic of biomes and landforms sculpted by the power of moving water and ice. Human survival requires adequate and safe water resources. The future of those resources hinge on our understanding of the delicate balance of water in our earth system.

The water balance is an accounting of the inputs and outputs of water. The water balance of a place, whether it is an agricultural field, region, or continent, can be determined by calculating the input and output of water. The major input of water is from precipitation and output is evapotranspiration. The geographer C. W. Thornth- waite (1899-1963) pioneered the water balance approach to water resource analysis. He and his team used the water-balance methodology to assess water needs for irrigation and other water-related issues.

Water is constantly moving within and above the earth in a cycle called the hydrologic cycle. Not only is the hydrologic cycle a cycle of water, it is a cycle of energy as well. There are six major components of this cycle: evapotranspiration, condensation, precipitation, infiltration, percolation and runoff.

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 2 and 3.
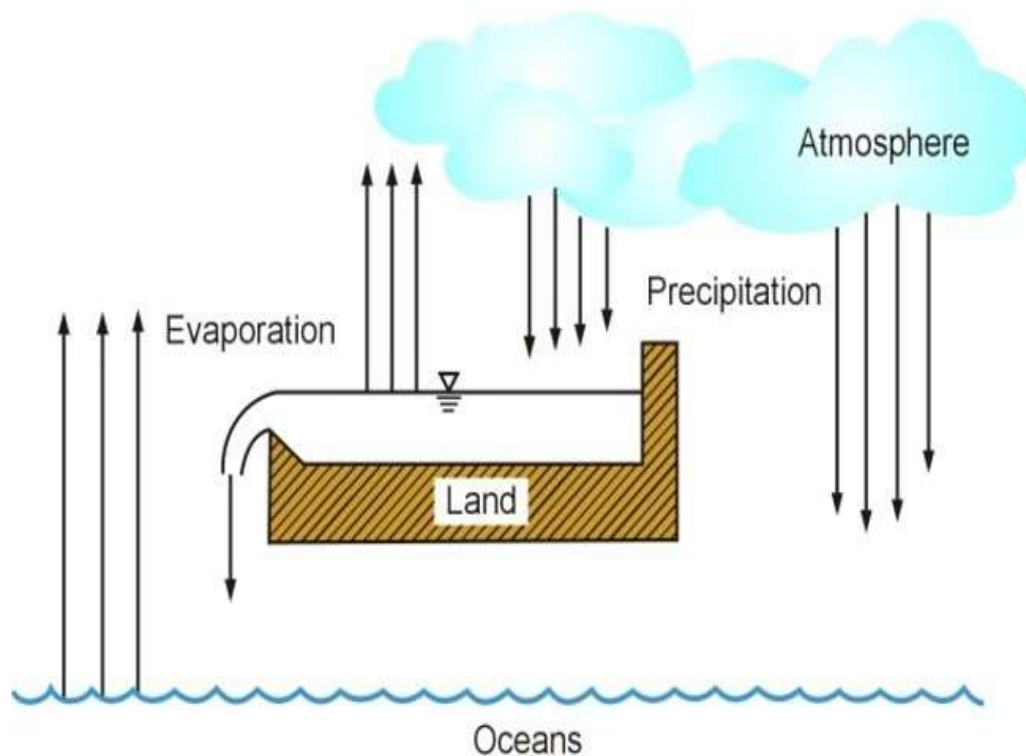

Figure 2 Global hydrological cycle are presented.

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.
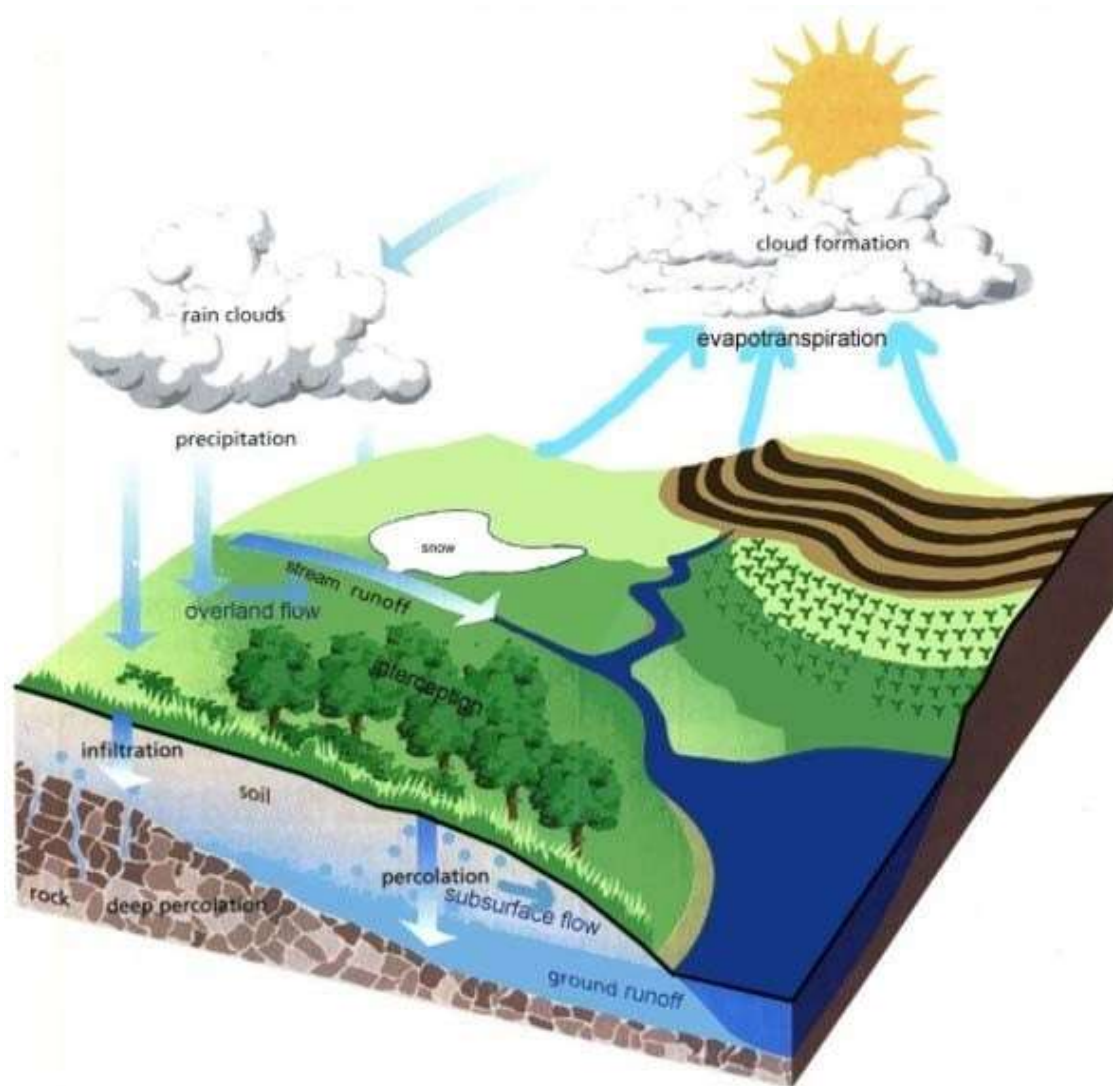

Figure 3. Terrestrial hydrological cycle.

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.

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 wind speed 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.

INTERCEPTION:-

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.

The water reaching the land surface begins to fill up first of all the 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 depression.

INFILTRATION: -

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 macrospores 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 macrospores. 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 macro porosity. 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.

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:

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.

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.

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.

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
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.

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 evaporated, 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.

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.

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, aquifer, 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.

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.

RUNOFF:-

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.

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.

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 hydro meteorological 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.

EVAPORATION :-

Definition
Evaporation is the phase change of liquid water into a vapor.

Evaporation is an important means of transferring energy between the surface and the air above. The energy used to evaporate water is called" latent energy". Latent energy is" locked up" in the water molecule when water undergoes the phase change from a liquid to a gas. Eighty-eight percent of all water entering the atmosphere originates from the ocean between 60 degrees north and 60 degrees' south latitude. Most of the water evaporated from the ocean returns directly back to the ocean. Some water is transported over land before it is precipitated out.

The conversion of water from a liquid into a gas. Approximately 80% of all evaporation is from the oceans, with the remaining 20% coming from inland water and vegetation. Winds transport the evaporated water around the globe, influencing the humidity of the air throughout the world.

Most evaporated water exists as a gas outside of clouds and evaporation is more intense in the presence of warmer temperatures. This is shown in the image above, where the strongest evaporation was occurring over the oceans and near the equator (indicated by shades of red and yellow).

CONDENSATION :-

Definition

Condensation is the cooling of water vapor until it becomes a liquid.

As the dew point is reached, water vapor forms tiny visible water droplets. When these droplets form in the sky and other atmospheric conditions are present, clouds will form. As the droplets collide, they merge and form larger droplets and eventually, precipitation will occur.

Condensation generally occurs in the atmosphere when warm air rises, cools and loses its capacity to hold water vapor. As a result, excess water vapor condenses to form cloud droplets. The upward motions that generate clouds can be produced by convection in unstable air, convergence associated with cyclones, lifting of air by fronts and lifting over elevated topography such as mountains.

OVER FLOW AND TRANSPORT :-

Definition: -

Transport is the movement of water through the atmosphere, specifically from over the oceans to over land.

Some of the earth's moisture transport is visible as clouds, which themselves consist of ice crystals and/or tiny water droplets. Clouds are propelled from one place to

another by either the jet stream, surface-based circulations like land and sea breezes, or other mechanisms. However, a typical 1-kilometer-thick cloud contains only enough water for a millimeter of rainfall, whereas the amount of moisture in the atmosphere is usually 10-50 times greater.

Most water is transported in the form of water vapor, which is actually the third most abundant gas in the atmosphere. Water vapor may be invisible to us, but not to satellites, which are capable of collecting data about the moisture content of the atmosphere. From this data, visualizations like this water vapor image are generated, providing a visual picture of moisture transport in the atmosphere.