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Article

Fundamentals of Climatology for Engineers

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Abstract: The study of long-term average weather patterns is known as climatology. It is a distinct field of study from meteorology and can be broken down into several subdivision. In order to predict the future, the knowledge of climatology is essential. In other words, with the help of climatology, we can figure out how likely it is that snow and hail will fall to the ground, and how much solar thermal radiation can reach a certain location etc. Climatology often focuses on how the climate has changed over time and how it has affected people and events. Both meteorology and climatology fall under the general term "meteorology", in particular, they are subdivision of research in the same field. In case of predicting the weather, meteorologists use variables such as humidity, air pressure, and temperature. This article's primary objective is to familiarize engineers with the fundamentals of climate and its processes so that they can effectively apply this knowledge to comprehend the climatic impact on water resources systems.

Keywords: Climate and Weather; Climate Model; Heat Transport; Radiation Balance; Atmospheric circulation; sea surface temperature; Planetary boundary layer; El Niño-Southern Oscillation (ENSO)

1. Climatology, climate and its Components

Climatology explains the physical processes of climate, including why it changes geographically, how it interacts with the environment and human activity. The phrase comes from the Greek terms 'klima' (equivalent to latitude) and logos (talk or study). Climatology and meteorology are closely connected yet have different time scales. Climate is a dynamic, interrelated system that includes the atmosphere, land, snow, ice, oceans, and living organisms. The air component of the climatic system is often called 'average weather' Climate is the mean and variability of temperature, precipitation, and wind across long timeframes (the classical period is 30 years). Internal dynamics and external impacts (called 'forcings') transform the climatic system over time. External forcings include volcanic eruptions, solar variations, and human-caused atmospheric changes, and the whole climatic system is solar-powered. There are three ways to alter the Earth's radiation balance: by changing the amount of solar radiation received (e.g., by changing the Earth's orbit or the Sun); by changing the fraction of solar radiation reflected (called 'albedo'); and by changing the long wave radiation emitted by the Earth back towards space (e.g., by changing greenhouse gas concentrations).

The observation and measurement of solar radiation, temperature, humidity, evaporation, cloudiness, precipitation, visibility, barometric pressure, and winds are required to comprehend the causes of heat exchange, moisture exchange, and fluid motion across time and space. Energy and mass transfers across latitude, height, land and water surfaces, mountain barriers, and local topography by prevailing winds, air masses, and pressure centers cause global and seasonal climate variation.

Applied Climatology examines the relationship between climate and other phenomena, their effects on human, and the possibility of manipulating the climate to satisfy human requirements. Applied climatology highlights the interconnection of many fields and the usefulness of climate data. New multidisciplinary climatology fields include bioclimatology, agroclimatology, medical climatology, building climatology, and urban climatology.

1.1. Scales of Climate

Scales of Climate may be written as:

- Microclimate is the local climate at a particular point location (e.g., a climate station). For instance, a long-term weather or temperature station.
- Topoclimate is the climate of a certain location (e.g., a valley or hill-slope). For example, the long-term climate of a river basin. In this context, we recommend [1,2] to the readers.
- Mesoclimate is the climate of a region (e.g., southern Oregon)
- Synoptic Climate is the climate of a large area (e.g., a continent).
- Global Climate is the climate of the planet.

1.2. Climate and Weather

Climate is what we expect, while weather is what we encounter. Climate is a probability distribution (PDF) over which the weather oscillates. Climate is determined by the properties of the Earth system and can be thought of as a boundary value problem (see this terminology [3,4] for more details), whereas weather is highly dependent on the system's evolution from moment to moment and can be thought of as an initial value problem (see this terminology [3,4] for more details).

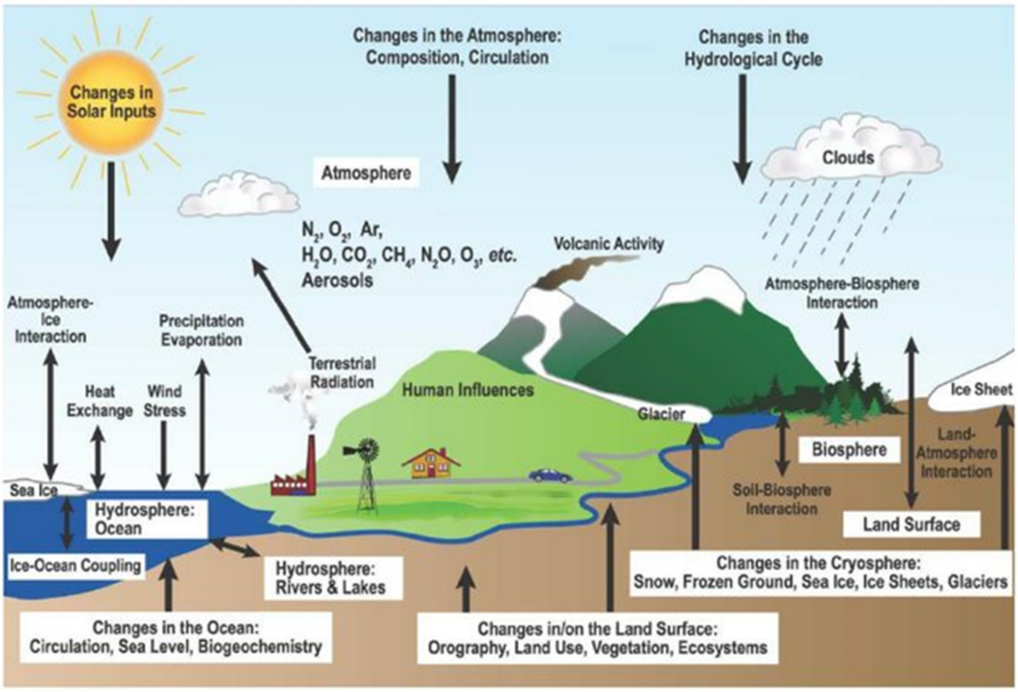

Figure 1. Schematic view of components of global climate system [5].

Weather forecasting is useful or possible for a few days, at most a week. Climate forecasting, on the other hand, is constrained by modeling capability and insufficient observation. This is not a failure of science, but rather of our capacity to comprehend the inherent properties of climate. Thus, climate simulations are constrained by gaps in our understanding of the hydrologic cycle, which encompasses cloudiness, ocean behavior, and small-scale processes, as well as the inherent unpredictability of various aspects of climate. Beyond several days, a fundamental dynamical property of the atmosphere imposes a significant constraint on weather forecasting.

Earth's climate is highly variable, affecting both the atmosphere and oceans, as well as the cryosphere and biosphere (see figure 1). The cryosphere, which includes the ice sheets of Greenland and Antarctica, continental (including tropical) glaciers, snow, sea ice, river and lake ice, permafrost, and seasonally frozen ground, is a critical component of the climate system. The cryosphere is critical to the climate system for a variety of reasons,

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including its high reflectivity (albedo) for solar radiation, its low thermal conductivity, its high thermal inertia, its ability to influence ocean circulation via freshwater and heat exchange and atmospheric circulation via topographic changes, its large potential for affecting sea level via land ice growth and melt, and its ability to influence greenhouse gases.

2. Radiation and Energy Balance in Atmosphere and Earth's Surface

On an average, 49 % (168 $Watt/m^2$) of total incoming solar radiation is absorbed by the Earth's Surface (see figure 2-3). This heat is radiated back into the atmosphere in three ways: as sensible heat, as evapotranspiration (i.e., latent heat), and as thermal infrared radiation. The majority of this radiation is absorbed by the atmosphere, which then radiates in both directions. The radiation lost to space originates in cloud tops and other regions of the atmosphere that are significantly colder than the surface. As a result, a greenhouse effect occurs. During the day, approximately 342 $Watt/m^2$ of energy reaches the top of the Earth's atmosphere. Approximately 30% of sunlight reaching the top of the atmosphere is reflected back to space. Approximately two-thirds of this reflectivity is due to clouds and small particles called 'aerosols' in the atmosphere. The remaining one-third of sunlight is reflected by light-colored areas of the Earth's surface, primarily snow, ice, and deserts.

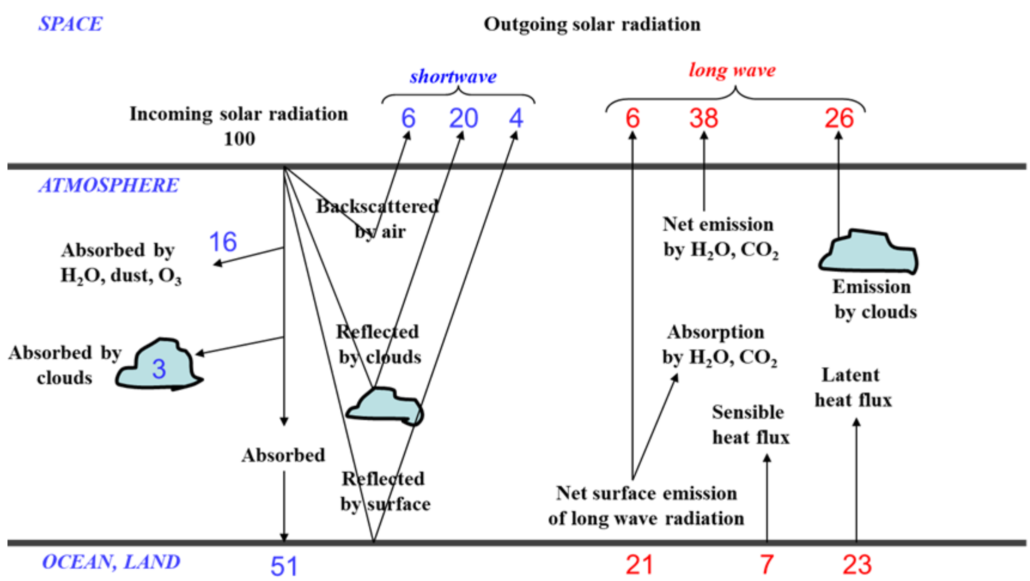

Figure 2. Radiation and Heat Balance in Atmosphere and Earth's Surface (global average components of the earth's energy balance).

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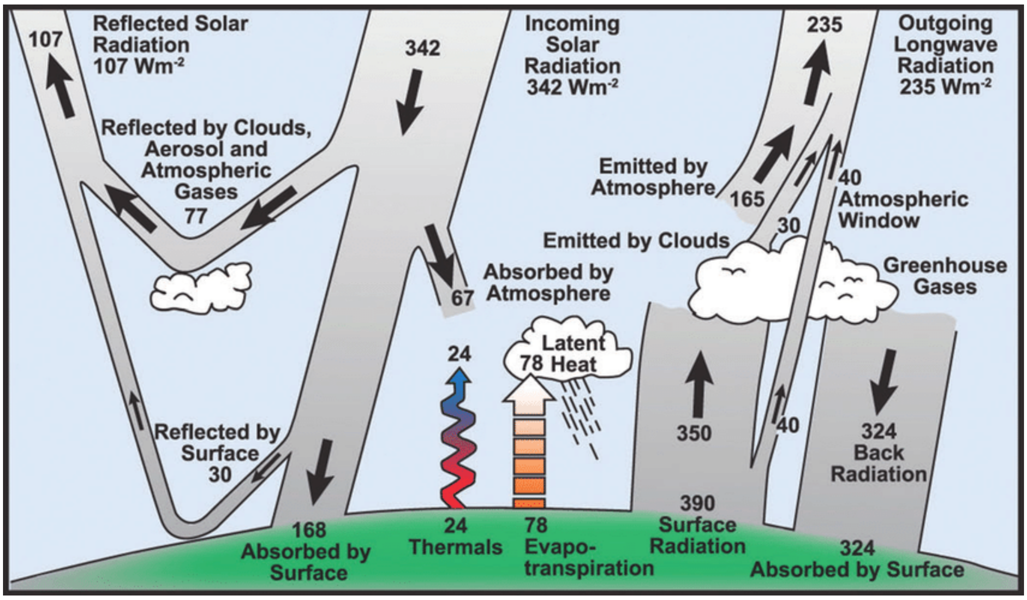

Figure 3. The Earth's annual and global mean energy balance [6].

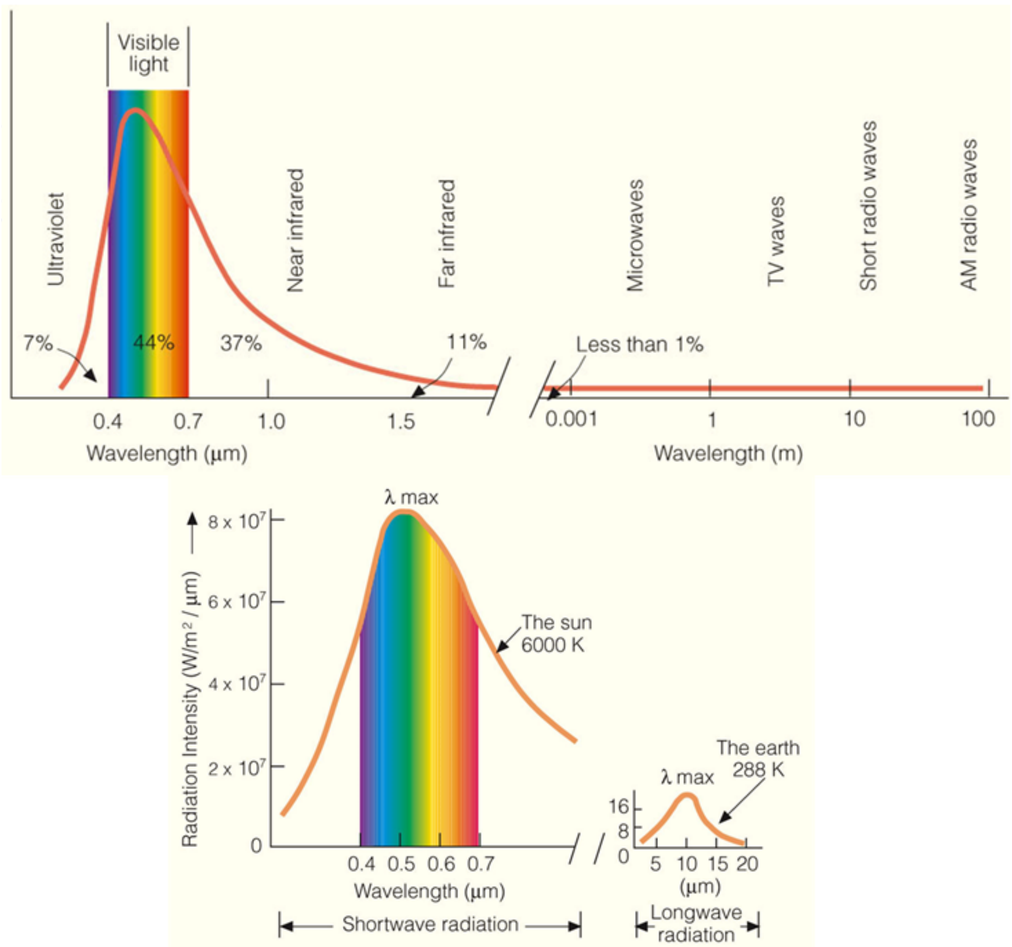

Figure 4. Electromagnetic spectrum of the Sun [7].

When large volcanic eruptions eject material very high into the atmosphere, the most dramatic change in aerosol-produced reflectivity occurs. While rain typically removes aerosols from the atmosphere within a week or two, when material from a violent volcanic eruption is projected well above the highest cloud, these aerosols typically influence the

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climate for about a year or two before falling into the troposphere and being carried to the surface by precipitation. Thus, major volcanic eruptions can result in a drop in the mean global surface temperature of approximately half a degree Celsius, which can last months or even years.

The remaining energy is absorbed by the Earth's surface and atmosphere. This equates to about 240 $Watt/m^2$. To compensate for the incoming energy, the Earth must, on average, radiate the same amount of energy back into space. The Earth accomplishes this through the emission of long-wave radiation. Everything on Earth continuously emits long-wave radiation (see figure 4). To emit 240 $Watt/m^2$, a surface must be around –19$°C$ in temperature. This is significantly colder than the actual conditions at the Earth's surface (the global mean surface temperature is approximately 14$°C$). Rather than that, the required –19$°C$ is found approximately 5 km above the surface. The Earth's surface is this warm because of the presence of greenhouse gases, which act as a partial blanket for the long-wave radiation emitted by the surface. The term "natural greenhouse effect" refers to this blanketing.

2.1. Energy Balance

After accounting for all energy inputs and outputs to and from the hydrologic system, the difference represents the rate of change of storage, just as was done for mass balance continuity. Sensible heat is the portion of a substance's internal energy, e, that is proportional to its temperature, $T°C$; $e = c_pT$, where c_p denotes the specific heat at constant pressure. Latent heat transfers occur at phase transitions, as indicated by vertical energy jumps during melting, sublimation, and vaporization.

2.2. Heat Transport Process

The heat transfer process occurs in three distinct ways, Conduction, Convection and Radiation. In conduction the molecules in higher temperature zones collide and transfer energy to molecules in lower temperature zones. It commonly measure by using Fourier's equation: $f_h = -k(\frac{dT}{dz})$, where k is the heat conductivity. In convection, heat energy is transported via the mass motion of a fluid. Mathematically, $f_h = -\rho cpK_h(\frac{dT}{dz})$, where K_h is the diffusivity. In radiation the energy transfer via electro-magnetic waves (which may occur in vacuum) at rates determined by their surface temperature. Stefan-Boltzman equation provides a way to measure radiation $R_e = E\sigma T^4$, where E is the emissivity of the surface, which is equal to one for a blackbody and 0.97 for water; Stefan-Boltzman constant $\sigma = 5.67 * 10 - 8W/m2.K4$, Wien's wavelength equals $\lambda = 0.0029/T$ meter, indicating that the Sun emits shorter wavelengths than the cooler Earth.

2.3. Radiation Balance at the Surface

The addition of aerosol, dust, and other pollutants to the atmosphere results in the greenhouse effect, in which some of the radiation emitted by the earth is reflected back by the atmosphere (see figure 5), resulting in an overall warming of the Earth. albedo $0 \leq \alpha \leq 1$, i.e., the fraction of radiation reflected, 0.006 for deep water, 0.9 for snow (see [8] for details).

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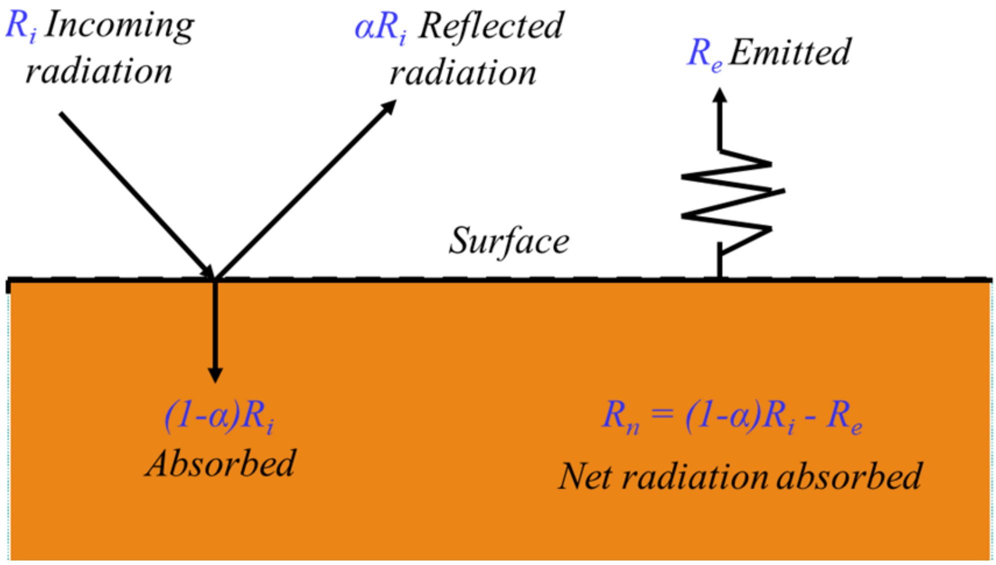

Figure 5. Radiation Balance at the Surface.

3. Earth's Atmosphere and Climate

3.1. Earth's Atmosphere

The atmosphere is extremely thin in comparison to Earth, reaching a maximum distance of approximately 560 *km* from the planet's surface. The structure is divided into four distinct layers (see figure 6): The troposphere, which extends from 8 to 14.5 *km* and accounts for approximately 90% of the atmosphere, has a temperature range of about 17 to -52°*C*; at the top of the troposphere, or tropopause (a thin layer), a cold trap causes condensation of the majority of the water vapor, forming clouds below. Stratosphere the dry atmosphere extends up to 50 *km* in height, with temperatures dropping to -3°*C* due to UV absorption; 99 % of the atmosphere remains below; stratopause separates the next layer. The mesosphere extends up to 85 *km*, the temperature drops to -93°*C*, the chemicals are excited due to the Sun's absorption of energy; the mesopause separates the next layer. The thermosphere reaches a height of 560 *km*, temperatures can reach 1727 °*C*, and chemical reactions occur much more rapidly than on the Earth's surface. Beyond the Atmosphere, the Exosphere begins at the thermosphere's top and continues to the interplanetary gases, or space. Hydrogen and Helium are the primary components of this region of the atmosphere and are only present at extremely low densities. The greenhouse effect, which keeps us from freezing, is caused by changes in atmospheric temperature with height (see figure 6).

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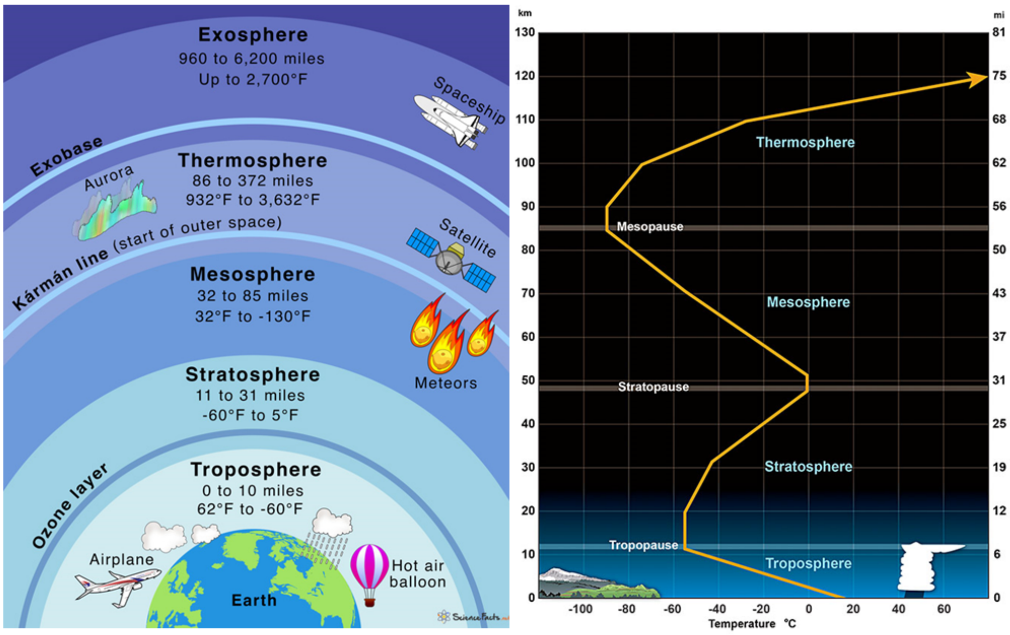

Figure 6. Atmosphere and Vertical Thermal Structure of the Earth [9].

3.2. Earth's Climate

The climate of the Earth is the result of the interaction of numerous properties and processes: solar radiation and orbital geometry, the planet's size, gravitational force, and rotation rate. The composition of the atmosphere, its circulation, and the hydrologic cycle Ocean properties and circulation, hydrology, biology, and geochemistry of the land surface, and the geography of continents, glaciers, mountain ranges, and oceans, among other things.

4. Earth's climatic processes

4.1. Heat Transport

There is a difference between incoming and outgoing radiation that varies with latitude. In specific, there is net radiative heating near the equator, and net radiative cooling near the poles (see in figure 7). This imbalance is compensated for by pole-to-pole energy transport. Thus, a fundamental coupling exists between the earth's radiation budget and the general circulation of the oceans and atmosphere. Ocean currents, particularly in the subtropics, transport heat pole-ward in both hemispheres. Heat is transported through the atmosphere in two ways: sensible heat (associated with the air parcel's temperature) and latent heat of the water vapor contained within the air parcel; this latent energy (2.5 kJ/gm water) is carried by evaporated water vapor until it is released into the atmosphere via vapor condensation. The atmosphere's latent heat transport contributes significantly to the earth's heat balance; it is a critical link between the hydrologic cycle and the global energy balance.

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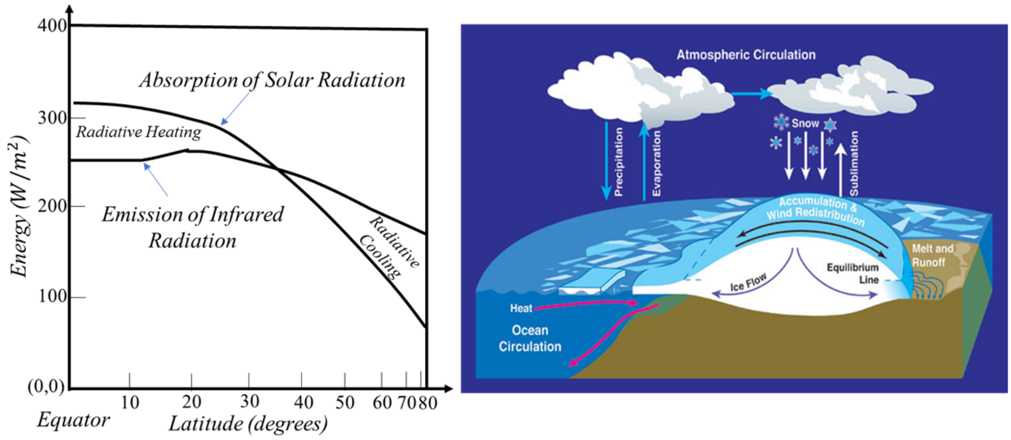

Figure 7. Absorbing solar radiation and emitting long-wave radiation averaged over latitude and Atmospheric circulation.

4.2. Atmospheric circulation

Three are types of atmospheric circulation, Low-Level Circulation, Upper-Level Circulation, and General Circulation.

4.2.1. Low-Level Circulation

This type of circulation displays significant regional variation in the mean low-level circulation, but a distinct pattern when averaged zonally across a latitude circle. Over the Atlantic and Pacific, the circulation pattern is most prominent. As latitude increases, a low pressure belt associated with extratropical storm tracks develops, concentrating eastward moving cyclone (low pressure) and anticyclone (high pressure) weather systems. In the Northern hemisphere, where the contrast between ocean and continent is greater, there are substantial seasonal variations in the atmospheric circulation. In the summer, subtropical oceanic highs expand northward and a region of relatively low pressure develops across south Asia. Wintertime brings storminess to the North Atlantic and North Pacific. Outside of the tropics, the earth's rotation is a key factor in determining the circulation-pressure relationship. Flow in a non-rotating system is primarily a direct response to the pressure gradient and is directed away from areas of high pressure and toward areas of low pressure. However, the planet's rotation produces an additional apparent Coriolis force, resulting in a substantial geostropic component of flow. This takes the form of a clockwise circulation around a high-pressure system in the northern hemisphere and a counterclockwise circulation around a low-pressure system in the southern hemisphere.

4.2.2. Upper-Level Circulation

Due to the thermal wind effect, the atmospheric circulation generally intensifies and simplifies as it ascends through the troposphere. Pressure decreases with height more rapidly in colder air than it does in subtropical air. Which results in a vertical pressure gradient along the north-south axis with height. Due to the geostropic relationship between pressure and circulation with altitude, in the subtropics and middle latitudes, strong westerlies develop with altitude. The jet stream is the rise of the strongest westerlies in the upper troposphere. Jet streams are constantly evolving, meandering, and decaying as part of the upper troposphere's wavy planetary-scale circulation systems. Upper tropospheric circulation characteristics are associated with the extratrophics' ever-changing regional weather regimes.

4.2.3. General Circulation

It was invented for the first time in 1735 by George Hadley. Hadley reasoned that the equatorward direction of low level easterly winds in the vast oceanic trade-wind belts

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of the subtropics reflects the lower branch of a hemisphere-wide thermal convection cell, with rising motion in the warm equatorial belt and sinking motion in the subtropics and colder temperature zones. The difference in solar heating between the tropics ($23^{0}27'$ N and S) and high latitudes would drive such a circulation. Hadley regarded transient circulations associated with day-to-day weather variation, as well as mean differences in the circulation around a latitude band, and the annual cycle of temperature and other metrological elements, as irrelevant "ornaments of the circulation" that are uncoupled from the fundamental processes of the general circulation. Even though mean low-level circulation varies significantly across regions, when data are averaged around a latitude circle, a distinct pattern (mean meridional) emerges (see figure 8).

The air entering from lower latitudes is generally warmer and will rise in order to cool. As some of this air sinks near the subtropics and returns to the equator, it produces what are known as trade winds, which were named after sailing ships used in foreign trade between Europe and the New World. The region near the equator where these winds cease to exist is referred to as the doldrums. Hadley cells are areas where air rises at the equator, sinks at 30° north and south, and then flows back to the equator. Although the trade winds are most distinct over the Atlantic and Pacific, they are frequently influenced by monsoon circulation over the Indian Ocean – Indonesian sector. While the majority of air sinks at 30° north and returns to the equator at 30° south, some air continues to move poleward. The belt of subtropical high pressure is located between 30° north and south latitudes; as one moves higher in latitude, a low-pressure belt associated with the extratropical storm track (between 40° and 60°) is formed, concentrating eastward-migrating cyclonic (low pressure) and anticyclonic (high pressure) weather systems. Ferrel cells are circulation cells that form between 30° and 60° north and south. At approximately 60° north and south, the air collides with cold polar air to form polar fronts; some of the air that rises at the polar fronts continues to move poleward, sinking at the poles and then returning to 60° north and south. There is a tendency for elevated pressure to build up over the cold polar cap. Hadley cells (see figure 8) are more robust than cells at higher latitudes. The Hadley circulation cannot be adequately described by an annual average; with the exception of brief periods during the equinox seasons (spring and autumn), the tropics' mean meridional circulation is dominated by a single summer hemisphere Hadley cell. The rising air zone moves seasonally, being in the SH from December to February (southern summer) and in the NH from June to August (northern summer). Outside of the tropics, the components of the mean meridional circulation are quite weak; they are comprised of weak indirect Ferrel cells and even weaker direct polar cells. Outside of the deep tropics, the earth's rotation dictates the relationship between circulation and pressure; in a non-rotating system, flow is largely a direct response to pressure gradients (high to low). Because the air that sinks at certain locations flows back along the earth's surface in a non-linear north-south path due to the Coriolis Effect, this results in a strong geostrophic component of flow. This manifests as a CW circulation around high-pressure (anticyclonic) systems and a CCW circulation around low-pressure (cyclonic) systems in the New Hampshire. In SH, the inverse relationship exists. Due to the thermal wind effect, atmospheric circulation generally intensifies and simplifies upward through the troposphere; there is an increasing N-S horizontal pressure gradient (pressure decreases more rapidly with height in the north than in the south) with height. Stronger westerlies develop with height in the subtropics and middle latitudes due to the geostrophic component; this axis of strongest westerlies in the upper troposphere is called the jet stream (speed: 160 km/hr). These upper tropospheric circulation characteristics are associated with the extratropics' ever-changing regional weather regimes; these synoptic characteristics are implied by the relatively low consistency of wind direction in mid-latitudes.

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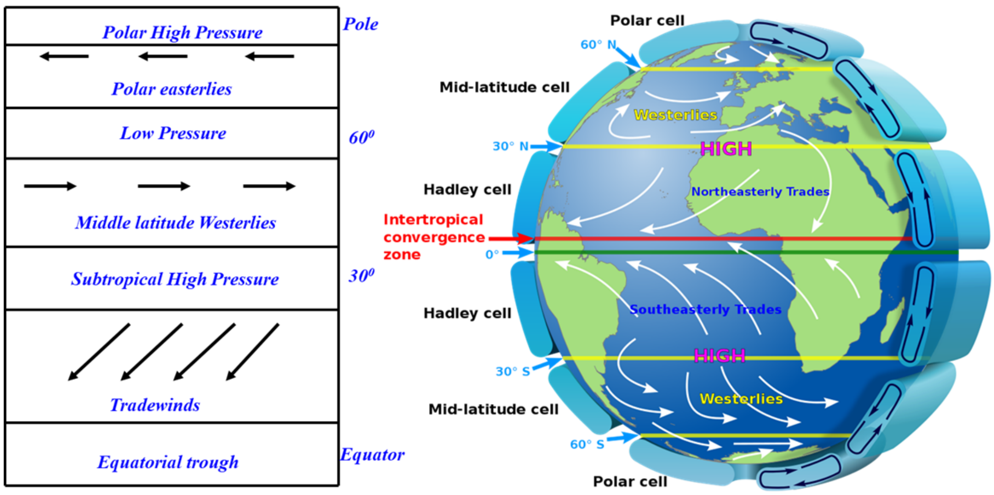

Figure 8. Representation of the wind and pressure belts at the surface [10].

The mean meridional (latitude averaged) circulation of the atmosphere between December and February, and between June and August are shown in figure 9; the values on the streamlines indicate the total mass circulation (1010 kg/s) between that streamline and zero streamline.

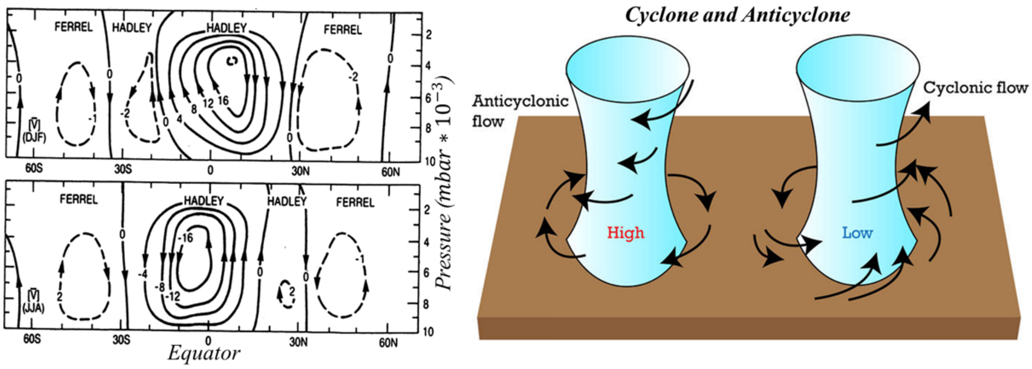

Figure 9. Mean meridional circulation of the atmosphere and cyclones.

4.3. Global Pressure, surface wind speed, and sea surface temperature (SST)

During northern summer, oceanic subtropical highs develop rapidly and expand northward; relatively low pressure develops over warm continents, most notably South Asia, where the monsoon low is well developed. During northern winter, a seasonal reversal of pressure between continents and oceans is observed; storminess increases over the North Atlantic and North Pacific as a result of pronounced low-pressure areas in northern latitudes and weak high-pressure areas in the subtropics (see figure 10).

71% of the earth's surface is covered by ocean water. Due to the fact that water has a greater heat capacity than land, the ocean serves as the primary heat storage or memory component of the climate system. Annual mean SST varies between 29°C in portions of the tropics to -1.8°C near the ice edge.

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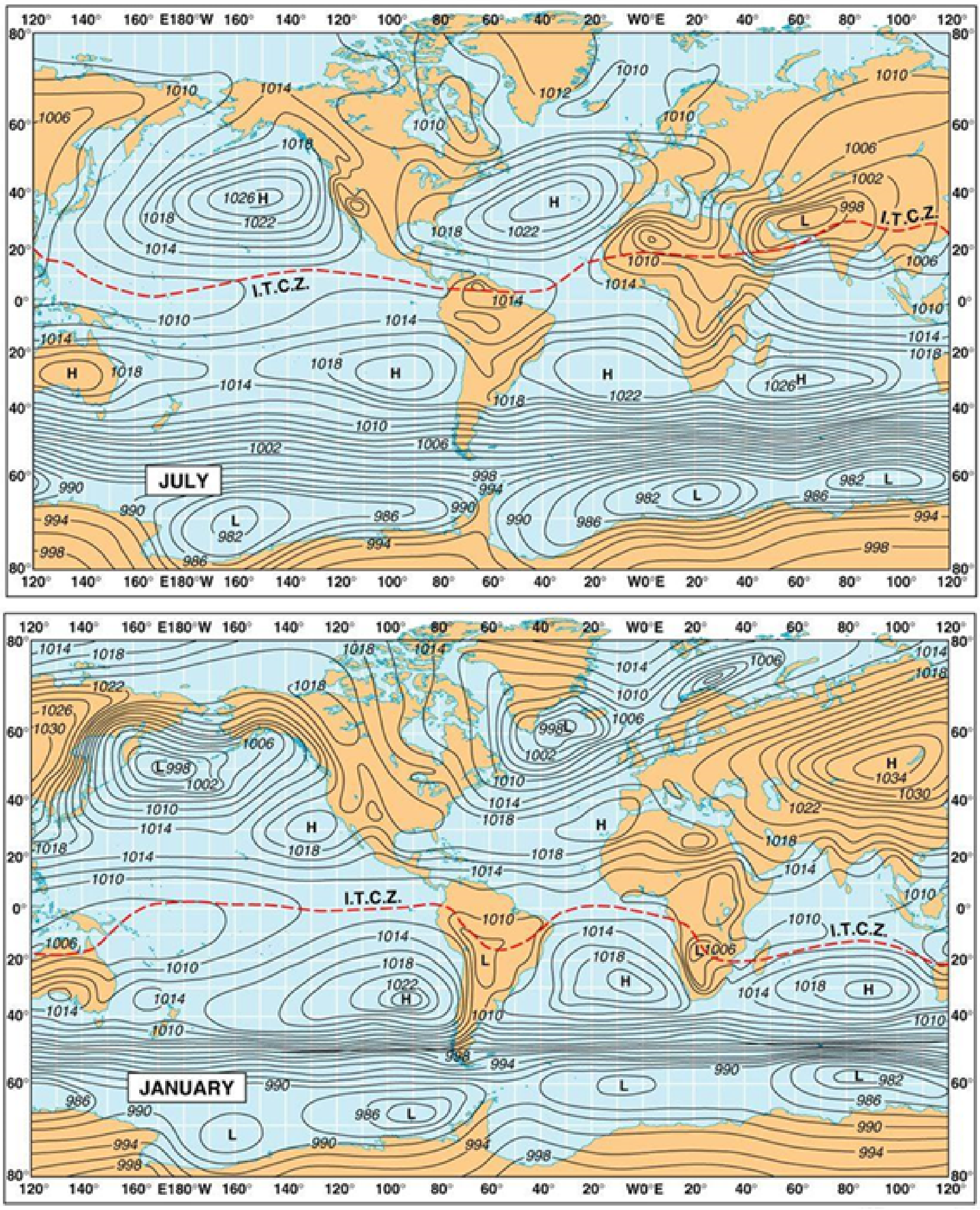

Figure 10. Global Pressure and surface wind for July and January.

4.4. Ocean Temperature and Salinity Profile

Because the ocean is heated and cooled predominantly from its upper surface, it has a thermal structure distinct from the atmosphere. Oceans are separated into upper and deep zones based on their temperature. Within the upper zone, a mixed layer exists where the temperature is similar to that of the surface. Below the mixed layer lies a transition layer called the thermocline, in which the temperature rapidly lowers. Temperature falls extremely slowly in the deep ocean. At mid-latitudes, a seasonal thermocline forms (see figure 11).

SST isotherms run broadly east-west across the majority of the ocean, but diverge towards the coasts and around the equator, where currents and upwellings influence temperature distribution. Ocean and atmospheric circulation systems are linked by energy and momentum exchanges at the air-sea interface. At its lower limit, the atmosphere absorbs energy via SST, whereas surface wind generates wind-driven circulations in the upper ocean. The ocean responds to surface heat fluxes and evaporation and precipitation,

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which alters the ocean's temperature and salinity and resulting in density variations that drive the ocean's deep thermohaline circulations (see figure 12).

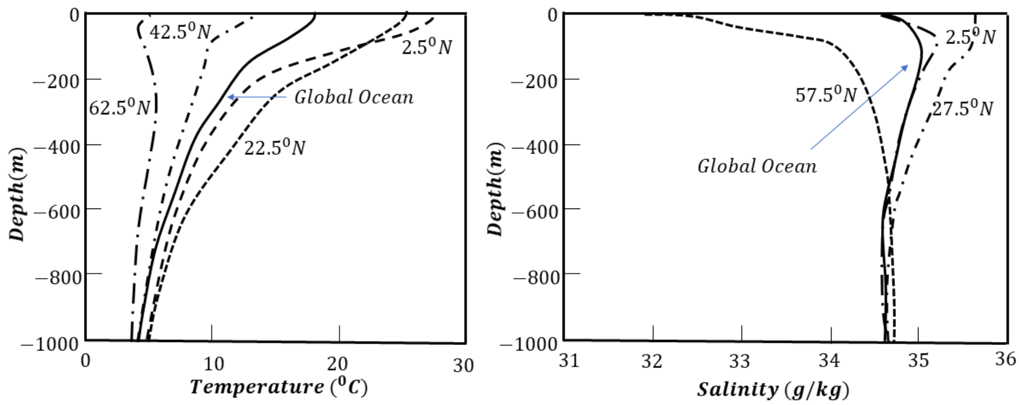

Figure 11. Ocean temperature and salinity profile.

4.5. Wind Driven Circulation and Ocean Currents

Wind-driven circulation is predominantly horizontal and is confined to the ocean's upper few hundred meters. The Antarctic circumpolar current is continuous around the latitude circle at the southerly latitudes of SH, where there are strong and persistent westerlies and no land barrier. Outside the deep tropics, the Atlantic and Pacific Oceans' largest-scale wind-driven circulation systems are the basin-scale subtropical anticyclonic gyres, which form as a result of prevailing winds, continental boundaries, and the earth's rotation. They include the following: Warm poleward current systems on the west side of ocean basins that are quite narrow and intense (transport heat): Gulf Stream in the North Atlantic, Kuroshio Current east of Japan. Currents on the east side of ocean basins that are cold and upwelling (carry cooler water): California Current in the North Pacific, Peru Current in the South Pacific. The equator's primarily east-west or zonal circulation regimes are significantly different from those seen at higher latitudes; large shifts in these equatorial currents are quick and more in tune with year-to-year atmospheric variability known as the El Nino-Southern Oscillation (ENSO).

4.6. Thermohaline Circulation

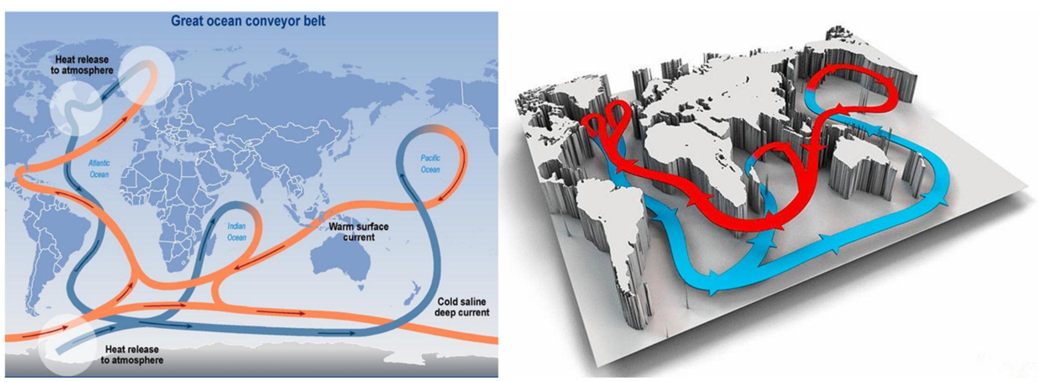

Figure 12. Great ocean conveyor belt [11,12].

Thermohaline circulation is the slow movement of water through oceans caused by density changes caused by temperature and salinity variations. It begins in polar latitudes as a vertical flow that sinks to the mid-depths or even lower, followed by horizontal flow. It is initiated by an increase in density at the upper surface, either directly through cooling and/or salinity increases, or indirectly through ice melting and ejecting salt, thereby

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increasing the salinity of the remaining water. The global scale redistribution of ocean water by thermohaline circulations occurs on time scales ranging from decades to centuries, with a typical cycle lasting approximately 1600 years, i.e., water from the surface to the deep ocean and back to the surface.

The great world ocean's water is constantly in motion. Currents that transport enormous amounts of water around the world have been dubbed The Great Ocean Conveyor (Broecker, 1991). Oceanic thermohaline circulation is what powers the Conveyor. It involves both heat, hence thermo, and salt, hence haline, for common table salt (halite). Temperature and salinity both affect the density of seawater, and the density differences between water masses cause the water to flow. Thermohaline circulation, also known as the Great Ocean Conveyor, results in the world's largest oceanic current. It operates similarly to a conveyor belt, thus the name, transporting massive amounts of cold, salty water from the North Atlantic to the Northern Pacific and returning with warmer, fresher water. Typically, descriptions of the Conveyor's operation begin with what occurs in the North Atlantic, beneath and near the polar region's sea ice. There, warm, salty water transported from tropical regions is rapidly cooled, resulting in vast quantities of frigid water. When this seawater freezes, the salt content of the water is removed (sea ice contains almost no salt), increasing the salinity of the remaining, unfrozen water. The salinity of the water makes it quite dense, and its frigidity makes it even denser. Due to the fact that this water is denser than the less saline, warmer surface waters moving in from the south, it sinks to the ocean floor. Oceanographers refer to this water as North Atlantic Deep Water (NADW), and it is responsible for today's oceanic thermohaline circulation. This water begins its great circuit through the world's oceans in the northernmost reaches of the North Atlantic. It travels south through the North Atlantic, then south through the South Atlantic, rounding Brazil, where it meets vast masses of similarly frigid and saline water flowing from beneath the sea ice surrounding Antarctica (dubbed Antarctic Bottom Water [AABW] or Antarctic Deep Water [AADW]), hugging the ocean bottom as it flows. This greatest of ocean currents then moves east, well north of the Antarctic mainland but well south of Africa (where a branch pushes northward along the east African coast past the Cape of Good Hope) and continues east across the entire width of the Indian Ocean north of Antarctica, swinging around south of Australia and far into the Pacific. As it continues its submarine migration, the current mixes with warmer water, warms, and rises, until it finally dissipates as a coherent entity in the northern Pacific. However, in the Pacific, a warm, shallow-sea counter-current has formed. This Counter-Current travels south and west through the Indonesian archipelago, across the Indian Ocean, continuing westward, and circles southern Africa just off the Cape of Good Hope. It passes through the South Atlantic, still on the surface (though it extends a *km* and a half below the surface), where tropical warmth increases evaporation, making the counter-current saltier. It then travels up the East Coast of North America and across the Atlantic to the coast of Scandinavia, where its warmth helps protect residents from the bitter cold of northern winters. When this saltier, warmer water reaches high northern latitudes, it cools and eventually transforms into North Atlantic Deep Water, completing the circuit.

5. Coupled Ocean and Atmosphere Processes

5.1. Annual Cycle and Monsoon Circulation

Annual climate cycle amplitude is significantly greater in the NH (60% ocean) than in the SH (80% ocean), because the seasonal cycle of land surface temperatures is moderated by ocean heat storage during the summer and release during the winter (via poleward transport). During the winter, high-latitude continents propel cold air masses southward and eastward from Asia and North America, bringing them to the warm waters of the western Pacific and Atlantic via transient cyclones and associated cold fronts. The heat fluxes from the oceans modify cold continental air masses as they travel eastward across the oceans, with the air assuming maritime characteristics well before it reaches the west coast of North America or Europe. While maritime influence extends far inland in Europe

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(there are no mountain barriers), it is largely confined to the area west of major costal mountain barriers in North America.

The annual cycle of land surface temperature has a greater amplitude than the tropical sea surface temperature cycle (SST). The annual cycle of land-ocean surface temperature difference, combined with the seasonal reversal of the SST difference between the hemispheres, is what drives the tropics' atmospheric monsoon circulations.

The Asian-Australian monsoon system is the world's dominant monsoon circulation. During the winter, there is a low-level flow of dry and cool air from the cold continent to the warm ocean, resulting in a light precipitation over land. During the summer, moisture flows from the tropical ocean to the warmer land, where upward motion of the heated air produces monsoon rains. The monsoon component of atmospheric surface circulation indicates the monthly mean surface circulation's departure from its annual mean value. Similar but less pronounced monsoon-type circulations also occur over western Africa, parts of Mexico, and Central America, extending as far north as the southwestern United States. Additionally, the $28°C$ and $27°C$ SST isotherms for the respective hemisphere's mid-summer month are shown; hatched areas indicate significantly higher precipitation than the average for all months. Apart from these shared characteristics, regional monsoon precipitation regimes vary significantly.

5.2. Tropical Cyclones

Each of the tropics' transient synoptic features, depressions and storms that comprises a cloud cluster, an area of intense convection. Cloud clusters work in concert to produce the large-scale precipitation patterns associated with monsoons and oceanic convergence zones. Tropical cyclones (with a warmer central core) are the most intense transient weather phenomena in the tropics; a storm reaches storm intensity when sustained winds surpass 17.5 m/s, while a hurricane reaches hurricane intensity when sustained winds exceed 33 m/s. The following three conditions must exist in order for powerful tropical cyclones to form:

- A warm ocean surface (min $26°C$ to $27°C$) is necessary to provide required fluxes of water vapor and sensible heat from ocean to atmosphere.
- Since strong rotation generates in regions of significant Coriolis force, these storms form beyond about $5°$ to $8°$ of the equator.
- A small change of wind with height is required if storm is to survive.

Tropical storms are more likely to form in the summer and fall over locations with a sea surface temperature of 27.5°C. Tropical storms do not form in the South Atlantic and eastern South Pacific due to the region's comparatively cool sea surface temperature. The ENSO cycle has an effect on tropical storm activity because it is connected with changes in both sea surface temperature and vertical wind shear.

6. El Niño-Southern Oscillation (ENSO)

El Niño, in its original sense, is a warm water current which periodically flows along the coast of Ecuador and Peru, disrupting the local fishery. This oceanic event is associated with a fluctuation of inter-tropical surface pressure pattern and circulation in the Indian and Pacific oceans, called the Southern Oscillation (SO). This coupled atmosphere-ocean phenomenon is collectively known as El Niño-Southern Oscillation, or ENSO. During an El Niño event, the prevailing trade winds weaken and the equatorial countercurrent strengthens, causing warm surface waters in the Indonesian area to flow eastward to overlie the cold waters of the Peru current. This event has great impact on the wind, sea surface temp and precipitation patterns in the tropical Pacific. It has climatic effects throughout the Pacific region and in many other parts of the world. The opposite of an El Niño event is called La Niña.

The Christ Child was first recognized as a warm surface countercurrent flowing down the Ecuadorian and Peruvian coasts around Christmastime, and thus was named by local fishermen. Discovered for the first time in 1795. El Niño were most recent and severe

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in 1953, 1957–58, 65, 72–73, 76–77, 82–83, 87–88, and 97–98, 2002–03. El Niño is cyclical, implying it occurs in cycles. Jacob Bjerknes (1969) demonstrated using the most recent wind, rain, and sea surface temperature (SST) data that the Southern Oscillation and El Niño are NOT distinct phenomena - abbreviated ENSO. El Niño's effects are not limited to Peru, but could affect the entire Pacific and even the entire world.

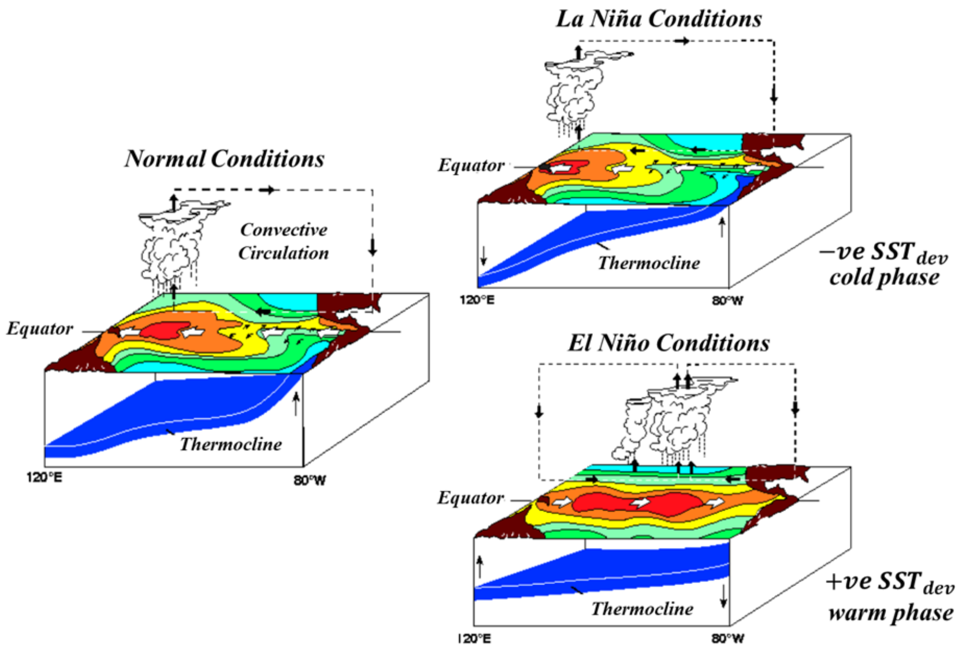

Figure 13. El Niño-Southern Oscillation (ENSO) [13].

During a La Nina event, eastern Pacific remains colder than western Pacific, atmospheric pressure in east is higher than west, so wind blows (Trade Wind) from east to west and push warm surface water and vapor to west, and Heated rising air over western Pacific causes rainfall in the west. On the other hand, during an El Nino event, eastern Pacific becomes warmer, trade Wind becomes weak and an opposing wind from west to east begins, so wind blows from west to east and push warm surface water and vapor to east, and heated rising air over eastern Pacific causes rainfall in the east (see figure 13).

7. Aspects of Land Surface Climate

7.1. Planetary boundary layer

The air close to the ground is more turbulent than the air at higher altitudes. Low-level turbulence is least during the night, occurring solely as a result of mechanical stirring of rough surfaces, and is highest in the afternoon, occurring as a result of heat convection and hence more vigorous stirring. Because atmospheric pressure falls with altitude, stirring an atmospheric layer cools increasing parcels of air by adiabatic expansion and warms sinking parcels via compression. The temperature decreases at a rate of $9.8°C$ per second, which is referred to as the adiabatic lapse rate (g/C_p), where the Cp specific heat of the atmosphere at constant pressure equals 1004 $J/°K/Kg$. The atmospheric layer immediately above the surface is sufficiently mixed to sustain a lapse rate close to or slightly greater than the adiabatic; this is referred to as the planetary boundary layer. Above the planetary boundary layer, the atmosphere is generally very stable (i.e., the lapse rate $<$ is adiabatic) due to latent heat released during condensation and precipitation, as well as upward heat transmission via large-scale atmospheric motion. In high-pressure zones over land, the planetary boundary layer is composed of three principal components: 1) a highly turbulent mixed layer, 2) a less turbulent residual layer containing former mixed-layer air, and 3) a

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nighttime stable boundary layer with intermittent turbulence; the mixed layer is further differentiated into a cloud layer and a subcloud layer (see figure 14).

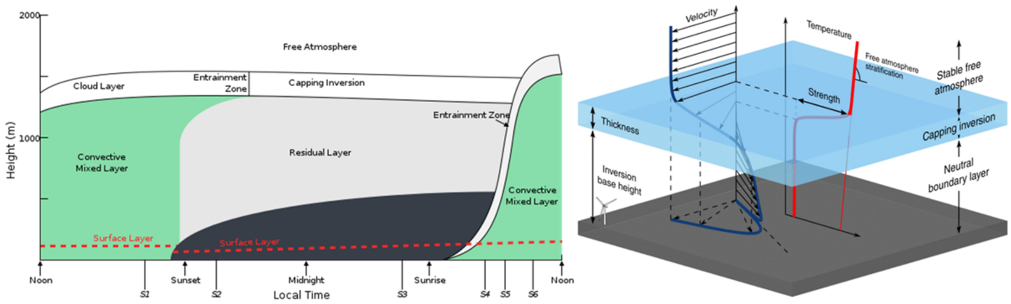

Figure 14. Planetary boundary layer [14].

8. Climate Variability and Climate Change

The 1930s and 1950s US Great Plains droughts, as well as the 1960s Sahelian rainfall deficit, are complicated by insufficient data records; dendrochronology (tree-ring chronology) is used to extend historical data series. Long-term variations in precipitation caused by natural processes will remain a source of contention until the governing processes are better understood. Numerous characteristics of multi-year variability are derived from global SST variations (internal dynamics). External factors such as volcanoes and solar variability may also be involved (though their significance is still debated); associations between drought and the 11-year sunspot cycle or the 22-year solar magnetic variation have been difficult to confirm.

8.1. Small-Scale Climate Variability

Due to a variety of local controls, the climate of a small area can differ significantly from that of the larger surrounding region. Local differences in terrain, land surface characteristics, and air pollution all affect airflow, cloudiness, temperature, and even precipitation via their effects on surface roughness, heat and water balance. Topography influences both mesoscale and microscale climate variations; regionally, mountain ranges force ascent for wind, condensation, and heavy precipitation, while subsiding air is relatively dry, resulting in a rain shadow effect thousands of *km* below; for example, the Rocky Mountains channel large-scale outflows of cold air from the polar regions; elevation affects temperature and precipitation type due to atmospheric lapse rate (which decrease at the rate of $9.8°C/km$); elevation affects temperature and precipitation type At one extreme, the summer afternoon lapse rate over arid regions may be nearly dry adiabatic down to several thousand meters; at the other extreme, a strong low-level temperature inversion frequently forms at night over middle- and high-latitude continents during winter under clear skies and light winds.

Local wind systems are prevalent in a wide variety of environments; they can be gravity-driven in mountainous regions and off ice fields and glaciers; thermally-driven by differential surface heating; or mechanically-driven by isolated hills or mountains. Local thermal circulations are most prominent in the tropics and middle latitudes during the warmer months, when large-scale temperature gradients and circulation are weak; thermally driven diurnal wind systems include mountain-valley winds, land-lake and land-sea breezes, and urban-rural contrasts.

differs partly as a result of topographic bias (e.g., in valleys), but primarily as a result of distinct land characteristics and air quality which is a phenomenon known as the heat island effect, which results in a higher nighttime temperature.

Due to high heat capacity of water, seasonal changes in water temperature lag those on land; water is colder in the spring and early summer than land, and warmer in the autumn and early winter that resulting in seasonally varying land-lake wind effects.

8.2. Drought

Generally associated with a sustained period of decreased soil moisture and water supply in comparison to the normal levels that have stabilized the local environment and society. Humidity is a rare and disruptive feature, Semiaridity is common and frequently catastrophic, and Desert is a meaningless concept. Drought is defined by the following criteria: precipitation, evapotranspiration, streamflow, runoff, groundwater levels, water supply, and water needs.

The most frequently used definitions of drought are as follows: Drought is defined meteorologically as a period of time, typically months or years, during which the actual supply of moisture at a given location falls cumulatively short of the climatologically appropriate supply. Drought is defined in agriculture as a period of insufficient soil moisture to meet the evapotranspiration demand required to initiate and sustain crop growth. Hydrologic drought refers to periods of reduced streamflow and/or depleted reservoir storage. Economic drought is caused by physical processes, but it primarily affects the economic sectors of human activity. Understanding the factors affecting the time scales, amplitudes, and frequency of droughts provides insight into the physical processes (most notably ocean-atmosphere interaction) that validate drought regimes.

9. Concluding Remarks

The major purpose of this article is to acquaint engineers with the foundations of climate and the processes that govern it so that they are able to properly utilize this information to comprehend the climatic impact on water resources systems.

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