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Atmosphere-Ocean Interaction Effect on Energy Balance of the Earth

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Abstract: Earth's atmosphere is made of a combination of several different gases with different volume percentages and also has different temperature layers. Oceans are also composed of different layers of temperature, pressure and salinity. The incoming solar radiation flux passes through atmosphere before hitting earth's surface while part of this radiation is absorbed by it, part of it is dissipated and the rest reaches the ground. Oceans, also interact with the incoming solar flux in the same manner of absorption, dissipation and passing. The ocean-atmosphere interactions include exchanging of heat, mass (water) and chemical components. These exchange mechanisms have different scales such as thermal conduction between water and air, between atmosphere and the upper layers of the ocean and tropical storms. The sun is the essential source of energy for earth. While the energy exchange between atmosphere and the surface isn't quite balanced, the surface radiates more energy than it receives and so the atmosphere's energy gain is greater. The amount of this imbalance depends on the difference of sensible and latent heat coefficients for the atmosphere and the surface. The sensible heat transfer is characterized by the direct exchange of heat between the lowest atmospheric layer and the surface which is disturbed by sun and then mixed with the air.

Key words: Ocean-atmosphere, the incoming energy balance, solar radiation flux, dissipation, Iran

INTRODUCTION

Ocean-atmosphere interactions: The ocean-atmosphere Interactions include exchanging of heat, mass (water) and chemical components. These exchange mechanisms have different scales such as thermal conduction between water and air, between atmosphere and the upper layers of the ocean and tropical storms. The sun is the basic source of energy for earth. The solar radiation reaches the Earth's surface after passing through the atmosphere. Part of the passing radiation is absorbed at the surface of the sea and the remaining penetrates into the ocean. Light can penetrate several tens of meters down the oceans and allows the energy to be transferred to the upper layers of the sea. This heat supplied to the ocean is then transferred to the atmosphere through heat exchange or evaporation of the oceans and thus provide its water vapor supply (Belkin and Cornillon, 2003). The presence of surface heterogeneities such as oceanic fronts (the difference in surface temperature for more than several tens of kilometers) or ices in sea, causes horizontal Differences in ocean-atmosphere interactions. It also heavily intensifies wind friction and increases the heat transfer between ocean and atmosphere and accelerates the atmospheric turbulences of the smallest scales. Vertical movement of air or water is dependent on the horizontal movement (wind and current), temperature and vertical gradient of humidity. Parallel to this dependence, the wind friction also causes local events (waves, sea spray and wave breaking) which in return affect the exchange of energy between water and air. Additionally, the exchange of water (through evaporation and rain), gases and solid particles, also takes place between the ocean and atmosphere. The amount of gas exchange depends on wind speed and water temperature (regarding the solubility of gases) (Belkin and Cornillon, 2003). The particles emitted from the evaporation of sea water and sea spray ("primary" aerosol) are carried to the land surface by raindrops. Also the organic components such as sulfur and volatile organic components cause chemical reactions inside the clouds.

Solar radiation: The sun is the ultimate source of energy in the atmosphere ocean system. Observations show that the flux of solar energy received by the Earth (Q) is almost 1366 W m^{-2}. At any time, only half of the Earth faces the sun, so the planet receives the amount of solar radiation equivalent to a disc with an area of πa^2 while represents the radius of Earth which is on average, 6.37×10^6 m. With its surface area equal to $4\pi a^2$, the average global solar radiation flux then is:

$$\frac{Q\pi a^2}{4\pi a^2} = \frac{Q}{4} = 342\ \frac{W}{m^2}$$

Solar radiation is evenly distributed around the world. Due to the long distance between the sun and earth, the

Table 1: Gaseous components of the atmosphere (Updated by Jacobson, 2005)

Gas	Volume (%)
Nitrogen (N_2)	78.08
Oxygen (O_2)	20.95
Argon (Ar)	0.93
Water vapor (H_2O)	0.00001-4.0
Carbon dioxide (CO_2)	0.039
Neon (Ne)	0.0015
Helium (He)	0.0005
Methane (CH_4)	0.00018
Ozone (O_3)	0.000003-0.001
Krypton (Kr)	0.0001

solar rays hitting earth are almost parallel to each other, meaning the received solar radiation is identical for any place on earth. Earth's surface is curved, so the surface of the planet is at an oblique angle to the solar radiations. The solar flux received at locations with a perpendicular angle to the surface is larger.

Solar zenith angle differs with latitude, season and time of day. A larger zenith angle means that the same amount of solar radiation is spread over a larger area. This spreading of solar radiation by increasing the zenith angle is one of the reasons the tropics are warmer than the Polar Regions. All the solar radiation reaching the ground is absorbed. A fraction of this amount is then reflected back into space; this fraction is called albedo.

The amount of surface albedo for sunlight, depends on the type of surface (Table 1) (Blanc, 1987). Oceans absorb a large part of solar radiation, so the albedo of ocean surface is small. On the other hand, fresh snow reflects a large part of solar radiation, so the albedo of an area covered with fresh snow is high. Albedo also varies with the solar zenith angle; larger solar zenith angles cause greater albedo. Total albedo of the planet is calculated by adding the solar radiation reflected from the top layers of the atmosphere to the radiation reflected from the lower atmosphere.

Seasonal cycle: The change of seasons is caused by the distribution of solar radiation with latitude due to the geometry of Earth's orbit.

Earth orbits around the sun in an elliptical path. The closest position to the Sun (perihelion), occurs around January the 4th while the farthest position (aphelion) occurs about July the 4th. The distance between the Sun and Earth at the apogee is $(1+e)\bar{r}$ where e represents the eccentricity of Earth's orbit (currently equal to 0.017) and r is the average distance between the Earth and sun. Similarly, for the perigee we have $(1-e)\bar{r}$. The radiation flux reaching the earth is calculated using the following equation:

$$Q = \frac{S}{4\pi r^2}$$

where, S is the flux radiated by the sun. The solar radiation reaching the Earth at perihelion is 3.3% less and at aphelion it is 3.5% more than the average. The total radiation emitted during a year is largely independent of the eccentricity of the orbit. The eccentricity of Earth's orbit over long periods (100000-400000 years) is 0.005-0.060. In the absence of other factors, the seasonal cycle of solar radiation can be determined using the eccentricity of the orbit: the summer is related to perihelion and the winter occurs at aphelion (Wells, 2012). However, the most important factor for the earth's seasonal cycle is the zodiacal deviation (the angle at which the Earth's axis is tilted relative to the orbital plane) which is currently± 23.5 degrees.

Because of the eccentricity effect, the radiation received in an Australian summer (December to February) is greater than the northern summers (June to August), but the seasonal cycle is also affected by the declination angle of the sun. The declination angle of the sun varies from 23.5±S in northern winter solstice (December the 21th to 22th) to N±23.5 in the northern summer solstice (June the 20th to 21th) and for the vernal and autumnal equinoxes in equator regions (March the 20th to 21th and September the 22th to 23th, respectively) this declination oscillates between a minimum of 22± and a maximum of 24.5± with a period of 41,000 years. These changes cause a great impact on the seasonal cycle of solar radiation. The third geometric factor, affecting the seasonal cycle is longitude of the perihelion of ±360 over a period of 21,000 year, so that in about 10,500 years the Earth's closest location to the sun occurs in July and in January its location is the farthest from the sun. The average temperature of the surface is highly sensitive to seasonal cycles. This sensitivity means that changes in orbital parameters over long periods of time can have a great influence on global climate. Studying the changes of global average temperature over the past and reconstructing it for the last 420,000 years using ice cores at Vostok, indicate significant climate changes for every 21000, 41000, 1000000 year intervals (Petit *et al*., 1999). This cycle is called Milankovitch's cycle.

More than 80% of Earth's lands is located in the northern hemisphere and the Southern Hemisphere contains about 63% of the area of global oceans and for about 60°S, there is almost no land on this hemisphere. Western hemisphere's asymmetry in lands and oceans is reversed in high latitudes. By containing Antarctica and its massive ice sheets, the Southern Hemisphere's surface temperature is very lower throughout the year. The Arctic Ocean also occupies some of Northern hemisphere. The first law of thermodynamics says the energy is conserved: although, most of the Arctic Ocean is frozen but the ocean itself is warmer than the regional average. While about 70% of the incoming solar radiation is absorbed, the mean surface temperature remains considerably constant. This suggests that the increase in energy due to the absorption of solar radiation needs to be balanced by the losing energy. This energy loss occurs in the form of electromagnetic radiation into space:

$$E = \sigma T^4$$

where, σ is the Stefan-Boltzmann constant: the Stephen law says emission from a blackbody happens equally for all wavelengths. A black body is theoretically a perfect absorber which absorbs all radiation regardless of its wavelength or angle of reception. Any object in thermal equilibrium, emits the same amount of radiation it absorbs. In fact, a blackbody is also a perfect emitter:

$$\left(\frac{Q}{4}\right)(1-\alpha) = \sigma\, T_e^4$$

where, σ is the mean global albedo: the approximate value of α is 0.3 and the value of Q is equal to 1366 W m^{-2} and the radiation temperature is 255°K or -18°C (Robinson, 2010). This balance is shown by the simplest meteorologic model. Despite the simplicity of this model, its results are especially related to the life on earth. Earth's average surface temperature of 15°C or 288°K is considerably warmer than the temperature of the exhaust gases. This difference suggests that our description of the energy balance is incomplete (Grodsky *et al*., 2012).

MATERIALS AND METHODS

Atmosphere (structure and composition): The Solar System and Earth were formed around 4.6 billion year ago and Earth's atmosphere, about 4.57 billion year ago. The Earth's early atmosphere is likely to have been made up of hydrogen and helium (He) because these two were the main constituents of the gas and dust orbiting the sun and forming up planets (Minobe *et al*., 2008). At that time, the Earth and its atmosphere were very hot. Particularly in high temperatures, hydrogen and helium molecules were moving fast. The movement speed of these molecules has been so great that they could eventually overcome earth's gravity and escape into space.

At about 2.7 or 3.5 billion year ago, the second appearance of the atmosphere was probably associated with volcanic activities. Volcanoes created this second atmosphere by releasing water vapor, ammonia, carbon dioxide. Other gases such as nitrogen and carbon monoxide also existed in the second atmosphere. The emergence of oxygen in the second atmosphere was due to the activity of simple bacteria and (shortwave) ultraviolet radiation on water vapor such that these rays with their great energy broke the vapor molecules into oxygen and hydrogen from which hydrogen left the atmosphere and the oxygen remained. After that, the photosynthesis begin to operate. Organisms such as cyanobacteria (cyan algae) received carbon dioxide and water and used them in photosynthesis to produce carbohydrates (sugars) and release oxygen. The detection of cyanobacteria in 3.5 billion years old limestones indicate that they have existed at that time (Quere *et al*., 2003).

Around 400 million years ago, many molecules of carbon dioxide were solved in the oceans and simple bacteria emerged that could produce oxygen. At the same time, the solar radiation broke down the molecules of ammonia, the nitrogen and hydrogen were separated and the lighter hydrogen escaped into space. At this time, Phanerozoic eon (the eon in which that animals became abundant on earth) started and the oxygen-breathing animals began different types of animal life. Earth's atmosphere started to evolve when it possessed oxygen. This molecule then caused the emergence of the ozone layer; a layer that protects life on Earth and prevented harmful ultraviolet rays from hitting the Earth's surface.

Earth's today atmosphere consists of a couple of high dense gases (nitrogen and oxygen molecules) and lots of other gases (such as water vapor, carbon dioxide and ozone). The atmospheric Water is available in both liquid and solid forms and there are also suspended particles in the air, in both solid and liquid phases. Ther thermodynamic state of atmosphere can be determined using its pressure, density and temperature (Liss and Merlivat, 1986):

$$p = \rho R_d T$$

This equation is derived from the equation of state of ideal gases. The water vapor content of the air is an important factor in its thermodynamic state (Cayula and Cornillon, 1992). According to its temperature gradient, density changes, pressure changes, gas mixtures and finally its electric properties, earth's atmosphere is divided into a number of different layers:

- Troposphere
- Stratosphere
- Mesosphere
- Ionosphere
- Exosphere

Troposphere: Troposphere, is the atmosphere's lowest layer which is made up of smaller sub layers. The accumulation of all the atmosphere's water vapor is what distinguishes this layer from the others; this is also the reason most weather-determinant activities related to humidity (such as cloudiness, raining, snowing, fog and lightening) take place just in this atmospheric layer. Since troposphere's heat source is the radiation energy of earth's surface, for higher altitudes in this layer the temperature will drop. The troposphere layer's thickness is subjected to different thermal conditions at different latitudes. The magnitude for this thickness, varies from 17-18 km at the equator, to 10-11 km at the temperate zones and 7-8 km at poles.

Stratosphere: The stratosphere layer is positioned above troposphere and its average thickness is about 23 km. Through the 1st 3 km, the air temperature is constant but at higher parts it increases with altitude. Cloud formation rarely happens in stratosphere and only under certain special circumstances mountain clouds named as pearl clouds, may appear at 21-29 km from the surface as a consequence of air's wavy movement caused by obstacles. One of the other important features of the stratosphere layer, is the formation of its ozone layer especially in distances of 20-30 km from the surface due to different photochemical reactions. The ozone content of this layer usually follows a seasonal pattern with a maximum amount in spring and a minimum in Autumn.

Mesosphere: The mesosphere is positioned above the hot ozone layer and its temperature decreases an amount of 0.3°C for every 100 m increase in height and thus the temperature in its upper border at 80-90 km from the surface reaches down to -80°C. As a result of this low temperature, the inconsiderable amount of water vapor existing in this layer freezes and creates Noctilucent Clouds. These clouds become observable in summer time and at higher latitudes. The mesosphere is the coldest atmospheric layer (Grodsky *et al*., 2012).

Ionosphere: From the upper boundary of the mesosphere to an average height of 1000 km in earth's atmosphere, there is an intensive electrical field that is caused by ions and free electrons. In fact, when the energetic sunrays from outer space, hit the upper parts of atmosphere, the molecular bonds break down and the atoms become ionized. By ionization, the electrons are released and the remaining atoms turn into ions; that's why this layer of the atmosphere is called "Ionosphere". The intensity of this ionization isn't identical for all altitudes; so there are different layers with different electron and ion densities in ionosphere; these layers become very important in radio communications. These layers are called: layers D-F.

Exosphere: The conditions named out for the mesosphere are also present in this layer which means gases in this layer still have their electric conductivity. The particle's speed in this layer are too high and in some cases reach 11.2 km sec^{-1}. The exosphere is the layer of transition of Atmosphere into space and the distance of its upper part from the ground is estimated to be >3000 km.

RESULTS AND DISCUSSION

The composition of atmosphere: The atmosphere of the earth with 90 km height from the surface, is a mixture of different gases and nitrogen, oxygen, argon, carbon dioxide and water vapor make up for 77.99% of its volume. Observations indicate that up to an altitude of 50 km, the mixture ratio of all atmospheric gases except water vapor remains appreciably unchanged. The mass of atmosphere is about 6.5×1014 tons.

Basically, the lighter gases (especially hydrogen and helium) are more abundant in the upper atmosphere. The basic changes refer to the amount of the two major non-permanent gases, water vapor and ozone.

Water vapor: In some regions, the air has almost no vapor while in some other places the moisture accounts for up to 4% of the air's volume. About 90% of all the atmosphere's water vapor is accumulated in its lowest layer with about 6 km height from the surface of the earth. Ozone: the ozone content of atmosphere is mainly concentrated in an atmospheric layer with 15-35 km in thickness. The ultraviolet rays that illuminate the upper atmosphere are responsible for the decoupling of oxygen molecules in layers between 80-100 km height (2012).

The changes with seasons and latitude: The ozone amount on the equator is small and especially in spring, it increases for latitudes upper than 50°N.

The ozone stored during "Polar Night", causes the creation of a rich layer of this gas for the early spring. The vapor content is exactly related to the heat and thus it's greater in summer and in the lower latitudes but the deserts of tropical areas are an exception (Leduc *et al*., 2012).

Changes with time: The quantity of Carbon dioxide and Ozone might become subjected to long term changes. Carbon dioxide is added to the atmosphere, mainly by the activity of organisms on the land and in the sea. Other small sources of this molecule include the decomposition of organic elements in soil and the combustion of fossil fuels. An increase in carbon dioxide content, causes the atmosphere to absorb too much energy from sunlight.

Since, the ozone absorbs both earth's and sun's radiation, if some changes happen for the ultraviolet rays of sun, relevant changes in connection to them, may also occur for the ozone content.

Ocean (structure and composition): Almost 71% of the Earth's surface (an area of about 361 million km^2) is covered with brackish water and this water is generally divided into several oceans and seas. The total volume of all the oceans is about 1. 3 billion km^2 and these oceans have an average depth of 3790 m. There are about 230,000 known species living in the ocean but many deep parts of the oceans still remain unexplored and it is being estimated that more than 2 million aquatic species exist there. The reason for the creation of Earth's oceans is unknown but scientists believe they first formed during Hadean eon and may have been the beginning source of life on Earth. More than half these oceans have depths of over 3,000 m (9,800 ft). The salinity of ocean water is at about 35 parts per thousand and near the seas this salt concentration is varies between 30-38 ppt.

The structure of the ocean: Just like the equation of atmosphere's state, the equation of state for seawater is also a function of pressure, temperature and density. Unlike the atmosphere, there is no general theory (like there is for ideal gases) for fluids and their equation of state for must be derived experimentally and based on a number of existing measurement. Additionally, while the atmosphere's density is a function of only temperature and pressure, the density in the ocean is a function of pressure, temperature and salinity (where this salinity is the concentration of dissolved ions or salts in water) (Minobe *et al*., 2008). The complete form of this equation is very complex because it must include the properties of all the solvent salts. Fortunately, there is not a significant difference between relative densities of the ions of different kinds of ocean salts and so, a special quantity may be determined that can account for the salinity of all different salt ions in seawater. Pressure change with depth is almost linear for the ocean, it is 1.019716 dbar pressure increase for every meter increase in seawater's depth. The pressure in a depth less than 10 m below the ocean's surface is about equal to the atmospheric pressure at the surface (Small *et al*., 2008).

So, the pressure in ocean is usually expressed as a fraction of the total pressure, with the assumption that atmospheric pressure is zero (Curry and Webster, 1999).

The local variations of the temperature and salinity in the ocean, is especially more sensible at depths of more than 300-500 m below the surface than the upper parts of ocean. As a part of the ocean-atmosphere interactions, surface winds result in the development of a shallow mixing layer, where temperature and salinity are almost constant. At the tropics and mid-latitudes, there is a severe reduction of this mixing sublayer's temperature. This area of extreme temperature decline is called "Thermocline" and it extends to a depth of about 1,000 m (Small *et al*., 2008). The temperature of thermocline region gradually reduces to about -3°C and the seasonal changes make very little difference in it. Permanent thermocline exist at polar latitudes because the water temperature is very low and the sea is often covered with ice. Furthermore, the average air temperature at the Arctic Ocean is below 0°C which reflects the fact that the freezing temperature of sea water decreases with an increase in its salinity. In contrast to temperature, the salinity levels strongly increase for high and polar latitudes. The changes in salinity, make an important effect on density changes at all latitudes and especially in high latitudes. The ocean's circulating currents are critical components of the ocean-atmosphere system. The salinity of water near the ocean's surface, varies significantly with longitude, so it's relatively fresh in the polar latitudes and relatively salty in the tropics (especially in the subtropics). These changes in salinity are passing and depend on freshwater flux in the ocean borders, including the precipitation and evaporation, the input from rivers and the melting of glaciers (Liss and Merlivat, 1986).

Sea water is about 1000 times denser than air. So, there is a very stable boundary between atmosphere and ocean. The density of seawater is a function of pressure, temperature and salinity and it's usually expressed as a deviation from the density of pure water at ±4°C (1000 kg m^{-3}) (Wanninkhoff, 1992):

$$\sigma\,(T, P, S) = \rho\,(T, P, S) - 1000 \text{ kg m}^{-3}$$

There are a number of different algorithms available to calculate the density of sea water (for example, look at sea water Python module). The Implementation of many of these algorithms, include the elimination of the pressure profile instead of using surface pressure as an argument and as a result we have (Reichler, 2012):

$$\sigma_t\,(T, S) = \rho\,(T, S) - 1000 \text{ kg m}^{-3}$$

In addition, the densest surface waters are located at high latitudes while the minimum density is in the tropics. Surface water is denser than lower waters and this causes

Table 2: The composition of ocean water with a standard salinity of 34.7 PSU (Wells, 2012)

Constituention	Mass mixing ratio in sea water (g kg^{-1})	Global salt (%)
Chloride (Cl^-)	19.215	54.96
Sodium (Na^+)	10.685	30.58
Sulphate ($SO^{-2}{}_4$)	2.693	7.70
Magnesium (Mg^{2+})	1.287	3.69
Calcium (Ca^{2+})	0.410	1.17
Potassium (K^+)	0.396	1.13
Bicarbonate ($HCO^-{}_3$)	0.142	0.41
Bromide (Br^-)	0.067	0.19
Boric acid (H_3BO_3)	0.026	0.07

the subsidence of surface waters. This profile suggests that surface water's subsidence at high latitude oceans (±65, ±55) is likely to be more intense than in the tropics.

The composition of oceans: On average, dissolved salts make up about 3.5% of seawater's volume. The seawater contains salty and ionic compounds because they are easy to dissolve in water. The breaking up of continent's Bedrocks, volcanoes, hot groundwater springs and atmosphere are all the sources of these salt. Ocean's salt often subsides on ocean floor sediments and also become transferred to Biosphere and Atmosphere (Reichler *et al*., 2012). The PH of ocean is relatively fixed and is suitable for the evolution of marine life. The ocean's total salinity has been constant for at least 100 million years now but some scientists believe that the salinity ratio of the ocean water has changed significantly over time. These changes refer to the use of different salts by biological organisms (for example, calcium ions are being used by crustaceans). Table 2 shows the major constituents of sea water. The ocean contains several minor compounds such as strontium, silicon, iron, lithium, phosphorus, iodine, oxygen and dissolved gases and nitrogen. These minor components are an important part of chemical and biological processes of the ocean. The first measurements of ocean showed that the salinity of every place is different in constituent ions but regardless of the location of sampling, salinity is largely constant (Reichler *et al*., 2012).

These observations have led to the defining of a single salinity parameter, in which the ratio of major constituents are assumed to be constant. Developing electrical conductivity methods to measure salinity in 1980, brought forth a significant improvement in measurement accuracy. At this time, salinity became redefined as the ratio of electrical conductivity of seawater to the electrical conductivity of a standard potassium chloride solution at a standard temperature and pressure and expressed using Practical Salinity Units (PSU). Recent advances in measurement technology have resulted in adoption of a more precise definition of salinity (absolute salinity) in 2009 (Reichler *et al*., 2012).

Interface processes and their role in the interconnected system: A surface's energy flux is the amount of energy that passes through each unit area of land surfaces during 1 sec and its unit is W m^{-2}. The energy balance is as follows (Jacobson, 2005):

$$G = Rnet\text{-}Fturb$$

Where:

G = The flux through the ocean
Rnet = Net radiation balance
Fturb = Heat flux turbulance

The radiation flux: Solar radiation spectrum covers a range of electromagnetic waves from ultraviolet to microwave including visible light and infrared. The wavelength distribution of the radiation from sun is a function of its temperature. According to Planck's law, the body that completely absorbs radiation is a blackbody. The maximum energy radiated by the sun correspond to the visible wavelength range at about 0.5 micrometers, whereas with the Earth's surface temperature at 290°K, its maximum energy transfer happens in the thermal infrared range (10-12 μm) (Gill,1982). The atmosphere is not completely transparent for solar radiation: gas molecules, droplets, cloud crystals and dust particles (aerosols) have a substantial absorption capacity for the infrared range. They also reflect the incoming radiation and diffuse it (mainly in the visible range as well as in particular infrared and microwave ranges). So, the atmosphere absorbs part of solar radiation and reflects infrared radiation both in up and down directions; clouds also absorb, reflect and defuse the radiation coming up as well as solar radiation. The total solar radiation, equals the sum of the radiation reaching the surface, the radiation weakened and defused by the atmosphere and the infrared radiation emitted from it. However, the balance of the energy received by the Earth's surface is completely dominated by short waves (the spectral range from visible to near infrared, about 3 μ >0.4 μm) and infrared range (from 3-12 μ).

The penetration of radiation in Ocean, is somewhat affected by phytoplankton activities and it makes the photosynthesis possible by being absorbed in phytoplankton cells and thus producing certain pigments containing chlorophyll. These cells, re-emit the radiation at lower energy wavelengths (through fluorescence) using specific organic substances called dye (often in solution form). The amount of energy stored chemically is low, but the absorbed energy is quickly converted to heat which leads to the warming of the ocean layers (Kiehl and Trenberth, 1997). These cells, like other marine particles of biological origin or minerals and different solutes, cause partial diffraction of light. These interactions lead to the creation of different thermal layers. Furthermore, at the presence of these particles the light penetrates less and the photosynthetic production at the deeper layers becomes avoided. The presence of organic and inorganic particles and the dissolved dye in the surface layers, also affects the reflection of shortwaves and so these two effects determine the radiation balance of earth's surface.

Horizontal diffusion: The atmosphere and oceans are fluids and the winds and currents are capable of horizontally moving these fluids. At the tropics (between about ±35S and ±35N), absorption of solar radiation is greater than long wave radiation output but it's less at other latitudes. These differences indicate the redistribution of energy by horizontal diffusion within the atmosphere and ocean system. This amount of energy transfer is calculated through integrating the difference between radiation output and the absorbed long wave solar radiation, from the Arctic to the Antarctic. Calculations indicate that both the atmosphere and oceans are responsible for about half the energy transferred to the poles (Gill, 1982). depicts the expected temperature of this diffusion, for a variety of scenarios. The first scenario (red line) relates to a situation without any penetration, (e.g., output radiation, solar radiation is absorbed everywhere). This scenario holds for a planet with zero thermal conduction between oceans and atmosphere. The second scenario (blue line) represents a situation with infinite diffusion (for example, the diffusion temperature is the same everywhere). The final scenario (black line) corresponds to earth conditions. The horizontal energy transfer which takes place between the atmosphere and the ocean is clearly an important element for earth's energy balance. In fact, this energy transfer is primarily done by wind and currents rather than emission. However, it's emission that is often used as a computationally efficient transfer element in those energy balance models which necessarily need to be highly simplified (Small *et al*., 2008).

An overview of the global energy balance: By combining satellite observations of radiant flux at the top of the atmosphere with complex radiative transfer models, the calculation of the energy budget of the earth-atmosphere system becomes possible. In the upper atmosphere, the average global flux of incoming solar radiation is balanced by the outgoing longwave radiation from the surface and atmosphere with an albedo of about 30% of the solar radiation. Budget level and atmospheric energy is primarily made up of the absorption of solar radiation and mainly long wave diffusion. The exchange of radiative energy between the surface and the atmosphere is not balanced and the surface emits more than the received energy and the atmosphere receives more energy than it. The amount of this imbalance depends on the difference of sensible and latent heat coefficients for the atmosphere and the surface. The sensible heat transfer is characterized by the direct exchange of heat between the lowest atmospheric layer and the surface which is disturbed by sun and then mixed with the air. Latent heat depends on evaporation of water from the surface and then passing this energy into the atmosphere during condensation and precipitation (Gill, 1982).

The ocean-atmosphere system: The interaction between the atmosphere and the oceans plays a major role in determining the status of Earth's climate for all periods of time. The atmosphere becomes quickly influenced by regional and global changes and passes on these effects to large distances. In this way, the atmosphere acts as a bridge. The ocean, has a tremendous capacity to store and release heat (1000 times more than the atmosphere) which makes considerable influence on the climate of the region. For example, the annual range of temperatures of an area enclosed in land such as Beijing, is more than 50°C while for the west coasts of North America on the same latitude, it's only about 10°C. The ocean prolongs the period of temperature changes by integrating the effects of weather disturbances (Reichler *et al*., 2012).

Heat exchange, takes place globally, regionally and locally between the atmosphere and oceans. Deep tropical convection quickly passes the surface conditions to the upper troposphere, where strong winds spread these changes throughout the world. The Wind stress on ocean surface, causes a commixture in the upper layer of the ocean. The ocean's reaction to climate changes in short periods of time are controlled through the depth of this upper layer of the ocean (which depends on the intensity of the wind pressure). Oceans are the largest water resources on the earth (about 97% of its total water) and evaporation from the ocean surface provides about 90% of atmosphere's water vapor that eventually turns into rain and snow. The distribution of evaporation and precipitation on the ocean surface, causes regional variations of salinity (amount of salt) in surface waters. These salinity changes, automatically leads to changes in the density of ocean water which in turn, improves the vertical movements and mixings in the ocean. The presence of ice in sea, fundamentally alters the heat and water fluxes between the atmosphere and the ocean (Kiehl and Trenberth, 1997). About >2% of the world's water is present as ice. However, most of the water (over 99%) is locked away in the form of ice sheets on land and permafrost. Although, only a small part of the global ice volume is related to sea ice, its horizontal expansion is significant. Sea ice (ice and ocean ice) plays a key role in determining the regime and diversity of the climate. Ocean-atmosphere system is distinct. It brings about considerable implications for global and regional climate and environmental health. A more prominent mode of interaction between the atmosphere and oceans, is El Niño Southern Oscillation (ENSO) which is also the largest source of climate variability. ENSO is the oscillation of the tropical Pacific sea's surface temperature and pressure, created by a positive atmosphere-ocean feedback mechanism (Sarachik and Cane, 2010). East Equatorial Pacific Ocean,is usually much cooler than the tropical western Pacific. Deep tropical convection occurs preferably for higher sea surface temperatures and it's generally focused on the tropical western Pacific. During the warm phase of ENSO (El Nino), East equatorial pacific ocean is warmer than normal and a location change of deep Tropical Convection occurs to the East. During the cold phase of ENSO (La Nina), East Equatorial Pacific become cooler than normal and the location change of tropical deep convection occurs more towards the West. The effects of these changes will be felt around the world (Kiehl and Trenberth, 1997).

The outlook of energy balance: The average temperature at Earth's surface is remarkably stable over time. This stability indicates that the flux of input energy to the Earth system is approximately equal to the output energy flux of the system. This fact, brings us to the simplest class of climate models, the model of energy balance. Although, the energy balance model represents a simplification in the climate system, it's a useful starting point for understanding the interaction of the ocean and atmosphere (Reichler *et al*., 2012).

CONCLUSION

The ocean-atmosphere interactions, play the central role in climate changes, through their energy transfer between the two. These changes are small-scale processes but are involved by temporal and regional changes on all scales. These, includes complex physical- chemical and biological processes carried out in the wavy and turbulent environment of ocean. The exact determining of the quantities and surface fluxes of the ocean is impossible due to its big size. However, by using a network of floating robots being controlled from a long distance in 1990, the simple techniques of ocean parameters measurement and the estimation of their average values were basically revolutionized.

The local examination of the mentioned changes has been made possible, but investigating the large-scale changes in the ocean-atmosphere system is still difficult. Satellite observations has made the inaccessible variables, accessible but the resolutions of the regional data is not accurate enough yet. The future challenges include the collaborative use of this information for regional and global modeling in suitable scales and for different climatic zones and also systematic understanding of the ocean-atmosphere system's climate simulation. For some regions, the determination of ocean-atmosphere interactions is more important. Among these regions, the coastal areas have a particular importance, especially in areas around the Eastern pacific basin where strong exchanges happen between the ocean and the atmosphere. The Arctic Ocean and its higher regions covered with ice and the Arctic and Antarctic are key areas where interactions of the ocean-atmosphere system are still poorly understood. Probably new methods will be needed for this subject.

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