IMPACT OF OCEAN-ATMOSPHERE INTERACTION ON WEATHER AND CLIMATE

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Abstract

In this project we have reviewed the impact of ocean-atmosphere interaction on weather and climate. We considered ocean-atmosphere interaction and exchange of energy by sensible heat, latent heat, wind stress, precipitation heat flux, and surface energy budget of the surface ocean mixed layer. In our work the mass exchanges between ocean-atmosphere boundary layers by the methods fresh water, precipitation, runoff, melting of sea ice, evaporation, inert and sparingly soluble gases in sea water. The radiation transfer at ocean and atmosphere boundary layer interrelations of micro wave and infrared radiation fluxes. The solar radiation by net shortwave irradiance, reflection at the sea surface, and absorption of solar radiation in the ocean. Terrestrial radiation occurs in two ways longwave emission from the sea surface and radiative transfer in the lower atmosphere. To analyzed and identified the impact of ocean-atmosphere interaction on weather and climate, those effects described by the following methods, air-sea feedback, their global distribution, the common practices to diagnose, ocean surface salinity budget, the planetary boundary layer, and the dynamical implications of air-sea interactions.

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Chapter 1

INTRODUCTION

The ocean and the atmosphere share a common boundary at the air-sea interface. This direct physical contact enables the two fluids to exchange energy and matter. The ocean is given energy by solar radiation passing through the atmosphere. There are also other mechanisms for heat exchange. The direct physical contact of the sea and air means that there is exchange of energy through collision of molecules in the surface layer of each fluid. This energy is known as sensible heat. There will also be an exchange of the molecules themselves, generally resulting in net evaporation and therefore transfer of latent heat, from the water surface. The momentum transfer that occurs, when the wind blows over the ocean results in motion within the water [1].

Water continually moves between the oceans, the atmosphere, the cryosphere, and the land. The total amount of water on Earth remains effectively constant on time scales of thousands of years, but it changes state between its liquid, solid and gaseous forms as it moves through the hydrologic system. The Sun provides the energy necessary to evaporate water from the surface. The horizontal and vertical movement of water vapor is critical to the water balance of land areas, the precipitation that falls on the land areas of Earth is water that was evaporating from ocean areas and then transporting to the land in the atmosphere. The excess of precipitation over evaporation in land areas supports the return of water from the land to the ocean in rivers [2].

The Earth receives virtually all of its energy from the Sun in the form of electromagnetic radiation. This radiation is absorbing, reflecting, and scattering by the Earth's surface, the ocean, and the atmosphere. The absorbing radiation is transforming into heat and other forms of energy and eventually it is returning to space as low temperature terrestrial radiation. The radiation is the fundamental importance to atmosphere ocean interaction [3].

Heating and cooling at the ocean surface determine the sea surface temperature, which is a major determinant of the static stability of both the lower atmosphere and the upper ocean. The surface heat fluxes at the air-sea interface are central to the interaction and coupling between the atmosphere and ocean. The ocean surface salinity budget plays an important role in determining the stability of the upper ocean. The saline surface water in the high latitude North Atlantic ocean is a key factor that allows surface water to sink deep into the ocean. On the other hand, fresh surface water acts to stabilize the mixed layer in the Arctic ocean and the tropical Western Pacific [4].

1.1 Objective of the Study

1.1.1 General objective of the study

The general objective of this project is to review the impact of ocean-atmosphere interaction on weather and climate.

1.1.2 Specific objective of the study

To describe the processes of heat exchanges at the oceanic and atmospheric boundary layer.

To review the mass exchanges that occur at the ocean-atmosphere interface.

To study the radiation processes at both oceanic and atmospheric boundary layers.

To recognize the implications of ocean-atmosphere interactions on weather and climate.

This project is organized into six chapters. Chapter one discuss about the introduction part of impact of ocean-atmosphere interaction on weather and climate and the objective of the study. In chapter two modeling heat transfer at air-sea boundary layer that includes ocean-atmosphere interaction, the ocean-atmosphere interface, exchange of energy, and the surface energy budget of the surface ocean mixed layer. In chapter three model of mass interchanges between the air-sea boundary layers that includes fresh water, precipitation, runoff, melting of sea ice, evaporation, inert and sparingly soluble gases in sea water and in chapter four model of radiation transfer at air-sea boundary layers, radiation models of oceanic and atmospheric boundary layer, interrelations of micro wave and infrared radiation fluxes, types of solar and terrestrial radiation, similarly in chapter five we discuss about impact of air-sea interaction on weather and climate that includes air-sea feedback, their global distribution, the common practices to diagnose this feedback, ocean surface salinity budget, the planetary boundary layer, and the dynamical implications of air-sea interactions and finally in chapter six summary and conclusions of the study are included.

Chapter 2

MODELING HEAT EXCHANGES AT AIR-SEA BOUNDARY LAYERS

2.1 Ocean-Atmosphere Interaction

The Earth's climate is determined by many complex physical, chemical, and biological interactions among the ocean, atmosphere, land, and ice/snow, subject to solar and tectonic forcing. Atmosphere ocean interactions are a consequence of exchanges between the Earth's atmosphere and the oceans. These exchanges can be explained in terms of fluxes of energy, momentum, and mass of substances (e. g, moisture, gases, etc). Across the ocean surface, and their redistribution within the ocean and with in the atmosphere, respectively. The chemical interaction between the atmosphere and ocean begins with the transfer of chemical species, gases, from one fluid to the other. This process depends on a number of physical and chemical parameters. The resulting control of the supply of oxygen and carbon dioxide to the ocean and atmosphere respectively. The direct physical contact of the sea and air means that there is exchange of energy through collision of molecules in the surface layer of each fluid. This energy is known as sensible heat. The momentum transfer that occurs, when the wind blows over the ocean results in motion within the water [1].

2.2 The Ocean-Atmosphere Interface

Air-sea fluxes are governed by processes acting on the interface from both above and below, as well interfacial processes which influence vertical exchange. The atmospheric process which governs most air-sea exchanges is turbulent transport through out the depth of the atmospheric boundary layer. The atmosphere boundary layer (ABL) may be defined as the thin layer extending from the earth's surface upward in which the airflow strongly experiences the effect of the earths surface friction. The ocean boundary layer (OBL) is the uppermost part of the ocean where interactions with the atmosphere and the underlying ocean influence water properties.

The two boundary layers adjacent to the air-sea interface are governed by a set of common physical and dynamical characteristics. Often capped by an overlying inversion, the atmospheric boundary layer has a depth which can range from tens of meters during strongly stable flow to several kilometers during convective conditions. The ocean mixed layer (OML) is order of $5-50$ meters deep, depending on stratification, wind stress, and the strength of the underlying thermocline. When normalized by density, the atmospheric boundary layer height and ocean mixed layer depth are of the same order of magnitude.

There are however a few unique differences between the atmospheric boundary layer and ocean mixed layer which one must note. Surface waves exert an influence on profiles of turbulence and fluxes in the adjacent boundary layers, extending to distances between 2 to 5 times the wave height in to both the atmospheric boundary layer and ocean mixed layer. In the atmospheric boundary layer, this depth of influence is insignificant with respect to the full depth of the atmospheric boundary layer, but in the ocean the depth of influence can at times encompass the entire depth of the ocean mixed layer.

The marine boundary layer (MBL) is the lowest part of the atmosphere directly influenced by the ocean's surface. Its depth typically ranges from 500 to $1,000$ meters, though this can vary based on factors such as sea surface temperature, atmospheric conditions, and geographic location. The depth of the marine boundary layer can change with the time of day and season. During the day, solar heating can deepen the boundary layer, while at night, cooling can cause it to become shallower. Seasonal changes in sea surface temperatures and atmospheric conditions also affect the marine boundary layer depth.

The marine boundary layer plays an essential part in weather and climate systems. It affects cloud formation, precipitation patterns, and the exchange of heat and moisture between the ocean and atmosphere [6].

2.3 Exchange of Energy by Heat Fluxes at the Air-Sea Interface

2.3.1 Exchange of heat (sensible heat)

Heat is transferred through conduction and convection. The direct physical contact of the atmosphere and ocean enables energy to be exchanged between them by conduction. Such energy exchange is known as sensible heat. This occurs due to collisions between the molecules of the two fluids at their interface, with energy being transferred to the cooler, and therefore, slower, molecules. It should be noted that this process is a statistical one. Sensible heat the energy transferred between the ocean and the atmosphere that causes a temperature change in the air or water. Heat is transferred from the warmer ocean surface to the cooler air above it through molecular interactions.

Convection involves the movement of heat through the transfer of fluid (air or water) as it circulates. When the ocean surface is warmer than the overlying air, the air in contact with the surface warms up, becomes less dense, and rises. This upward movement of warm air allows cooler air to descend and take its place, creating a convective heat transfer.

Sensible heat transfer therefore depends on the temperature difference between the near surface air and the sea surface. Turbulence, and high wind speeds, encourage conduction by mixing air from higher in the atmosphere with that at the surface, allowing the ocean to interact with more air than that in the shallow surface layer. The sensible heat transferred to the atmosphere is again difficult to measure, so the empirical formula is often used.

$$Q_s = \rho_a C_H U(T_S - T_A)[1], \tag{2.3.1}$$

where Q_s is the sensible heat, ρ_a is the density of air, C_H is the specific heat of air, U is wind speed, and T_S and T_A are sea surface temperature and air respectively. If $T_S > T_A$, $Q_s > 0$ and the sensible heat flux is out of the ocean which therefore cools. The global average temperature of the surface ocean is indeed 1 or 2 degrees warmer than the atmosphere and so, on the average, sensible heat is transferred from the ocean to the atmosphere [1, 2].

2.3.2 Exchange of moisture (latent heat)

Latent heat the energy absorbed or released during phase changes of water (evaporation, condensation) without changing the temperature of the water. Evaporation is the liquid water at the ocean surface changes into water vapour (gas). When the ocean surface is warmer than the overlying air, water molecules at the surface gain enough energy to escape into the atmosphere. This process absorbs latent heat from the ocean, which cools the ocean surface. Condensation is the water vapour in the air changes back into liquid water. Moist air rises and cools, it reaches its dew point and condenses into water droplets or ice crystals.

Radiation dominates the exchange of heat between the atmosphere and ocean. The most important of these is latent heat transfer. When water is evaporated from the ocean surface energy is supplied to the molecules to free them from the strong inter molecular bonds with in liquid water. When the water molecules condense to form water droplets, in a cloud , the energy latent with in the vapour is released as heat and contributes to the driving energy of the cloud producing process.

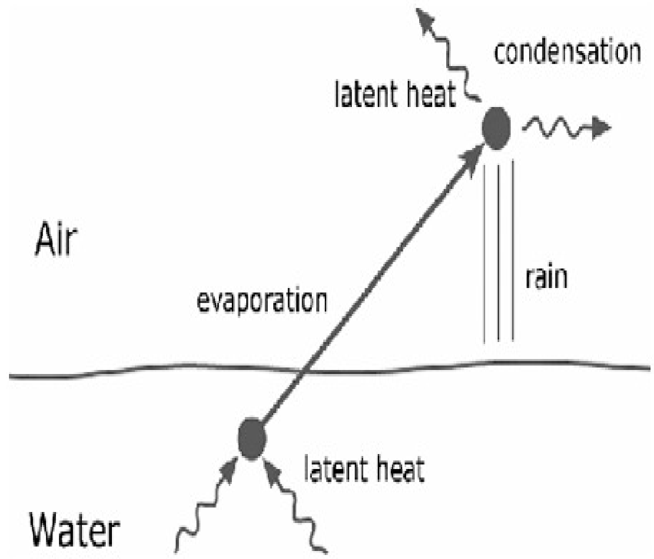

Figure 2.1: Latent heat is taken from the ocean to evaporate water that is subsequently released into the atmosphere when the vapor condenses to form rain [9].

The latent heat added to the atmosphere by the ocean is,

$$Q_E = L_v E[1], \tag{2.3.2}$$

where Q_E is the latent heat, E is the evaporation, and L_v is the latent heat of vaporization. It depends slightly on temperature, $L_v = 2.5 \times 10^6 Jkg^{-1}$ [9, 10].

2.3.3 Exchange of momentum (wind stress)

Wind stress is the primary mechanism by which momentum is transferred from the atmosphere wind to the ocean surface. This interaction movement of surface water can lead to the generation of ocean currents (are continuous, directed movements of sea water generated by various factors, wind, temperature and salinity difference and earth's rotation) and surface waves (are waves that travel along the interface between the ocean and the atmosphere) and is the main factor in the ocean's circulation. These currents are the upper layer of ocean circulation driven directly by the wind.

Wind stress is the force exerted by the wind (or the work done by the wind) on the ocean surface. Wind stress contributes to ocean mixing, which affects the distribution of heat, nutrients, moisture, and salinity in the ocean and impacting weather systems and climate. It results from the frictional force of the wind blowing over the ocean surface.

The magnitude of wind stress depends on wind speed, wind direction, and the drag coefficient of the ocean surface. Wind stress obtained using the following formula,

$$\tau = \rho_a C_d U^2 [11], \tag{2.3.3}$$

where τ is the wind stress, ρ_a is air density, U is wind speed, and C_d is the drag coefficient. Which depends upon the height of the wind measurement and the atmospheric stability as well as wave characteristics [10, 11].

2.3.4 Precipitation heat flux

Precipitation heat flux the heat energy transferred to or from the ocean surface (or other surfaces) due to precipitation processes. It is an important component in the exchange of heat between the atmosphere and the ocean. Precipitation heat flux is the flux of heat associated with the addition or removal of heat through precipitation events, such as rain or snow, impacting the ocean surface or other land surfaces. When rain falls on the ocean surface, it can either warm or cool the surface, depending on the temperature difference between the rain and the ocean surface. If rain is warmer than the ocean surface, it will transfer heat to the ocean. If rain is cooler, it will absorb heat from the ocean.

From the surface energy budget equation of ocean,

$$F_{Qo}^{net} - F_{Qo}^{adv} - F_{Qo}^{ent} = F_{Qo}^{rad} + F_{Qo}^{SH} + F_{Qo}^{LH} + F_{Qo}^{PR}, \tag{2.3.4}$$

where F_Q is the flux density of energy, F_{Qo}^{rad} is the net surface radiation flux, F_{Qo}^{SH} is refers to the surface turbulent flux of sensible heat, F_{Qo}^{LH} is the surface turbulent flux of latent heat, F_{Qo}^{PR} is the heat transfer by precipitation, F_{Qo}^{adv} is the the transport of energy into or out of the ocean mixed layer via fluid motions, F_{Qo}^{ent} is the transport of energy into or out of the ocean mixed layer via entrainment and/or molecular diffusion at the base of the ocean mixed layer, and F_{Qo}^{net} is the ocean heat storage term. Sign convention a term adding heat to the mixed layer is positive and a term removing heat from the mixed layer is negative.

The term F_{Qo}^{PR} is the heat flux at the surface due to rain or snow. Heat transfer by precipitation occurs if the precipitation is at a different temperature than the surface. If a falling raindrop is in thermal equilibrium with its surroundings, then the temperature of the raindrop at a given height will be the same as the wet-bulb temperature of the atmosphere at that height. Assuming that such an equilibrium exists, we can write,

$$F_{Qo}^{PR} = \rho_l c_{pl} p_r (T_{wa} - T_o), \tag{2.3.5}$$

where Qo is the heat flux from the rain, ρ_l is the density of the liquid water in the raindrop, c_{pl} is the specific heat capacity of liquid water, p_r is the rainfall rate in m/s, T_{wa} is the atmospheric wet-bulb temperature, and T_o is the temperature of the raindrop.

The value of F_{Qo}^{PR} are greatest for large rainfall rates and for large differences between the atmospheric wet-bulb temperature and sea surface temperature. Except for rare circumstances $T_{wa} < T_o$, and the heat flux from rain cools the ocean. During heavy rainfall events, values of F_{Qo}^{PR} may be the largest term in the surface energy budget however, when is averaged over longer time scales, the contribution of this term to the surface energy budget is quite small and is commonly neglected.

In the presence of snowfall the term F_{Qo}^{PR} is more complex, since the ocean must provide latent heat to melt the snow. Hence we have the following expression for the heat flux associated with snowfall into the ocean:

$$F_{Qo}^{PR} = \rho_s c_{ps} P_s (T_{Ia} - T_o) - \rho_s L_{il} P_s [4], \tag{2.3.6}$$

where ρ_s is the density of snow, c_{ps} is the specific heat capacity of snow, P_s is the snow precipitation rate, T_{Ia} is the ice bulb-temperature of the atmosphere just above the surface, T_o is the ocean temperature, and L_{il} is the latent heat of fusion of snow. The precipitation heat flux is a major component of the surface energy budget, influencing ocean temperatures and the exchange of heat between the ocean and atmosphere [4].

2.4 The Surface Energy Budget Of the Surface Ocean Mixed Layer

The energy budget within the mixed layer of the ocean, which is the top layer of the ocean where temperature and salinity are relatively well mixed due to turbulence and wave action. The mixed layer typically extends from the ocean's surface to depths ranging from several meters to about 150 meters, depending on various factors such as wind strength, solar heating, and local meteorological conditions.

The surface energy budget of the surface ocean mixed layer, we use various formulas to quantify the energy exchanges. The surface energy budget of ocean mixed layer calculating the balance between incoming and out coming going energy at the ocean surface, including solar radiation, long wave radiation, sensible heat flux, and latent heat flux. The general formula for the surface energy budget can be expressed as:

$$Q_{net} = Q_{in} - Q_{out}, \tag{2.4.1}$$

where Q_{net} is the net energy at the ocean's surface, Q_{in} is the incoming energy at the surface, and Q_{out} is the outgoing energy from the surface.

Incoming solar radiation (short wave radiation) is the energy from the sun that reaches the ocean surface. Some of it is absorbed directly by the ocean, while the rest is reflected back into the atmosphere. The proportion reflected is known as the albedo.

$$Q_{sol} = (1 - \alpha) I_{sol}, \tag{2.4.2}$$

where Q_{sol} is the absorbed solar radiation by the ocean surface, I_{sol} is the incident solar radiation, and α is the albedo of the ocean (reflectivity of the ocean).

Outgoing long wave radiation the ocean surface emits infrared radiation (heat) back into the atmosphere. This is influenced by the temperature of the ocean surface.

$$Q_{lw} = \epsilon \sigma T_s^4, \tag{2.4.3}$$

where Q_{lw} is the outgoing longwave radiation from the ocean surface, ϵ is the emissivity of the ocean surface (usually close to 1 for the ocean), σ is the Stefan-Boltzmann constant (5.67×10^8 W/m^2 k^4), and T_s is the temperature of the ocean surface.

Latent heat flux is the energy absorbed by the ocean surface during the evaporation of water. The energy required for the phase change from liquid to vapor is drawn from the ocean surface, leading to cooling.

$$Q_{lh} = \rho L_v E, \tag{2.4.4}$$

where Q_{lh} is the latent heat flux, ρ is the density of the water, L_v is the latent heat of vaporization, and E is the evaporation rate.

Sensible heat flux is involves the transfer of heat between the ocean surface and the overlying atmosphere via convection and conduction. It is related to the difference in temperature between the ocean surface and the air above.

$$Q_{sh} = \rho c_p h(T_s - T_a), \tag{2.4.5}$$

where Q_{sh} is the sensible heat flux, ρ is the density of air, c_p is the specific heat capacity of air at constant pressure, h is the heat transfer coefficient, and T_s and T_a are he temperature of the ocean surface and the air above the ocean respectively.

Net heat flux the net heat flux is the sum of all the positive and negative energy fluxes. It determines whether the ocean surface is gaining or losing heat.

$$Q_{net} = Q_{sol} - Q_{lw} - Q_{sh} - Q_{lh}[11], \tag{2.4.6}$$

Positive (Q_{net}), indicates that more energy is absorbed by the ocean surface than is lost, leading to warming of the ocean. Negative (Q_{net}), indicates that more energy is lost from the ocean surface than is absorbed leading to cooling of the ocean [11].

Chapter 3

MODEL OF MASS EXCHANGES BETWEEN THE AIR-SEA BOUNDARY LAYERS

3.1 Exchange of Mass at the Air-Sea Interface

3.1.1 Fresh water

In the annual mean, the fresh water leaving the ocean through evaporation must equal that being returned through precipitation and river runoff if the ocean fresh water balance is in equilibrium. The estimate of the flux of fresh water across the ocean surface. Evaporation dominates in the subtropics, while under the tropical atmospheric convergence zones and at middle and high latitudes precipitation provides the major contribution. Further inhomogeneity is provided by the river out flow and the flow of saline water. Just as for heat, the oceans must transport fresh water to compensate for these imbalances.

The exchange of mass at the ocean-atmosphere interface involves various processes that contribute to the dynamic balance between the ocean and the atmosphere. Fresh water exchange the movement of freshwater into and out of the ocean. Fresh water at the ocean-atmosphere interface primarily refers to the input of fresh water into the ocean from various sources. This input affects ocean salinity, circulation, and climate. This includes inputs from rivers, precipitation, and melting ice, as well as out puts like evaporation.

The sources and sinks of fresh water at the ocean surface are a crucial component of the global water balance. The ocean loses fresh water through evaporation than it gains through precipitation. The remaining is contributed by river runoff, with the change in storage of fresh water in the oceans being a much smaller residual. The surface fresh water flux determines upper ocean mixing and the thermohaline component of ocean circulation. The stabilizing effect of a local net fresh water surface input limits the efficiency of ocean atmosphere communication. It impedes heat exchange with subsurface layers, nutrient transport to the upper ocean, and trace gas exchange with the atmosphere. In the atmosphere these fluxes determine atmospheric moisture content and latent heat that drive tropospheric convection, horizontal advection, and precipitation patterns.

The air-sea flux fresh water is simply the difference evaporation lost from the ocean surface and precipitation gained by the ocean from the atmosphere, often we written,

$$F = E - P[13], \tag{3.1.1}$$

where F is the net fresh water flux, E is the evaporation rate, and P is the precipitation rate [13, 28].

3.1.2 Precipitation

Precipitation is the moisture flux from the atmosphere to the Earths surface coupling the atmospheric and terrestrial branches of the hydrologic cycle and serving as the downward directed mass flux for climate of the second kind. Rain and snow provide the primary moisture inputs to the land phase of the hydrologic cycle. Dew and fog drip provide small moisture inputs for specific locations that are very important locally, but the moisture quantity delivered by these processes is small compared to rain and snow.

Precipitation is the water vapor in the atmosphere condenses into liquid droplets or ice crystals and falls to the Earth's surface. Precipitation directly adds fresh water to the ocean, which can impact the salinity of sea water. This input can be significant in regions with high rainfall and can influence ocean stratification and circulation patterns. When rainfalls on the ocean, it dilutes the sea water, reducing its salinity. This decrease in salinity can affect the density of the ocean water and influence ocean currents and mixing. Changes in salinity due to precipitation can affect the formation of deep water currents. For example, fresh water input can disrupt the formation of dense, cold water that drives deep ocean circulation.

Precipitation is produced in a series of stages beginning with super saturation of ascending and cooling air. Condensation, droplet formation, and successful descent of the droplets or particles to the Earths surface complete the atmospheric cycling of moisture. The amount of water vapor present in the air is a fundamental factor to precipitation formation [14, 27].

3.1.3 Runoff

Runoff the process by which water, primarily from precipitation and snow melt, flows over the land surface and eventually makes its way into bodies of water such as rivers, lakes, and oceans. It is a critical component of the hydrological cycle and plays a significant role in various environmental and ecological processes. The mass exchange at the ocean atmosphere interface, including runoff, is a dynamic process that affects not only the distribution of water but also the chemical and biological processes in both systems.

The climatic role in the runoff process is characterized by the path followed by water in arriving at the stream channel after it has been delivered to the surface by precipitation. This perspective provides a basis for distinguishing the climatic influence as distinct from the geomorphologic influence imposed by terrain characteristics related to the watersheds size, shape, and relief.

3.1.4 Melting of sea ice

When sea ice melts, it releases fresh water into the ocean. Since sea ice is formed from sea water, its melting adds fresh water without significantly affecting the ocean's salinity. This can lead to changes in ocean stratification and circulation patterns. The addition of fresh water from melting sea ice can lower the salinity of surface waters, which can impact the density of sea water. This change in density affects ocean circulation and the formation of deep water masses, which play a crucial role in global ocean circulation patterns.

Sea ice has a high albedo, meaning it reflects a large portion of incoming solar radiation back into space. When sea ice melts, it exposes the darker ocean surface, which absorbs more heat. This can lead to further warming and additional ice melt in a feedback loop, accelerating climate change [13].

3.1.5 Evaporation

The process of loss of water in the form of vapour from a wetted land surface, or from a water surface, such as that of ponds, rivers, lakes or open oceans, whenever the vapour pressure of the air above the surface falls short of its saturation value at the temperature of the underlying surface is called evaporation. Evaporation requires an energy source at a surface that is supplied with moisture, the vapour pressure in the air must be below the saturated value and air motion removes the moisture transferred into the surface layer of air. The saturation vapour pressure increases with temperature.

Evaporation is water converted from liquid to vapor and moves from the ocean surface into the atmosphere. This removes fresh water from the ocean, and the vapor can later contribute to precipitation. The primary driver of evaporation is solar radiation. When the sun's energy heats the ocean surface, water molecules gain enough energy to transition from a liquid state to a vapor state. The energy required to change water from liquid to vapor is known as latent heat. This process absorbs heat from the environment, which influences local and global climate conditions. Higher surface temperatures increase the rate of evaporation because more water molecules have enough energy to escape the liquid phase. Warmer ocean waters and air temperatures thus enhance evaporation rates.

Evaporation move both heat and fresh water between the ocean and the atmosphere. The rate of evaporation is directly proportional to the oceans latent heat loss and is controlled by the ocean atmosphere temperature difference. Therefore the ocean influences evaporation through its surface temperature. The addition of fresh water through river discharge, spring runoff from snow melt on land and through the melting of sea ice, this results in a low salinity surface ocean in the equatorial and high latitude regions and high salinity.

The evaporative flux is the amount of water lost per unit area of the surface per unit time and may be easily derived using the equation of state for water vapour and expressed by the relation.

$$E = -\rho K_D(\varepsilon/P)\partial e/\partial z, \tag{3.1.2}$$

where E is the evaporative flux, K_D is the coefficient of eddy diffusivity for water vapour, ρ is the density of air, ε is the ratio of the molecular weights of water vapour to dry air ($\varepsilon = 0.625$), p is the atmospheric pressure, e is the vapour pressure of water, and z is the small height above the surface. The evaporative heat flux is then given by,

$$H_e = -L_\rho K_D(\varepsilon/P)\partial e/\partial z = LE, \tag{3.1.3}$$

$$H_e = LE[7], \tag{3.1.4}$$

where H_e is the evaporative heat flux, E is the evaporative flux, and L is the latent heat of vaporization of water at the temperature of the under lying surface [7, 16, 28].

3.1.6 Inert and sparingly soluble gases in sea water

The atmosphere is almost entirely a mixture of gases. These gases can enter the ocean across the air-sea interface. If no molecules of a particular gas were in solution in the upper ocean then this transfer would act as a drain on the atmospheric store of the gas. Once sufficient molecules from the atmosphere accumulate in the water for as many to leave the sea as enter in a given time, equilibrium is reached and the water is said to be saturated with respect to the gas. Atmospheric gases are generally close to a state of saturation in surface waters of the ocean.

Some soluble gases are carbon dioxide, methane, and nitrous oxide, all of these are greenhouse gases and oxygen is non greenhouse gas. Such gases also show an enhanced temperature dependence of their solubility. The solubility of a gas determines the relative ease with which it can be absorbed into the ocean if there were no other gases present in the atmosphere, and if an equilibrium state was achieved. The troposphere however, is a mixture of many gases and the basic driving force for exchange of gas across the air-sea interface is the difference in gas concentration, or partial pressure, between the two media.

Gas flux at the air-sea surface is the transfer of gases between the atmosphere and the ocean, specifically how gases such as carbon dioxide (CO_2), oxygen (O_2), nitrogen (N_2), and other trace gases move across the interface between the air and the water. This flux can occur in both directions, gases can be absorbed by the ocean from the atmosphere is gas absorption or released from the ocean into the atmosphere is gas emissions. The gas flux of a gas across the interface into the ocean is often written as,

$$F = K_T(P_a - P_s), \tag{3.1.5}$$

where F is the gas flux, P_a and P_s are the partial pressures of the gas in the surface atmosphere and ocean respectively, and K_T is the transfer velocity. As its name suggests K_T has the units of a velocity, it represents the variability of the rate of exchange due to the sea state and atmospheric stability.

Moisture flux at the air-sea surface is the transfer of water vapor between the ocean and the atmosphere at the interface between the two. It is the process by which moisture is exchanged across the surface of the sea due to differences in temperature, humidity, and pressure between the ocean and the overlying air. This exchange plays fundamental role in the water cycle and impacts weather patterns, climate and ocean dynamics. The general equation for moisture flux at the air-sea surface is,

$$F = K_w A(Q_a - Q_s), \tag{3.1.6}$$

where K_w is transfer coefficient or the gas exchange velocity for water vapor, Q_a and Q_s are specific humidity in the atmosphere and at the sea surface respectively, and A is the area of the ocean surface through which the exchange is occurring.

The aerosol flux at the air-sea interface is the movement of aerosol particles between the atmosphere and the ocean surface. These particles are tiny solid or liquid substances suspended in the air, and their transfer can occur in different ways. They can originate from natural sources like sea spray, volcanic eruptions, and dust storm or anthropogenic (human made) sources like industrial emissions and vehicular exhaust. The process is important because it influences both the atmosphere and ocean in several significant ways, impacting climate, weather, marine ecosystems and atmospheric chemistry.

$$F = K_a A(C_a - C_w), \tag{3.1.7}$$

where K_a is transfer coefficient or the gas exchange velocity for water vapor, and C_a

and C_w are the concentration of aerosols in the air and water respectively.

The heat flux at the air-sea surface is the transfer of heat between the ocean surface and the atmosphere. This exchange plays a major factor in regulating the Earth's climate, weather patterns, and oceanic circulation. Heat fluxes are driven by various factors, such as solar radiation, wind, humidity, and temperature differences between the ocean and the atmosphere. Heat flux can occur through radiation, conduction and convection processes, and it is driven by temperature differences between the ocean surface and the atmosphere.

$$Q = hA(T_s - T_a)[15], \tag{3.1.8}$$

where h is heat transfer coefficient, and T_s and T_a are temperature of the ocean surface and air respectively.

The ocean and the atmosphere exchanges of the long lived greenhouse gases such as, carbon dioxide (CO_2), nitrous oxide (N_2O), and methane (CH_4) and non-green house gas oxygen O_2 are important for the global climate change.

The exchange of methane (CH_4), at the ocean atmosphere interface involves the transfer of methane gas between the ocean and the atmosphere. Wind driven waves and turbulence increase the contact between the ocean surface and the air, enhancing the rate of methane transfer. The breaking of waves creates bubbles and turbulence, which helps mix gases between the ocean and the atmosphere.

The exchange of nitrous oxide (N_2O), at the ocean atmosphere interface involves the transfer of this greenhouse gas between the ocean and the atmosphere. Turbulence and wave action increase the surface area where the ocean and atmosphere interact, enhancing the transfer of gases. Wind and wave driven turbulence help mix nitrous oxide across the ocean atmosphere boundary.

The exchange of carbon dioxide (CO_2), at the ocean atmosphere interface is component of the global carbon cycle. Wind and wave driven turbulence increase the contact area between the ocean surface and the air, facilitating the transfer of CO_2. The solubility of CO_2 in sea water decreases with increasing temperature.

The exchange of oxygen (O_2), between the ocean and the atmosphere is a vital process that impacts marine life, climate regulation, and the global oxygen cycle. Oxygen moves from areas of higher concentration to lower concentration through the water column and across the air sea interface. At the ocean surface, the concentration of oxygen in the water is influenced by atmospheric oxygen levels [15, 28].

Chapter 4

MODEL OF RADIATION TRANSFER AT AIR-SEA BOUNDARY LAYERS

4.1 Radiation Models of Oceanic Boundary Layer

The most important characteristics of the sea water are their temperature and salinity. Permittivity of water ϵ, is a function of the wavelength λ. The dependence of permittivity ϵ on the wavelength, temperature, and salinity is increase in ionic conductivity at a given wavelength λ due to an increase in salt content in the solution. Of all the salt solutions present in natural water, the most extensively studied in the radio wavelength range is N_aC_l the solution proper to sea water (in concentration) upto $3-5$ moles L^{-1}.

At wind speeds U of about $5-7\ ms^{-1}$, foam appears on the ocean surface. The water surface area covered by foam increases with a rise in wind speed. Foam may also be formed as a result of turbulence of the surface and subsurface waves, and also in regions of plankton by a reduced pressure caused by the motion of internal. Theoretical estimates show the real and imaginary parts of the foam dielectric constant to vary inversely with the air concentration and constitute several units at concentrations in excess of 0.9. The spectral dependence of foam dielectric constant is due to the frequency dependence of the water dielectric constants.

The dependence of the system ocean-atmosphere (SOA) brightness temperature T^b of a calm (not exited by wind force) water surface on temperature and salinity. Which include the characteristics of this medium as the complex dielectric constant ϵ, observation angle θ and polarization.

$$T^b = \chi T_w, \chi = 1 - R, \tag{4.1.1}$$

where T^b is the brightness temperature, χ is the coefficient of polarization, T_w is the water surface temperature, and R is the Fresnel reflection coefficient, with vertical observation $\theta = 0$,

$$\chi_v = \chi_h = \chi_o, \tag{4.1.2}$$

where χ_v is the coefficient vertical polarization, χ_h is the coefficient horizontal polarization, and χ_o is the coefficient common polarization.

$|\epsilon| = \epsilon_1\sqrt{1 + tg^2\delta}$, $0 \leq \theta \leq 0.5\Pi$, $tg\delta = \frac{\epsilon_2}{\epsilon_1}$ is the dielectric loss angle. where $|\epsilon|$ is the magnitude of the dielectric constant, ϵ_1 is the real part of the dielectric constant, $tg\delta$ the tangent of the dielectric loss angle δ, and ϵ_2 is the imaginary part of the dielectric constant.

The angular dependence of the coefficients of horizontally and vertically polarized radiation can be approximated with an accuracy of $2 - 10$ % by the following polynomials.

$\chi_h(\theta) = \chi_0(1 - A_h\theta^2)$, $0 \leq \theta \leq 0.5\Pi$,

$$\chi_v(\theta) = \chi_0(1 + A_v\chi_v\theta^2 + B_v\theta^4), 0 \leq \theta \leq 0.4. \tag{4.1.3}$$

When the receiving antenna is oriented at an angle α relative to the observational plane,

$$\chi_a(\theta) = \chi_v \cos^2 a + \chi_h(\theta) \sin^2 \theta. \tag{4.1.4}$$

The thickness of the efficiently emitting layer (the skin layer) is determined as the inverse of the absorption factor:

$l_e = \frac{1}{\gamma}$, where,

$$\gamma = \frac{2\Pi\sqrt{2}}{\lambda}\sqrt{|\epsilon| - \epsilon_1}. \tag{4.1.5}$$

Analysis of the fresh water dielectric constant data discloses the relation:

$$l_{e1}/l_{e2} = f(\lambda_1/\lambda_2). \tag{4.1.6}$$

Which is valid for k from $0.94 - 1$ in the $5 - 30$ cm wavelength range.

$$l_e(\lambda) \approx 0.01\lambda^{-2}[16], \tag{4.1.7}$$

where λ is the wavelength of the electromagnetic radiation, l_e is the skin layer thickness, γ is the absorption factor, and f is the frequency of the electromagnetic radiation. Where l_e , λ are in centimeters.

The frequency dependent relations between the system ocean-atmosphere (SOA) radiation characteristics at microwaves and water surface temperature and salinity. The emissivity of a smooth water surface covered with a uniform foam or an oil layer can be approximated by the equations derived for the stratified models which account for the dielectric characteristics of individual layers and multiple reflections of electromagnetic radiation fluxes from the boundaries of individual layers.

4.2 Radiation Models of Atmospheric Boundary Layer

A standard cloudless atmosphere is characterized by such parameters as temperature, air density, and water vapor pressure. The height temperature profile T_a is described by the broken line functions T_{a1}, T_{a2}, and T_{a3}.

$$T_{a1} = T_0 - ah, 0 \leq h \leq 11Km, \tag{4.2.1}$$

$$T_{a2} = T_a(11Km), 11Km \leq h \leq 20Km, \tag{4.2.2}$$

$$T_{a3} = T_a(11Km) + h - 20Km, 20Km \leq h \leq 32Km, \tag{4.2.3}$$

where T_0 is the temperature at sea level, typically taken as 288.15 K, T_{a1} T_{a2}, and T_{a3} are function of temperature at layer 1, layer 2, and layer 3 respectively, and a is the lapse rate, which is given as 6.5 K/km. Here, T_0 is the temperature at sea level ($h = 0$), and a is the temperature vertical gradient. Due to one of the known (US) standards, T_0= 288.15 K, and $a = 6.5$ K km^{-1}.

For an approximate calculation, we can use the simplified formula:

$$T(h) = T_0 \exp^{-h/H}, H_T = 4.4km, \tag{4.2.4}$$

where $T(h)$ is the temperature at height h, T_0 is the temperature at sea level (288.15 K), h is the height above sea level in kilometers, and H_T is the scale height, given as 4.4 km.

For many preliminary calculations, we can also use the exponential altitude model of atmospheric pressure P:

$$P(h) = P_0 \exp^{-h/H}, P_0 = 1013mb, H_p = 7.7Km. \tag{4.2.5}$$

The exponential model is used here to describe the volumetric water vapor density in the atmosphere:

$$\rho_e(h) = \rho_0 \exp^{-h/H}, \tag{4.2.6}$$

where $P(h)$ is the atmospheric pressure at height h, P_0 is the standard atmospheric pressure at sea level, H_p is the scale height for pressure, given as 7.7 km, ρ_0 is the water vapor density at sea level (at $h = 0$), and $\rho_e(h)$ is the water vapor density at height h. Here, ρ_0 is the atmospheric water vapor density at sea level and depends on climatic and seasonal time scales. On the average, it varies from 10^{-2} $g\ m^{-3}$ for a dry climate upto 30 gm^{-3} for a warm climate. The models from equation 4.2.1 to 4.2.3 are using in the climatic and seasonal time scales.

The main absorption and radiation components of the atmosphere are its water vapor and oxygen, which are entirely determined by the atmosphere temperature T, air pressure P, water vapor density ρ_e, and their vertical distribution. Water vapor and oxygen have absorptive lines at wavelength 1.35 cm ($f = 22.235$ GHz) and in the region of 5 mm, respectively.

The brightness temperature of the system ocean-atmosphere (SOA) depends on the radiation properties of the water surface and the absorption of the atmosphere at microwaves, the system ocean-atmosphere (SOA) brightness temperature T^b is given by the following relations.

$$T^b = T_1^b + T_2^b + T_3^b, \tag{4.2.7}$$

where,

$$T_1^b = T_w^b \exp(-\tau) \tag{4.2.8}$$

is an intensity of radiation flux from the water (oceanic) surface T_w^b attenuated in the atmosphere (the quantity T_w^b is proportional to emissivity of the water surface and its thermodynamic temperature T_w).

$$T_2^b = \int_0^H T_0(h) \exp[\tau(h) - (H)] dh \tag{4.2.9}$$

is an intensity of the up-going atmosphere radiation flux;

$$T_3^b = R \int_0^H T_a(h) \exp[\tau(h) - (H)] dh \tag{4.2.10}$$

is an intensity of the atmosphere down-going radiation flux reflected by the water surface.

$$\tau(h) = \int_0^h \gamma(h^{'}) dh^{'} [16], \tag{4.2.11}$$

where T^b is the brightness temperature of the water surface (the ocean surface), T_1^b T_2^b, and T_3^b are individual components that contribute to the overall brightness temperature, T_w^b is the intensity of radiation flux from the water surface, T_w is the thermodynamic temperature of the water surface, τ is the integral attenuation of the radiation by the atmosphere, which depends on the linear absorption factor and the thickness of the absorbing layer counted from the ocean surface ($h = 0$), R is the coefficient of reflection of the atmosphere down-going radiation flux from the water surface, γ is the absorption factor, and h is the thickness of the absorbing layer.

The system ocean-atmosphere (SOA) brightness temperature values at millimeters and centimeters, where the resonance effects become apparent. Besides molecular attenuation in water vapor and oxygen, the microwave propagation at centimeters and millimeters is also greatly affected by various hydro meteors, first of all, rain and cloud [16].

4.3 Interrelations of Micro Wave and Infrared Radiation Fluxes

In the framework of a plane layered model of thermal radiation (absorption) in the ocean-atmosphere system shown in the figure 4.1, it is possible to analyze the processes of the electromagnetic energy transfer in various layers, as well as to estimate an intensity of these radiation components, which can be registered with satellite (A), airplane (B), and shipboard (C) equipment.

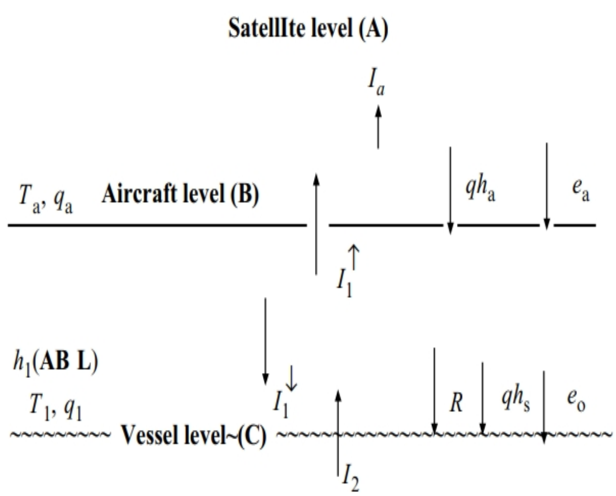

Figure 4.1: Parameterization scheme of the main characteristics of thermal and electromagnetic energy transfer in the ocean-atmosphere system [17].

A: In satellite observations, the ocean-atmosphere natural microwave radiation I_{sat} is composed of the intensity of the free atmosphere radiation I_a and an intensity of the up-going radiation flux $I_1^{\uparrow}$ at the top of the atmosphere boundary layer attenuated in the free atmosphere (multiplier G_a):

$$I_{sat} = I_a + I_1^{\uparrow} G_a, \tag{4.3.1}$$

where $I_1^{\uparrow} = I_1 + (I_1^{\downarrow} R_{21} + I_2) G_1$ is an intensity of the up-going radiation flux at the top of the atmosphere boundary layer, $I_1^a = I_1 + I_a$, G_1 is an intensity of the down-going radiation flux at the bottom of the atmosphere boundary layer, $I_1(G1)$ is an intensity of the integral (total) attenuation in the atmosphere boundary layer, I_2 is an intensity of the atmosphere boundary layer natural radiation, G_1 is the integral attenuation of radiation in the atmosphere boundary layer, and R_{21} is the coefficient of reflection of the down-going radiation flux $I_1^{\downarrow}$ from the water surface. It is assumed that the electrophysical parameters of the atmosphere boundary layer and the free atmosphere, in spite of the difference (in a general case) between their temperature and humidity characteristics, are consistent with each other, that is, the reflectivity at their interface is absent or negligibly small compared with R_{21}.

B: In observations from an aircraft (at the top of the atmosphere boundary layer, the ocean atmosphere radiation intensity I_{air} is determined solely by the component $I_1^{\uparrow}$.

$$I_{air} = I_1^{\uparrow}. \tag{4.3.2}$$

C: In shipboard observations (at the lower boundary of the atmosphere boundary layer), an intensity of radiation I_{ship} is computed as follows:

$$I_{ship} = I_2 + I_1^{\uparrow} R_{21}. \tag{4.3.3}$$

The characteristics of natural radiation of the boundary (I_1) and free (I_a) atmosphere in expressions (4.3.1) – (4.3.3) are related to the corresponding values of temperature T_1, T_a and of the integral absorption G_1, G_a of this media:

$$I_1 = T_1(1 - G_1), I_a = T_a(1 - G_a). \tag{4.3.4}$$

An intensity of the thermal radiation of the water surface I_2 is proportional to

$I_2 = \epsilon T_2$ in the micro wave length range,

$I_2 = \delta B(T_2)$ [17], in the infrared band, where T_2 is the ocean boundary layer temperature, $B(T_2)$ is the Plancks function with T_2 as an argument, ϵ is the emissivity of the water surface at microwaves, and δ is the infrared emissivity of the ocean surface.

In the micro wave range of wavelengths approximation is valid, respective values of the brightness temperature T^b are used as a measure of the radiation intensity of different components of the ocean atmosphere system. To characterize the infrared radiation intensity I here and below, we will use the concept of the effective (radiation) temperature T^r, defining it from the equation $B(T^r)$, that is, as a thermodynamic temperature of the absolute(ideal) black body with a radiation intensity is equal to I [17].

4.4 Solar Radiation

The Sun is an average star in our galaxy. It is a gaseous sphere with a diameter of 1.42×10^6 km. It's distance to the Earth is 150×10^6 km, where as the next closest star is 3×10^5 times as far away.

The interior of the Sun, where the nuclear reactions occur that ultimately lead to life on Earth, is incredibly hot, at a temperature of several million degrees Celsius. However, the electromagnetic radiation that provides the energy for the climate system is derived from the outer layers of the Sun. The greatest amount of radiation comes from the photosphere, a layer some 300 km thick in the solar atmosphere. This varies in temperature from 10000 K at the bottom to 5000 K at the top. Outside the photosphere are much less dense regions the chromosphere and corona. While these outer regions are at much higher temperatures, up to millions of degrees in the corona, their low density means that they radiate relatively little energy. Most of this is at very short, X-ray and gamma-ray wavelengths which affect the upper atmospheres of the planets but do not penetrate into the lower atmosphere.

The Sun appears to us as (almost) a black body. That is, the spectrum and total energy of electromagnetic radiation emitted from the Sun (as from all surfaces, and indeed molecules) is a function of its temperature.

The total energy flux emitted by a black body follows:

$$E = \sigma T^4, \tag{4.4.1}$$

where E is the total energy flux, σ is the Stefan-Boltzmann constant, and T is the temperature in degrees Kelvin. The energy density, E_λ or radiant energy per unit wavelength λ, per unit volume per second, is given by

$$E_\lambda = \frac{8\pi c}{\lambda^5}\left[\frac{1}{e^{\frac{hc}{\lambda kT}} - 1}\right] [1], \tag{4.4.2}$$

where c is the speed of light, k is Boltzmanns constant, h is Plancks constant, and T is the temperature in degrees Kelvin.

The vast majority of the energy that reaches the Earth comes from the ultraviolet through visible to infrared part of the spectrum. The variation in the amount of energy emitted by the Sun is probably small on non-geological time scales. At the Earths distance from the Sun this solar constant is about 1.38 kWm^{-2}. On very long time scales, comparable with the life of the planet, astrophysicists believe that the Suns irradiance varies dramatically as the supply of fuel within the Sun changes. The variation in the Earths orbit can affect the amount of energy reaching the Earths surface by a few percent, on time scales of thousands of years [1].

4.4.1 The net shortwave irradiance at the sea surface

The shortwave radiation that reaches the sea surface has passed through the atmosphere. The atmosphere is only partly transparent for shortwave radiation. Ultraviolet radiation with wavelengths shorter than 0.29 μ is absorbed by ozone. Although the total amount of ozone in the atmosphere is less than 1 part in $10,000$, its presence shields life on land and in the upper ocean from lethal high frequency radiation. The atmosphere is also partly transparent for infrared shortwave radiation (0.7-4 μ,m). Oxygen, water vapour, and carbon dioxide are the main absorbing constituents, as can be deduced from figure 4.2. The amount of shortwave infrared radiation, which passes through the atmosphere and is absorbed at the sea surface, can lead to significant warming of the surface layer itself. The atmosphere is quite transparent to

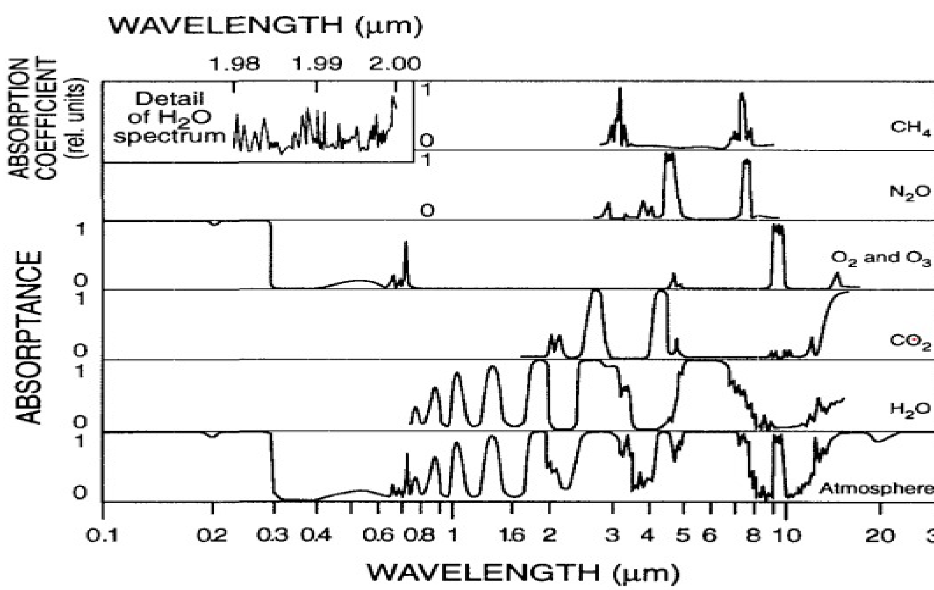

Figure 4.2: Absorption spectra for H_2O, CO_2, O_2, O_3, N_2O, CH_4, and the absorption spectrum of the atmosphere [3].

visible radiation $(0.4-0.68)$ μm, which contains nearly 60 % of the energy emitted by the Sun. On a clear, cloudless day, the visible radiance of direct sunlight is reduced only by scattering. On the other hand, scattering increases the light that comes down to the surface from other directions of the Sky. The diameter of air and water molecules is so small compared to the wavelength of light that higher order practically negligible, and Rayleigh scattering, inversely proportional to the fourth power of the wavelength, is produced. The Sun radiates most intensely at a wavelength of 0.47 μm in the blue green part of the spectrum. This light is scattered about four times more effectively than red light with a 0.66 μm wavelength. The well known explanation for the blue colour of the Sky and of the sea is based on this phenomenon.

The composition of sunlight varies with the Sun's elevation. When the Sun is about 20^o above the horizon, the rays have to pass through three times as much air as when the Sun is overhead. The longer the path through the atmosphere, the more blue light is scattered out of the direct solar beam. As a result, the peak of the spectrum of direct sunlight is shifted at low solar angles from blue green in planetary space to yellow or red at the earth's surface. Part of the blue light that has been scattered out of the direct beam is lost into space; the remainder reaches the surface as sky radiation, which is nearly though not entirely diffuse. Its contribution to the irradiance of the sea surface in clear weather increases with the Sun's zenith distance from 16 %, when the height of the Sun is 60^o, to about 37 % , when it is 10^o above the horizon. The radiance is more nearly omnidirectional when clouds reduce the amount of directly transmitted light.

4.4.2 Reflection at the sea surface

Some of the radiation that reaches the sea surface is reflected back into the atmosphere. The reflectance of the surface, defined as the ratio of the reflected to the incident radiance, is not identical with the albedo. The albedo of the sea surface is the ratio of all shortwave radiation leaving the surface to the incident irradiance. It includes light from specular reflection at the surface alone, with upwelling light that passes the interface from below, after scattering within the water.

When the Sun is low, more light is reflected and the fraction of radiant energy that penetrates the water is reduced. The same applies to that part of the diffused Sky radiation which comes from the regions close to the horizon. Clouds increase the reflectance when the Sun is high because their presence then increases the fraction of radiation that comes from near the horizon. When the Sun is low, clouds allow more light to penetrate the water because they then increase the light scattered down vertically.

When the sea is rough, the local reflectance becomes a function of time. Radiation that comes from a high elevation is in this case more likely to meet a sloping wave surface, causing a higher fraction of the irradiance to be reflected. When radiation from near the horizon reaches a rough sea, it does not have a grazing incidence as on a calm surface; rather, it meets wave flanks that slope toward the source of the light. The local angle of incidence is therefore smaller, causing less of this horizontal radiation to be reflected and more to penetrate into the sea.

In the below figure 4.3, the solid line represents the relation obtained from glitter pattern observations. The individually plotted points are surface slope estimates based on infrared measurements.

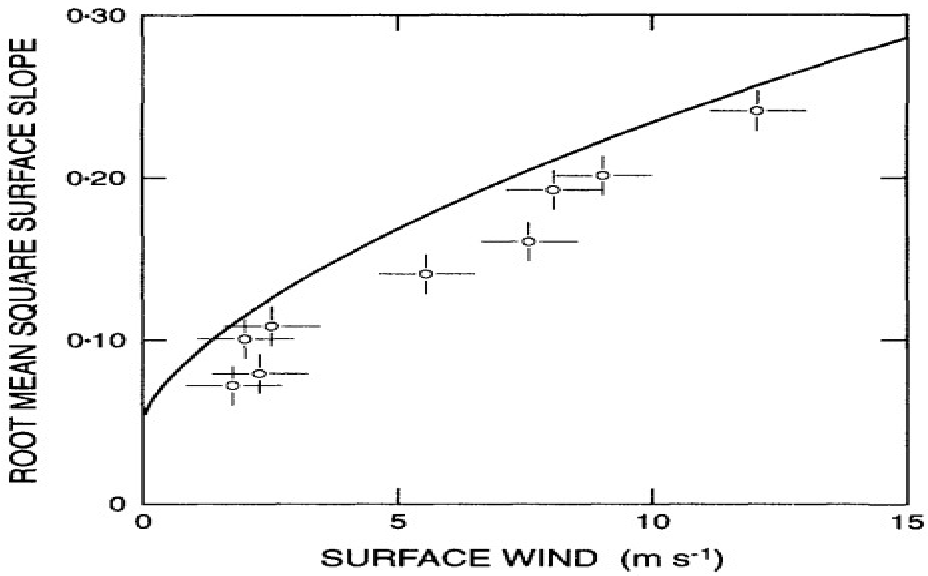

Figure 4.3: Root mean square sea surface slope as a function of winds peed [3].

The reflectance of the sea surface is influenced by other factors as well. Bubbles scatter light, and this gives whitecaps a diffused reflectivity which generally is much larger than that of any singly connected water surface. The albedo in sea ice can vary from less than 0.4 for melting, dirty ice to as much as 0.9 for ice which is covered with fresh snow.

4.4.3 Absorption of solar radiation in the ocean

Most of the solar radiation that does pass the interface is visible and is ultimately absorbed by the ocean. In clear water, blue radiation may penetrate to considerable depths, whereas red radiation is mainly absorbed in the first few meters. The transmittance at various wave lengths of a layer water for several ocean areas, coastal waters, and for Crater Lake.

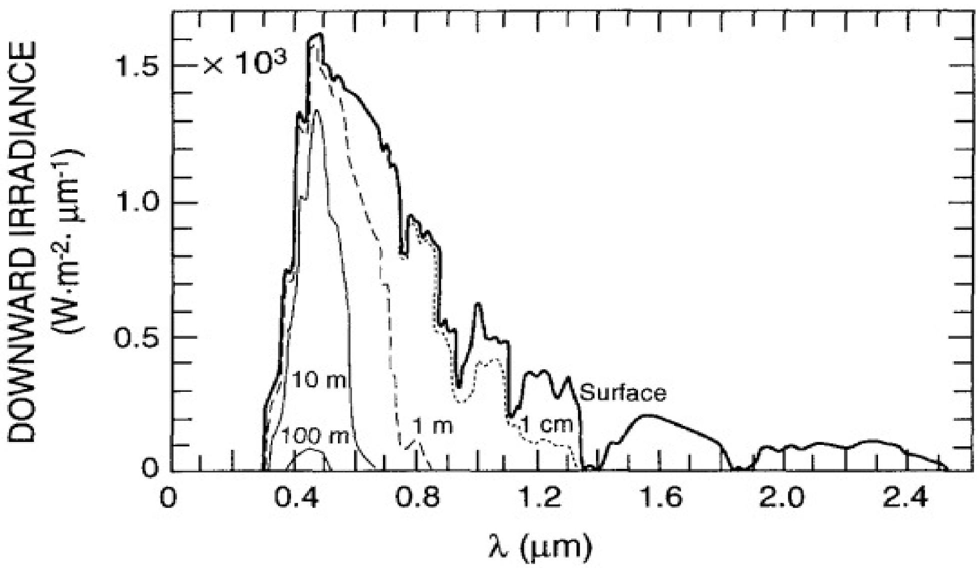

Figure 4.4: The complete solar spectrum of downward irradiance in the sea at various depths [3].

All radiation that is absorbed in the ocean is converted into internal energy. This solar heating is distributed over various depths as we have seen from the transmittance variability. The spectra of downward irradiance in clear ocean water are presented in figure 4.4. We see that in the infrared practically all radiation is absorbed in the top 1 m and that a substantial fraction of this is absorbed in the top 1 cm. The strong absorption in this surface layer leads to relatively rapid warming near the surface. The larger amount of radiation in the visible range is distributed over a deeper layer and contributes slowly to the heating of that layer.

The absorption of shortwave radiation may be approximated by a series of exponentials in the form

$$F_{is}(z) = F_{is}(0) \sum_n a_n \exp_n^b z[3], \tag{4.4.3}$$

where F_{is} is the shortwave irradiance that penetrates the sea surface, a_n and b_n are coefficients, with the condition that $a_n = 1$.

The rate of sensible heating of the oceans by the sun is slow because of the great heat capacity of the oceans. Even in the tropics, where the irradiance may be as high as 1000 Wm^{-2}, the heating is only 0.036^o C per hour if all of this radiation is absorbed in the upper 10 m. The cumulative effects are, of course, very large. The amount of solar energy that is absorbed directly by the oceans is between 1.5 and 3 times as large as that absorbed directly by the entire atmosphere. It is more than three times as large as that absorbed by all the global land surfaces. A relatively small fraction of this absorbed energy is transported over large horizontal distances by ocean currents. The larger part is stored locally or in the immediate vicinity, and is later transmitted to the atmosphere mainly by evaporation and by longwave radiation.

4.5 Terrestrial Radiation

4.5.1 Longwave emission from the sea surface

Longwave radiation is absorbed and emitted in the top $1mm$ of the sea. For this radiation the sea surface approximates a black body. The exitance, F_{el} from the sea surface to the atmosphere may be approximated by

$$F_{el} = \epsilon_f \sigma T_s^4 - (1 - \epsilon_f) F_{il} \tag{4.5.1}$$

where ϵ_f is the flux emittance of the sea surface for all wavelengths of the black body spectrum, ϵ is actual emittance is about 0.98 for wavelengths from 3 to 50 um with a slight dependence on temperature and salinity, and T_s is the temperature of the sea surface. Which may be slightly different form the bulk temperature near the sea surface, i.e., a few cm below the surface. The flux emittance is related to the actual emittance by

$$\epsilon_f = 2\int_0^{\Pi/2} \epsilon(\theta)\cos\theta\sin\theta d\theta, \tag{4.5.2}$$

assuming that T_s appears to be the same for all directions. This is usually the case but it may not be quite valid when there is a strong temperature gradient from the surface down. The emittance, $\epsilon(\theta)$, is a function of the zenith angle because in this case,

$$\epsilon(\theta) = 1 - r(\theta)[3]. \tag{4.5.3}$$

The reflectance, r, behaves in similar ways for longwave and shortwave radiation. Thus, under a clear sky, the longwave radiance of a calm sea is largest when seen directly from above ($\theta = \pi$).

When clouds are present, the longwave radiation from the sky is almost diffuse and the difference between vertical and horizontal radiances diminishes greatly. The longwave absorption and emission of clouds depends on their liquid water content. For a liquid water path of 40 gm^{-2} which corresponds to a cloud thickness of 100-300 m at the top of the boundary layer and somewhat more for higher clouds, the radiance is almost equal to blackbody radiance at cloud temperature. The result is that, with a cloud cover, the radiance from the sea surface, including the reflected radiance from the clouds, approaches the blackbody radiance.

When the sea surface is rough and the sky is overcast, the radiance from the surface is equal to the blackbody radiance at the sea surface temperature for all practical purposes. Nevertheless, there still may be transfer of radiation from the sea surface to the cloud base. This may have some impact on the cloud layer, especially when the cloud layer is cooler than the sea surface. The cloud base may be heated by several degrees Celsius per day at the base, which enhances the unstable stratification and turbulence within the cloud layer.

4.5.2 Radiative transfer in the lower atmosphere

The upward and downward longwave irradiances depend on the vertical distributions of temperature, water vapour, CO_2, and other trace gases in the atmosphere. The dependence can be treated numerically. We start by asking what the contribution to the irradiance is of a volume element at height z with thickness dz to a reference level z_r as illustrated in figure 4.5.

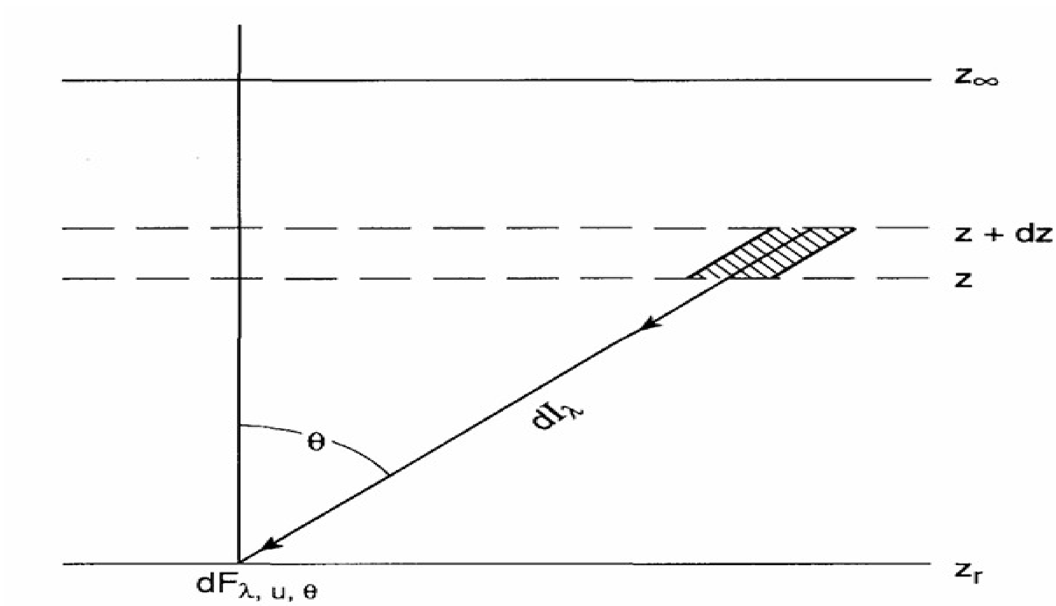

Figure 4.5: Contribution to irradiance $(dF_{\lambda,\upsilon,\theta,})$ at $z = z_r$ made by a differential element at level z and zenith angle θ [3].

The monochromatic radiance, dI_λ ,emitted by the element in the directions θ is given in the form of $k_\lambda I_\lambda^* \sec\theta d\upsilon$, where u is the optical thickness measured in the vertical. The radiance is attenuated portion of the radiance increment dl_λ, that arrives at the reference level, is

$$dI_\lambda \exp(-K_\lambda u \sec\lambda) = I^*_\lambda K_\lambda \sec\theta du \exp(-K_\lambda u \sec\theta). \tag{4.5.4}$$

We now assume that the atmosphere is uniformly stratified in the horizontal, so that the radiance coming from all elements with the same angle θ is the same and has the same attenuation. Thus, all volume elements seen at zenith angle θ contribute an equal increment to the irradiance at the reference level. Keeping in mind that $d\omega = 2\Pi \sin\theta d\theta$ and using ($F_i = \int_0^{2\Pi} I \cos d\omega$), the increment to the irradiance $dF_{\lambda,u,\theta}$ from the ring may be expressed by,

$$dF_{\lambda,\upsilon,\theta} = 2\Pi I^*_\lambda K_\lambda \sin\lambda \exp(-K_\lambda \upsilon \sec\theta) d\theta d\upsilon, \tag{4.5.5}$$

This expression must now be integrated over all zenith angles from 0 to $\frac{\Pi}{2}$ over all layers $d\upsilon$, and over all wavelengths that contribute to longwave radiation. the spectra flux transmittance $\tau_{f,\lambda}$, is defined as the proportion of the incidence irradiance that is transmitted through the layer.

$$\tau_{f\lambda} = \frac{F_\lambda}{F_{\lambda,0}} = \frac{F_\lambda}{\Pi I_\lambda 0}, \tag{4.5.6}$$

and the transmitted irradiance F, may be calculated by integrating the transmitted radiance, over the solid angle, $F_i = \int_0^{2\Pi} I \cos\theta d\theta$ and $F_i = \Pi I$. This yield,

$$\tau_{f\lambda} = 2\int_0^{\Pi/2} \exp(-K_\lambda \upsilon \sec\theta) \sin\theta \cos d\theta \tag{4.5.7}$$

Upon differentiation we find that,

$$\frac{\partial \tau_{f\lambda}}{\partial \upsilon} = -2K_\lambda \int_0^{\frac{\Pi}{2}} \exp(-K_\theta u \sec\theta) \sin\theta d\theta \tag{4.5.8}$$

After integrating (4.5.5) over the solid angle, (4.5.8) may be introduced into it with the result

$$dF_{\lambda,u} = -F^*_\lambda \frac{\partial \tau_{f\lambda}}{\partial u} du \tag{4.5.9}$$

The flux $\epsilon_{f\lambda}$, is related to by $\pi_{f\lambda} = 1 - \tau_{f\lambda}$ and therefore,

$dF_{\lambda,u} = F^*_\lambda \frac{\partial \epsilon_{f\lambda}}{\partial u} du = F^*_\lambda d\epsilon_{f\lambda}$

Upon integration over wavelength and optical depth, the radiation from a cloudless atmosphere reaching the reference level may be expressed formally by

$$-F_i = -\int_0^{\epsilon_f s} F^* d\epsilon_f = \int_0^{\epsilon_f s} \sigma T^4 d\epsilon_f, \tag{4.5.10}$$

where ϵ_{fs} represents the flux emittance from the top of the atmosphere to the reference level.

Determination of ϵ_f as a function of height is especially useful, if ϵ_f is independent of temperature over the range encountered in the vertical column of air that contributes to the irradiance. This is the case when the area under the blackbody envelope, which represents energy absorbed in an atmospheric layer, does not change as the black body envelope changes with temperature. This condition is approximately fulfilled for water vapour and carbon dioxide.

The upward irradiance consists of a contribution from the atmosphere below the reference level plus a contribution from the ocean surface. This may be written as,

$$F_e = \int_0^{\epsilon_{fs}} \sigma T^4 d\varepsilon_f + (1 - \epsilon_{fs})\sigma T_s^4 \tag{4.5.11}$$

where the index s refers to the surface. The reflected radiation from the sea surface has been neglected in this equation because it is a minor contribution.

Equations 4.5.11 and 4.5.12 can be evaluated when ϵ_f is known as a function of optical depth. In figure 4.6, experimental observations of flux emissivity are given for

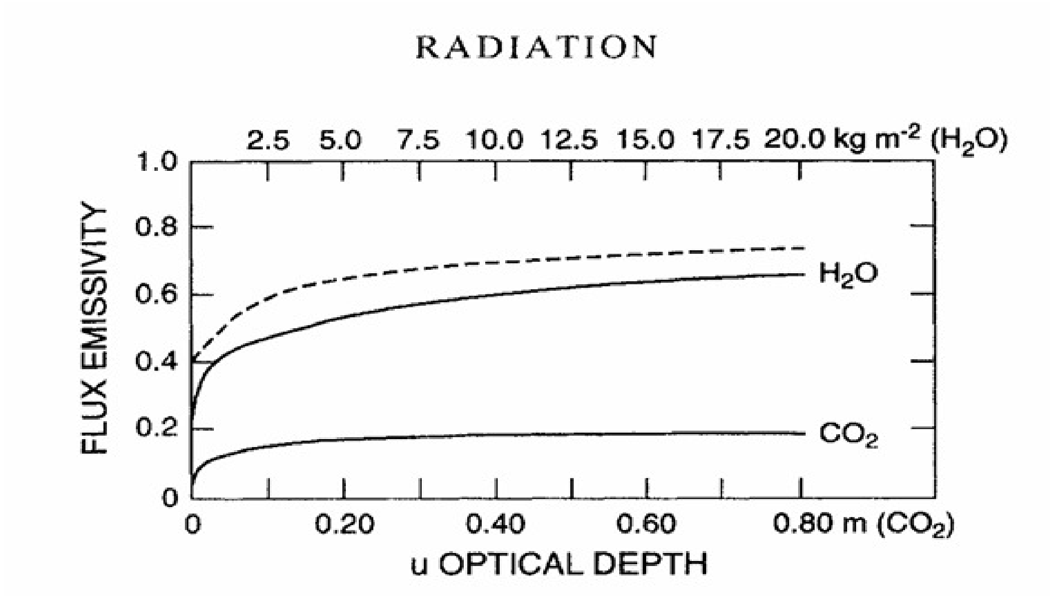

Figure 4.6: Experimental observations of emittance of pure water vapour, of carbon dioxide, and of an atmospheric mixture of CO_2 (0.032 %) and H_2O (mixing ratio 5 g kg^{-1}) as a function of optical depth for temperature of $10^o c$ and pressure of 1013 mb [3].

water vapour and carbon dioxide. When the effects of H_2O and CO_2 are combined, care must be taken to avoid doubling the contribution from overlapping absorption bands.

Because the temperature and the absorbing gases are not analytical functions of height, equations 4.5.11 and 4.5.12 may be integrated graphically by dividing the atmosphere in finite layers and by summation over the layers. This allows us to calculate the divergence of the net longwave radiation and, consequently, whether the layers are heating or cooling, and to what degree. From conservation of energy we find,

$$\rho c_p \frac{\partial T}{\partial t} = -\frac{\partial}{\partial z}(F_{il} + F_{el}) = -\frac{\partial}{\partial z} F_{nl}[3]. \tag{4.5.12}$$

Under clear skies, the divergence of the net longwave radiation typically cools the air in the atmospheric boundary layer by $1-3$ K per day. When clouds are present, the divergence of net longwave radiation is small in the mixed layer below the clouds, because the clouds usually approximate black bodies. At cloud top, however, the cooling may be very large because the cloud droplets may radiate directly to space through the atmospheric window from about 8 to 12 μm [3, 27, 28].

Chapter 5

IMPACT OF AIR-SEA INTERACTION ON WEATHER AND CLIMATE CHANGE

5.1 Air-Sea Feedback

The surface energy budget of the ocean mixed layer of which sensible heat flux, latent heat flux, and precipitation heat flux are a parts dictates the temperature of the sea surface, which by virtue of its role in determining the stability of the lower atmosphere and the upper ocean dictates the fluxes across the interface. It may be noted that the surface ocean mixed layer of the ocean refers to the depth up to which surface turbulence plays an important role in mixing so that the density is approximately the same as the surface. The entire mixed layer is active in transferring heat to the ocean atmosphere interface. This forms a feedback loop between sea surface temperature, stability of the lower atmosphere and the upper ocean, and the air-sea fluxes. Therefore determination of air-sea fluxes is a form of diagnosis of the coupled ocean-atmosphere processes. Similarly, the fresh water flux is part of the ocean surface salinity budget that dictates the stability of the upper ocean, which feeds back to the airsea fluxes.

The influence of cloud radiation interactions on sea surface temperature is an important aspect of air-sea feedback mechanisms. Clouds can have both warming and cooling effects on the sea surface temperature. They influence the Earth's radiation balance by reflecting incoming solar radiation and trapping outgoing long wave radiation. In the summertime, especially over regions like the North Pacific around $35°N$, positive cloud feedback can occur. During summer, increased cloud cover can lead to more trapped longwave radiation. The increased cloud cover can reduce the amount of solar radiation reaching the ocean surface, but the net effect of this can still result in a positive feedback.

Wind induced surface heat exchange (WISHE) phenomenon is another form of air-sea feedback, which refers to the positive feedback a loop between wind speed and surface heat fluxes. As wind speed increases, the exchange of heat between the ocean surface and the atmosphere, includes both sensible heat and latent heat with wind speed that results in increased fluxes leading to cooling of the sea surface temperature as the wind speed increases. The increased heat fluxes lead to more evaporation and consequently to a cooling of the sea surface temperature.

The negative correlation between wind speed and sea surface temperature are apparent at large spatial scales across the subtropical oceans as well as in mesoscale features. As wind speed increases, the rate of heat transfer from the ocean surface to the atmosphere also increases. This is primarily due to enhanced evaporation, which absorbs and removes heat from the ocean surface. The increased heat loss from the ocean surface leads to a decrease in sea surface temperature. Essentially, the higher wind speeds contribute to a greater rate of cooling at the sea surface. This cooling effect due to higher wind speeds is a part of a feedback mechanism. In tropical storms or hurricanes, for example, this interaction can significantly impact the storm's intensity and development. The asymmetry of the inter-tropical convergence zone (ITCZ) in the Northern Hemisphere, particularly in the Eastern Pacific and Atlantic Oceans, can indeed be explained through the wind induced surface heat exchange (WISHE) phenomenon. In the Eastern Pacific and Atlantic, strong trade winds and associated cooling effects can result in lower sea surface temperature [20, 21].

5.2 Their Global Distribution

Direct measurements of air-sea fluxes are indeed essential for understanding and accurately modeling the exchanges between the ocean and the atmosphere. These measurements though sparse, provide the foundational data needed to develop, calibrate and verify parameterizations schemes used in larger-scale models. Direct measurements offer real world data that can be used to calibrate and validate models and parameterizations schemes. Without accurate direct measurements models may rely on assumptions or approximations that could lead to errors. Direct measurements help capture the variability of air-sea fluxes due to changes in atmospheric and oceanic conditions, which is essential for creating accurate predictive models. Parameterizations schemes are simplifications used in models to represent complex processes.

A commonly used parameterizations scheme for air-sea fluxes is the bulk-aerodynamic formulae. The bulk-aerodynamic formulas are indeed a fundamental approach in parameterizing air-sea fluxes. These formulas are used to estimate the fluxes of momentum, heat, and moisture between the atmosphere and the ocean. The key idea is to use measurable state variables (such as wind speed, temperature, and humidity) to estimate these fluxes based on the concept of air-sea interaction. The momentum flux or wind stress is calculated using the wind speed at a certain height above the sea surface.

$$\tau = \rho_a C_d U^2, \tag{5.2.1}$$

where τ is the wind stress or momentum flux, ρ_a is the air density, C_d is the drag coefficient, and U is the wind speed. The drag coefficient C_d is a dimensionless number that represents the drag per unit surface area that the wind exerts on the water. The drag coefficient is influenced by various factors, including wind speed, surface roughness, and atmospheric stability.

The sensible heat flux, which is the transfer of heat from the sea surface to the atmosphere or vice versa, is calculated as:

$$Q_h = \rho_a C_p C_h U(T_s - T_a), \tag{5.2.2}$$

where Q_h is the sensible heat flux, ρ_a is the air density, C_P is the specific heat capacity of air, C_h is the heat transfer coefficient, U is the wind speed, and T_s and T_a are the sea surface and air temperature respectively.

The latent heat flux, which represents the transfer of water vapor, is given by:

$$Q_e = \rho_a L_v C_e U(q_s - q_a)[22], \tag{5.2.3}$$

where Q_e is the latent heat flux, ρ_a is the air density, L_v is the latent heat of vaporization, C_e is the transfer coefficient for water vapour, U is the wind speed, q_s is the specific humidity at the sea surface temperature and air pressure, and q_a is the specific humidity of the air above the water surface. In all these formulas, the transfer coefficients C_d, C_h, and C_e play a main role and are often derived from empirical observations or parameterized based on the stability of the atmosphere and sea surface conditions.

Wind stress (τ) is proportional to the mean wind shear between the surface and a reference height, typically 10 meters above the surface. This shear is indicative of how wind speed changes with height and is crucial for estimating the stress exerted by the wind on the ocean surface. Sensible heat flux (Q_h) is proportional to the vertical temperature gradient between the sea surface and a reference height, typically 2 meters above the surface. This gradient drives the heat transfer between the ocean and the atmosphere. Large values of sensible heat flux are often observed along the western boundary currents of middle latitude oceans, such as the Gulf Stream in the North Atlantic. This phenomenon occurs during winter when cold continental air moves over relatively warmer ocean currents.

The Bowen ratio is the ratio of sensible heat flux to latent heat flux, indeed exhibits a latitudinal gradient that reflects the varying climatic conditions across different regions. Bowen Ratio (β) It is given by:

$$\beta = \frac{Q_h}{Q_e}[23], \tag{5.2.4}$$

where β, is the Bowen ratio, Q_h, is the sensible heat flux, and Q_e, is the latent heat flux

The positive correlation between sea surface temperature and surface wind speed indicates a dynamic mechanism where reduced atmospheric static stability due to warmer sea surface temperature enhances vertical momentum mixing, leading to accelerated surface winds [22, 23, 24].

5.3 The Common Practices to Diagnose This Feedback

The variability of sea surface temperatures in the middle latitudes is significantly influenced by atmospheric variations. This interaction can be understood by examining the correlations between atmospheric heat fluxes and changes in sea surface temperature over time. Atmospheric variations, particularly latent heat flux and sensible heat flux, play a major role in driving sea surface temperature variability in the middle latitudes. These fluxes are affected by atmospheric conditions such as wind speed, temperature, and humidity. The analysis typically shows that atmospheric heat fluxes have a significant impact on sea surface temperature variability. The interaction between atmospheric heat fluxes and sea surface temperature can vary based on regional factors such as ocean currents, local weather patterns, and seasonal changes. In some regions, positive correlations might be observed, indicating that warmer sea surface temperature are associated with higher heat fluxes, while in others, the relationship might be more complex or show negative correlations.

In the North Atlantic, the variability of sea surface temperature is strongly influenced by atmospheric conditions. For instance, during winter months, strong Westerly winds and associated heat fluxes can lead to significant changes in sea surface temperature, impacting regional climate and weather patterns. Latent heat fluxes, driven by high evaporation rates, often contribute to sea surface temperature cooling, while sensible heat fluxes can either warm or cool sea surface temperature depending on atmospheric temperature gradients.

In the Pacific ocean, atmospheric variations such as the position and strength of the jet stream, as well as variations in wind patterns, significantly impact sea surface temperature variability. The interaction between atmospheric heat fluxes and sea surface temperature can lead to pronounced regional effects, including variations associated with phenomena like the El Nio-Southern Oscillation (ENSO). Similar to the North Atlantic, latent and sensible heat fluxes in the Pacific ocean affect sea surface temperature, with latent heat fluxes often cooling sea surface temperature through increased evaporation and sensible heat fluxes influencing sea surface temperature based on temperature differences between the ocean and atmosphere.

The relationship between atmospheric variability and sea surface temperature in the Eastern equatorial Pacific ocean is a key aspect of understanding climate dynamics, particularly in the El Nio-Southern oscillation (ENSO). In the Eastern equatorial Pacific, atmospheric variability is significantly influenced by underlying sea surface temperature variations. Changes in sea surface temperature affect atmospheric conditions such as pressure, wind patterns, and humidity, leading to variations in weather and climate. The Eastern equatorial Pacific is a critical region for ENSO, where sea surface temperature anomalies (both warm during El Nio and cool during La Nia) drive significant changes in atmospheric conditions and global weather patterns. The simultaneous correlation between latent heat flux and the tendency of sea surface temperature in the Eastern equatorial Pacific is typically positive. This means that as sea surface temperature increases, latent heat flux also tends to increase. The positive correlation between latent heat flux and sea surface temperature tendency indicates a feedback mechanism where rising sea surface temperature enhance evaporation and latent heat flux, which can further affect atmospheric conditions and potentially amplify sea surface temperature anomalies.

Higher sea surface temperature leads to increased evaporation rates because warmer waters have higher energy available for evaporation. This results in higher latent heat fluxes, as more heat is transferred from the ocean to the atmosphere. Increased latent heat flux contributes to atmospheric moisture and cloud formation. This, in turn, can influence local and regional weather patterns, including precipitation and storm activity [25, 26].

5.4 Ocean Surface Salinity Budget

The processes that contribute to the ocean surface salinity budget includes precipitation, evaporation, formation and melting of sea ice, river runoff, and storage and transport below the ocean surface. Analogously to the surface energy budget equation,

$$F_{Qo}^{PR} = \rho_s C_{ps} P_s (T_{la} - T_o) - \rho_s L_{il} P_s \tag{5.4.1}$$

where F_{PR} is the net longwave radiative (irradiance) flux, Qo is the net heat flux at the ocean surface, ρ_s is the density of water at the surface, C_{ps} is the specific heat capacity of sea water, P_s is the surface pressure, T_{la} is the air temperature near the surface, T_o is the surface water temperature, and L_{il} is the latent heat of evaporation at the surface.

We can write a general ocean surface salinity budget for either the air/ocean interface or the ice/ocean interface.

$$F_{so}^{net} - F_{so}^{adv} - F_{so}^{ent} = (-\rho_l P_r - \rho_s P_s + E_o - R)s_o + \rho_i \frac{dhi}{dt}(s_o - s_i), \tag{5.4.2}$$

where F_{so}^{net}, is the net flux of surface ocean salinity, F_{so}^{adv} is the advective flux of salinity, F_{so}^{ent} is the entrainment flux of salinity, ρ_l is the density of the freshwater from the river, ρ_s, is the density of sea water, P_s is the surface precipitation, E_o is the evaporation from the ocean surface, R is the river runoff, ρ_i is the density of sea ice, hi is the thickness of the ice, s_o is the surface salinity of the ocean, s_i is the salinity of the ice, and dt is the change with respect to time. where the subscripts o and i refer to the ocean surface value and the sea ice value, respectively. The terms in equation 5.4.1 are the flux density of salt at the surface, in units $kgPsum^{-2}s^{-1}$. A positive term denotes an increase in surface salinity. F_{so}^{net} is the ocean storage term and F_{so}^{adv} and F_{so}^{ent} represent the transport of salt into the ocean mixed layer via fluid motions and turbulence, respectively.

The first two terms on the right hand side of equation 5.4.1 represent the freshening associated with rain and snowfall. Precipitation acts as a negative flux of salt, since the near surface ocean water is diluted by the precipitation as if there were a loss of salt. Because of the momentum of rain as it reaches the ocean surface, some of the drops submerge into the ocean, depending on the size (and thus terminal velocity) of the drops. The raindrops that are not submerged remain to form a freshwater skin. The surface salinity depression associated with rainfall normally does not exceed 5 psu. Snowflakes do not submerge into the ocean because of their low density. The term E_o in equation 5.4.1 is the evaporative flux of water from the ocean surface. Evaporation of water from the ocean increases the concentration of salt in the ocean and thus the salinity. The evaporative flux of water E_o, is given by,

$$E_o = \rho(wq_v)_o = -\frac{F_{Qo}^{L_H}}{L_{lv}}, \tag{5.4.3}$$

where E_o is the evaporation flux, ρ_w is the density of water, q_v is the specific humidity, Q_o is the heat flux, L_H is the heat of vaporization, and L_{lv} is latent heat of vaporization.

The term R in equation 5.4.1 arises from the transport of freshwater from river runoff into the ocean. Typically, about 40 present of the precipitation that falls on a continent is transported into the global ocean through river runoff. River runoff acts analogously to precipitation by diluting the ocean water and acting as negative salt flux. River runoff influences the surface salinity directly only in coastal regions. In equation 5.4.1 the term hi denotes sea ice thickness, Pi the density of sea ice, and si the salinity of sea ice. The term $\frac{dhi}{dt}$ reflects the growth or melting of sea ice. Because growing sea ice rejects the salt back into the melt, sea ice freezing acts effectively as salt source for the ocean mixed layer. Melting sea ice freshens the ocean mixed layer, and thus acts as a negative salt flux. Over most of the global ocean, away from the coast and from the high latitude regions that are influenced by sea ice and snowfall, we can write a simplified version of 5.4.1 as,

$$F_{so}^{net} - F_{so}^{adv} - F_{so}^{ent} = (-\rho_1 P_r + E_o)s_o[4], \qquad (5.4.4)$$

where F_{so}^{net}, is the net salinity flux at the ocean surface, F_{so}^{adv} is the advective flux of salinity, F_{so}^{ent} is the entrainment flux, ρ_1 is the density of the river water, E_o is the evaporation from the ocean surface, and s_o is the surface salinity of the ocean.

The evaporative flux shows a general decrease with latitude away from the equator, analogous to the surface latent heat flux. Precipitation exceeds evaporation in the equatorial regions, and thus there is a net freshening of the ocean surface. In the subtropical latitudes, the evaporation term dominates the surface salinity budget, and there is a net positive surface salinity flux. However, the zonal average hides important meridional differences and differences between ocean basins. For example, comparing the values of the Atlantic and Pacific Oceans (Table) shows that the net surface salinity flux is positive (salinating) in the Atlantic and negative (freshening) in the Pacific.

Most of the fiver runoff occurs in the Northern Hemisphere, because of the larger land mass. The Amazon River provides the largest source of river runoff. The large amount of river runoff into the Arctic Ocean results in an upper ocean salinity of around 30 *psu*. Most of the freshwater input to the Southern Ocean comes from glaciers via iceberg calving, basal melting under ice shelves, and wall melting.

	$\dot{P}$	$\dot{E}_0$
Arctic Ocean	97	53
Atlantic Ocean	761	1133
Indian Ocean	1043	1294
Pacific Ocean	1292	1202
All oceans	1066	1176

Table 5.1: Precipitation and evaporation rates ($mmyr^{-1}$) for four ocean basins [4].

The salinity budget is a dynamic interplay between various sources and processes influencing salinity changes. Factors such as geography, climate particularly latitude, and the presence of river runoff or sea ice can drastically affect local and ocean basin salinity levels [4].

5.5 The Planetary Boundary Layer

Planetary Boundary Layer (PBL) is the lower most portion of the atmosphere, adjacent to the earth's surface, where maximum interaction between the Earth surface and the atmosphere takes place and thereby maximum exchange of physical properties like momentum, heat, moisture, and salinity etc., are taking place. Exchange of physical properties in the planetary boundary layer is done by turbulent motion, which is a characteristic feature of planetary boundary layer. Turbulent motion may be convectively generated or it may be mechanically generated.

If the lapse rate near the surface is super adiabatic, then planetary boundary layer becomes positively buoyant, which is favourable for convective motion. In such case planetary boundary layer is characterized by convective turbulence. Generally over tropical oceanic region with high sea surface temperature this convective turbulence occurs. If the lapse rate near the surface is sub adiabatic then the planetary boundary layer is negatively buoyant and it is not favourable for convective turbulence. But in such case, if there is vertical shear of horizontal wind, then vortex (cyclonic or anti cyclonic) sets in, in the vertical planes in planetary boundary layer. This vortex motion causes turbulence in the planetary boundary layer, known as mechanical turbulence.

If the planetary boundary layer is positively buoyant as well as, if vertical shear of the horizontal wind exists, then both convective and mechanical turbulence exits in the planetary boundary layer. The depth of the planetary boundary layer is determined by the maximum vertical extent to which the turbulent motion exists in planetary boundary layer. On average it varies from few cms to few Kms. In case of thunderstorms planetary boundary layer may extend up to tropopause. Generally at a place on a day depth of planetary boundary layer is maximum at noon and in a season it is maximum during summer [3, 18].

5.6 The Dynamical Implications of Air-Sea Interactions

El Nio-Southern oscillation (ENSO) is probably the most widely recognized low frequency cyclic event associated with atmospheric circulation. This phenomenon is a coupling of the atmosphere and ocean with a non-periodic recurrence of 3 to 7 years that has been suggested as the strongest source of natural variability in the Earths climate system. It is particularly relevant for hydroclimate because it highlights the link between the climate system and the hydrologic cycle and the recurrence of floods and drought and other extreme conditions. In simple terms, El Nino is an oceanic circulation component consisting of a huge pool of anomalously warm water along the equator in the Eastern Pacific ocean. La Nina is another oceanic component identifiable as anomalously cool sea surface temperature in the equatorial Pacific ocean. The Southern oscillation is an alteration of the trade wind circulation resulting in a reversal of flow over the tropical ocean. Typical ENSO events tend to develop during summer to early fall, mature during winter, and terminate the following spring.

El Nio(warm phase) the trade winds, which usually blow from East to West across the tropical Pacific weaken or even reverse direction. This leads to the Eastward movement of warm surface waters from the Western Pacific to the central and Eastern Pacific, including along the coast of South America. The walker circulation the normal atmospheric circulation pattern weakens or reverses, which causes a shift in the location of low and high pressure systems. The warming of the Eastern Pacific causes an area of lower atmospheric pressure to develop there, resulting in heavy rainfall along the Western coast of the Americas. Conversely, the Western Pacific experiences drier conditions, as the typical rising air and rainfall shift to the central and Eastern Pacific.

El Nio events can also cause shifts in the jet stream, leading to altered weather patterns around the world. For example, some regions in North America may experience warmer or wetter-than-usual conditions, while other areas may face drought or wildfires. During an El Nio, the air pressure in the Western Pacific becomes higher, while in the central Pacific it is lower. This change in pressure is linked to the reversal of the trade winds and the warmer sea surface temperatures.

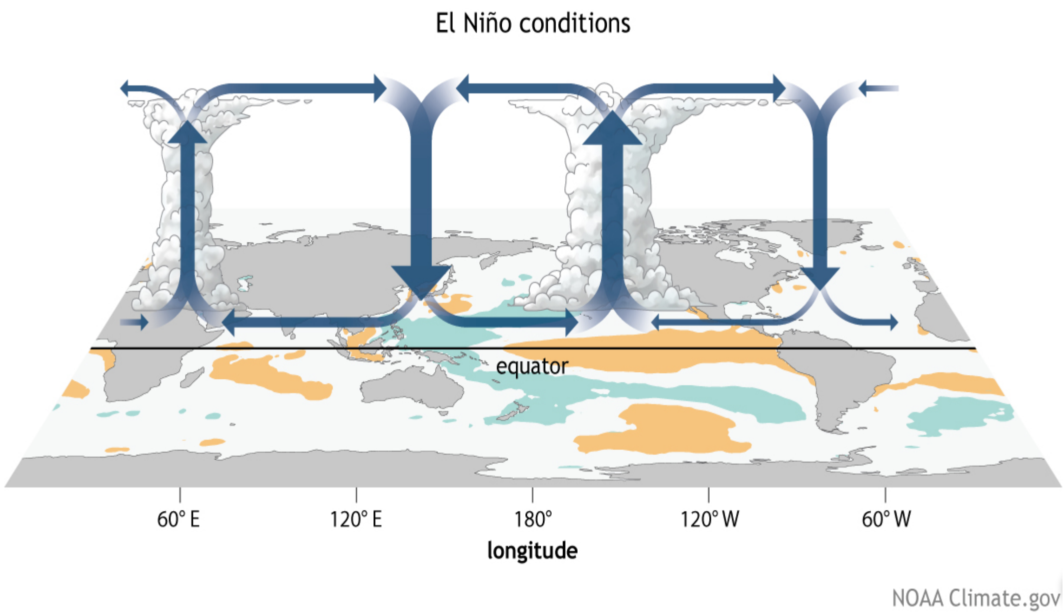

Figure 5.1: An El Nio phase occurs when the trade winds weaken, El Nio is the warmer phase of ENSO [29].

La Nia(Cool Phase) is the opposite of El Nio. During a La Nia event, the trade winds strengthen, pushing more warm water towards the Western Pacific and allowing for enhanced upwelling of cold water along the coast of South America. This leads to cooler than normal sea surface temperatures in the central and Eastern Pacific.

The walker circulation becomes stronger, reinforcing the normal atmospheric pattern of rising air in the Western Pacific and sinking air in the Eastern Pacific. The Western Pacific, including Southeast Asia and Australia, experiences wetter-than-usual conditions and more tropical cyclones.

The Eastern Pacific, including the Western coasts of North and South America, tends to experience drier conditions and reduced rainfall. During a La Nia the higher atmospheric pressure over the Western Pacific (near Darwin) and lower pressure over the central Pacific (near Tahiti). This reinforces the trade winds and enhances the cool conditions in the Eastern Pacific.

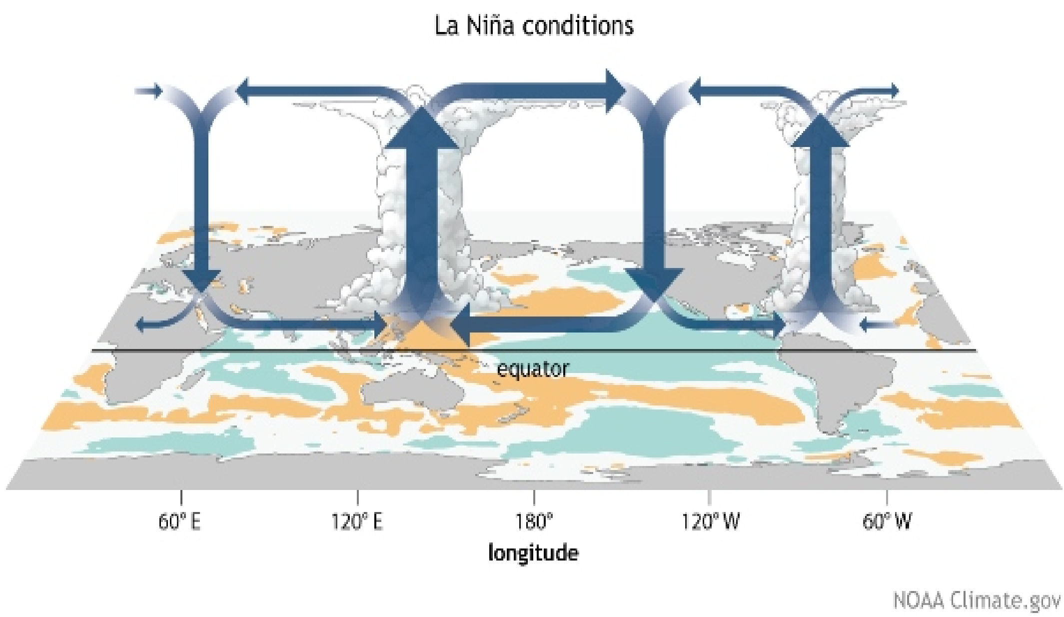

Figure 5.2: The La Nia phase occurs when trade winds are stronger than normal, La Nia is the cooler phase [29].

Neutral phasein the neutral phase of ENSO, the conditions in the tropical Pacific are neither in an El Nio nor a La Nia state. The trade winds blow from East to West, and the sea surface temperatures in the central and Eastern Pacific are closer to average. The walker circulation and atmospheric pressures are also relatively normal.

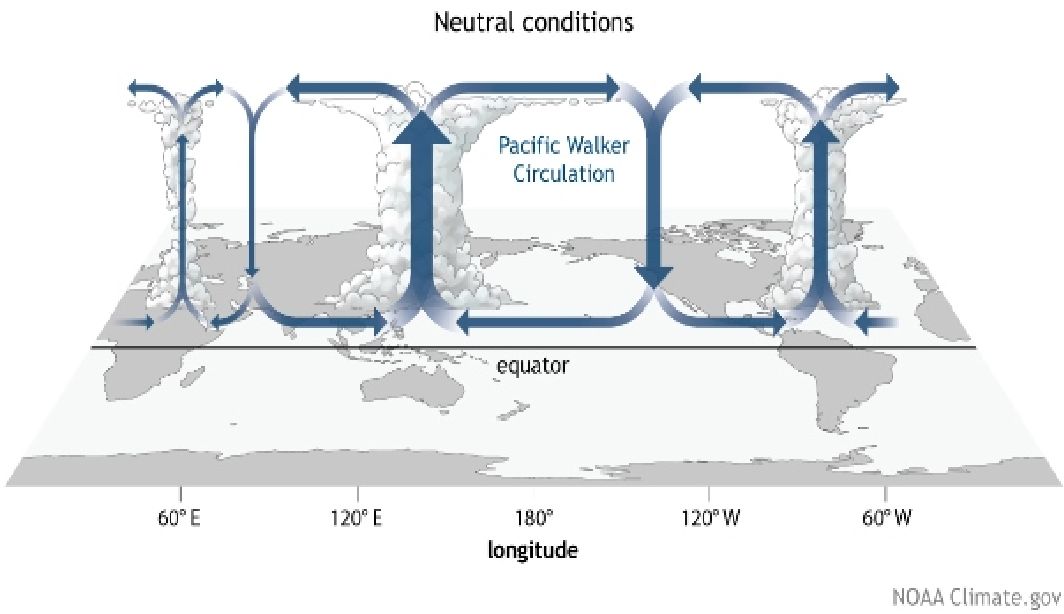

Figure 5.3: The neutral phase is the walker cell functioning normally, the tropical Pacific Ocean's temperatures, winds, rainfall, and convection are close to average [29].

Under normal conditions, coastal waters supplied by the upwelling are cool and nutrient rich. If the trade winds relax and the walker circulation weakens, the entire equatorial Pacific adjusts, and El Nio conditions start to develop. First, the upwelling in the East weakens and cuts off the supply of colder deeper waters. A strong warming and deepening of the thermocline follows in the East, the Western warm pool starts to extend towards the central equatorial Pacific, and the convective area in the atmosphere shifts Eastward. La Nia conditions are the opposite of El Nio with anomalously cold conditions in the Eastern equatorial Pacific and a stronger walker circulation.

ENSO can cause temperature anomalies, with some areas experiencing warmer than average conditions (e.g., parts of North America during El Nio) and others cooler than average (e.g., Southeast Asia during El Nio). ENSO influences the distribution and amount of rainfall, with wet conditions often occurring in areas like the central Pacific and dry conditions in regions like Australia and the Western Pacific. ENSO can contribute to flooding, droughts, wildfire, and hurricanes.

Rainfall in East Africa is tightly linked to the El Nio Southern Oscillation (ENSO), with more rainfall during El Nio years and severe droughts in La Nia years. Wind speeds show the inverse pattern with an intensified windy season during La Nia years in contrast to reduced conditions in El Nio years.

A key process maintaining normal conditions in the equatorial Pacific as shown below in the figure 5.4, the three states of the ENSO cycle. Normal conditions are characterized by a tilted thermocline the location between the cold deeper water and the warm surface water at around 150 m depth at the equator, resulting in cold sea surface temperatures in the East and a warm pool in the West. Cold surface conditions generate high sea level pressure, whereas low pressure and convection prevail over the Western warm pool. This promotes trade winds which drive warm surface waters Westward and shallow the thermocline in the East. This is the lower branch of the atmospheric walker circulation.

The ocean circulation is composed of a wind driven component and a thermohaline component. The wind-driven circulation dominates the surface currents, but it is largely restricted to the topmost few hundred meters. The circulation deeper in the oceans is dominated by the slower thermohaline circulation. The timescale in which a parcel completes a circuit of this so called thermohaline circulation is on the order

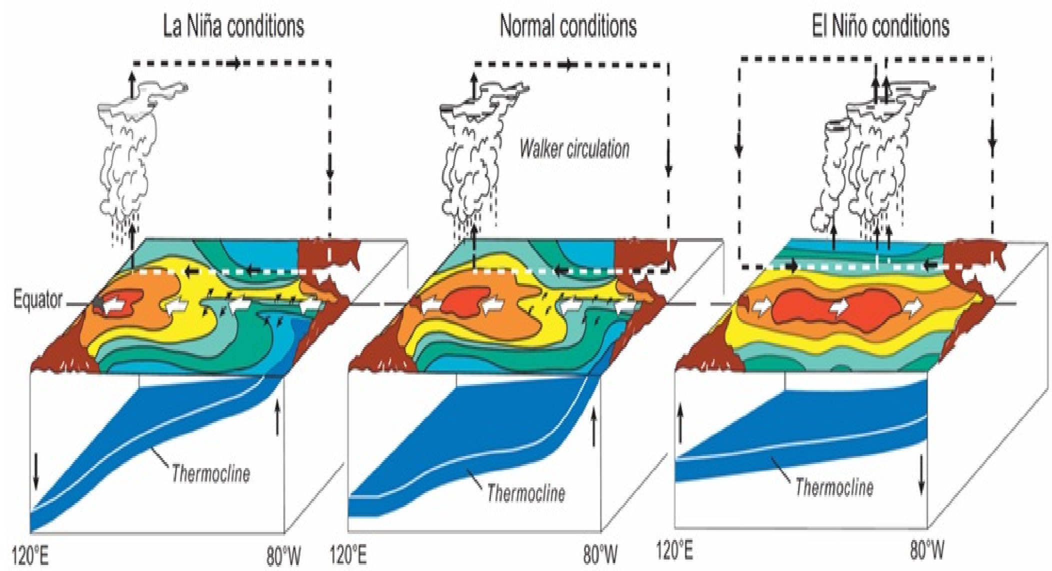

Figure 5.4: The three states of the ENSO cycle showing the location of the main convective system and circulation in the atmosphere, sea surface temperature anomalies and the location of the ocean thermocline [5].

of hundreds of years

By generating ocean waves, surface winds transfer horizontal momentum from the atmosphere into the ocean. The waves stir the uppermost layer of the ocean, mixing the momentum downward. The momentum, as reflected in the distribution of surface currents the pattern of surface winds at subtropical latitudes and at subpolar latitudes.

The walker circulation is the East to West flow of air across the tropical Pacific. The walker circulation is an atmospheric circulation pattern in the tropical Pacific, driven by temperature differences between the Western Pacific and the Eastern Pacific. Air rises over the warm waters of the Western Pacific, moves Eastward at high altitudes, then descends in the Eastern Pacific, creating a pattern of low pressure over the West and high pressure over the East. This circulation helps reinforce the trade winds and ensures that there is consistent upwelling of cold water off the South American coast.

The trade winds push surface waters from East to West across the tropical Pacific ocean. As a result, warm surface waters are pushed toward the Western Pacific. This movement of warm water is accompanied by the upwelling of cooler water along the Eastern Pacific coast, especially near the equator. This upwelling brings nutrient rich cold water to the surface, supporting marine life and maintaining a relatively cooler temperature in the Eastern Pacific.

The Indian ocean is the third largest ocean, covering about 20 % of the Earth's water surface. It is bounded by Asia to the North, Africa to the West and Australia to the East. Surface winds over the tropical Indian ocean are dominated by the seasonally reversing monsoon circulation, consisting of a broad arc originating as a Westward flow in the winter Hemisphere, crossing the equator, and curving Eastward to form a belt of moisture-laden Westerly winds in the summer Hemisphere, as depicted for the Northern Hemisphere (i.e., boreal) summer. The monsoon is driven by the presence of India and Southeast Asia in the Northern Hemisphere subtropics versus the Southern Hemisphere subtropics. Surface temperatures over land respond much more strongly to the seasonal variations in solar heating than those over ocean.

Monsoon is seasonal reversal of wind patterns, typically characterized by heavy rainfall during the summer months, caused by large scale atmospheric circulation changes driven by differential heating between land and ocean, resulting in a distinct wet and dry season in a region. The seasonally reversing Asian monsoon circulation influences weather patterns and precipitation in East Africa, resulting in two distinct rainy seasons: the long rains in boreal spring (April-May) and the short rains in boreal fall (October-November).

During the long rains, the East African region often experiences consistent and abundant rainfall, primarily driven by the moist air transported from the Indian ocean. This period is characterized by strong monsoonal winds that enhance convection and precipitation, contributing to favorable conditions for agriculture and water resources.

In contrast, the short rains occurring in the fall season are marked by greater variability. The October-November short rains have been found to correlate strongly with the Southern oscillation, especially during La Nia events. In these cases, cooler sea surface temperatures in the central and Eastern Pacific can disrupt the typical weather patterns, leading to below normal rainfall in parts of East Africa.

Summer monsoon occurs approximately from June to September and is characterized by heavy rainfall in South Asia and parts of East Africa. The Southwest monsoons bring moisture-laden winds from the Indian ocean, leading to abundant precipitation. Winter monsoon occurs from December to February, with dry air flowing from the Northeast. This monsoon results in significantly less rainfall across the region.

The Indian ocean, the Ethiopian highlands, and the monsoon systems significantly influence the climates and weather patterns of East Africa and South Asia. Here's an overview of how the Indian ocean and the Asian monsoon system interact with the climate of East Africa and specifically Ethiopia. The Ethiopian Highlands, also known as the Roof of Africa, significantly influence local weather patterns. The moist air from the Indian ocean is forced upwards, resulting in precipitation on the windward

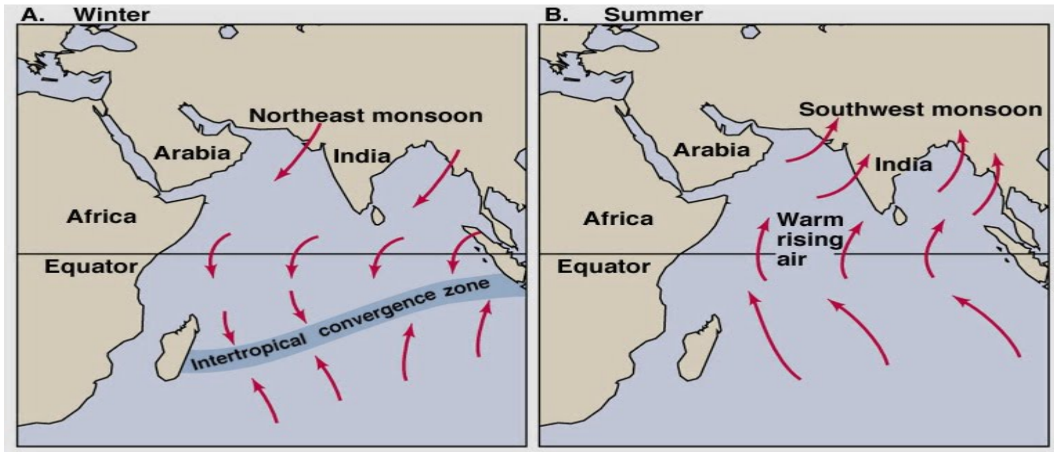

Figure 5.5: Indian ocean winter and summer monsoon [30].

side of the mountains. These phenomena in the Indian ocean can lead to significant variances in rainfall patterns in East Africa, including Ethiopia. For instance, El Nio years might lead to excessive rainfall and flooding, while La Nia can result in drought condition [14, 19, 28].

Chapter 6

SUMMARY AND CONCLUSION

In this study we have seen the impact of ocean-atmosphere interaction on weather and climate. Atmosphere ocean interactions are a consequence of exchanges between the Earth's atmosphere and oceans. These exchanges can be explained in terms of fluxes of energy, momentum and mass of substances, governed oceanic and atmospheric dynamics. Heat transfer occurs through several mechanisms includes radiation, sensible heat, and latent heat transfer. The surface energy budget, which encompassed both incoming and outgoing energy fluxes, dictated whether the ocean is gained or losed heat.

The mass interchanges between ocean-atmosphere boundary layers by the methods freshwater movement, investigated the dynamic equilibrium between evaporation, precipitation, runoff, melting of ice and its impact on ocean salinity, mixing and circulation. The fresh water flux F = E - P, is a major determinant of stability in the water column. The study explored the exchange of gases, between the ocean and atmosphere is importance of greenhouse gases like CO_2, N_2O, and CH_4, as well as O_2, within the context of climate change and global cycle.

We have also seen the radiation transfer processes at air-sea boundary layers, focused on both the oceanic and atmospheric boundary layer models. Radiation models of oceanic boundary layer fundamental properties of seawater, temperature and salinity that influence its permittivity. Interrelations of microwave and infrared radiation fluxes the transfer of thermal and electromagnetic energy between the ocean and atmosphere. The radiation is measured through satellite, aircraft, and shipboard observations, with specific equations quantified the contributions from various atmospheric and oceanic sources. The terrestrial radiation the dynamics of longwave radiation emitted from the sea surface and its interaction with atmospheric layers are detailed.

Finally we have discussed the air-sea feedback mechanisms, the factors such as sensible heat flux, latent heat flux, and precipitation fluxes contribute to the surface energy budget of the ocean mixed layer. The ocean surface salinity budget outlines the processes contributed to salinity changes, emphasized the impacts of precipitation, evaporation, river runoff, and sea ice dynamics. The planetary boundary layer (PBL) emphasizes its role in mediating exchanges between the ocean and atmosphere, impacted by wind and temperature gradients. The dynamical implications of air-sea interactions, used the Asian monsoon system and the El Nio-Southern Oscillation (ENSO) as examples of how these interactions can modulate regional and global.

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