CHAPTER 1

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1.1 GENERAL INTRODUCTION

Boundary Layers are one of the widely encountered phenomena by researchers working in the field of aerodynamics, hydraulics, fluid mechanics and heat transfer, as well as by the meteorologists and physical oceanographers. In a broad sense, a boundary layer can be defined as the layer of a fluid flow in the immediate vicinity of a material surface in which significant transfers of momentum, heat and mass between the boundary surface and the fluid occurs. In the atmospheric context, it has never been easy to define precisely what the boundary layer is. Nevertheless, the Atmospheric Boundary Layer (ABL) also referred to as the Planetary Boundary Layer (PBL) can be defined as the lowest layer of the air directly above the earth's surface that forms as a consequence of the interactions between the atmosphere and the underlying surface (land or water) over time scales of a few hours to about one day (Arya, 1988; Garrett, 1992; Stull, 1988). Within the ABL, the influence of surface roughness, heating, and other properties are quickly and efficiently transmitted through the mechanism of turbulent mixing. The turbulent nature of the ABL is one of its most conspicuous and important features. Almost the entire biosphere is either contained in, or depends on, the ABL. The ABL transfers heat and moisture from the surface and disperses them both horizontally and vertically, effectively air-conditioning the biosphere and providing a conduit for energy to power weather systems on all scales. In many aspects, the ABL can be considered as the circulatory system of the biosphere.

The ABL over the oceanic surface is commonly referred to as the Marine Atmospheric Boundary Layer (MABL). More than two-thirds of the earth’s surface being composed of water in the form of oceans, seas and lakes, complete understanding of the structural characteristics of the MABL becomes imperative for meteorological and atmospheric studies leading to global and long-term weather predictions (Edwards, 2000). The structure and characteristics of the MABL and its
interactions with the oceanic surface and the overlying free atmosphere are known mainly as a result of a synergistic combination of observational studies, numerical and laboratory simulations, and dimensional analysis (Lenschow, 1986). Planning observational studies requires an understanding of instrument capabilities and limitations, strategies for designing and implementing a field study, and techniques for data handling, analysis and synthesis. It may be mentioned that the amount of funding required for open ocean experimental work is rather large: a group of scientists, engineers, technicians and researchers has to work together for a number of years in order to gain experience in obtaining undisturbed and reliable measurements. In the last four to five decades, our understanding of the structural characteristics of the MABL has improved vastly through observational and modelling studies over the tropical oceanic regions during various field experiments [to mention a few: International Indian Ocean Expedition (IOE, 1962-1965); Atlantic Trade-wind Experiment (ATEX, 1969); GARP Atlantic Tropical Experiment (GATE, 1974); Joint Air-Sea Interaction Experiment (JASIN, 1978); Humidity Exchange Over the Sea (HEXOS, 1986) Programme, Tropical Ocean Global Atmosphere (TOGA) and TOGA-Coupled Ocean Atmosphere Response Experiment (TOGA-COARE, 1985-1993)].

However, most of the oceanographic field experiments in the last four to five decades were confined to the tropical Atlantic and Pacific Ocean. Due to lack of coordinated field experiments, historical observations over the tropical Indian Ocean are a few. In this context, the meteorological data collected over the data-sparse region of the western tropical Indian Ocean during the various phases of the Indian Ocean Experiment (INDOEX) attains prime significance (Mitra, 2001, 1999; Mitra et al., 2000; Ramanathan et al., 2001; 1996). INDOEX, being a multi-disciplinary international field experiment, had several objectives focused towards developing a comprehensive analysis off the interactive role of radiation, clouds, and anthropogenic and continental aerosols transport for a better understanding of the role of aerosols on natural and climatic forcing and its feedback on regional and global climate from experimental observations over the western tropical Indian Ocean (Ramanathan et al., 2001). Though the field experiment was aimed mainly towards aerosols and radiations studies, it essentially included a meteorological component for assessment of the magnitude of solar absorption at the surface and in the
troposphere including Inter Tropical Convergence Zone (ITCZ) cloud system. The campaign also had an objective to assess the role of ITCZ in the transport of trace gases and pollutants and their radiative forcing. The present thesis aims at improving our understanding of the structural characteristics of the MABL and its associated dynamics over the data-sparse region of western tropical Indian Ocean by making use of the meteorological data collected onboard Oceanic Research Vessel (ORV) Sagar Kanya and Kaashidhoo Climate Observatory (KCO, located in the Republic of Maldives) during the Intensive Field Phase of the INDOEX (hereafter referred to as INDOEX, IFP-99) campaign. Section 1.9 of this chapter gives the details of the objectives and theme of the thesis. In brief, the objective of the study presented in this thesis is three fold: (1) study of air-sea interaction processes through estimates of air-sea exchange parameters and role of wind speed in the estimates (2) study of thermodynamic structure of the MABL in vertical in relation to degree of convection and (3) study of diurnal variability of the ABL over a tiny island of Kaashidhoo through observations and testing the applicability of mesoscale models for simulation of thermodynamic profiles over the region.

1.2 ATMOSPHERIC BOUNDARY LAYER (ABL)

The ABL can be considered as the region of the lower troposphere, which is directly influenced by the Earth’s surface. The winds in the ABL are affected by the frictional forces acting across the surface, whilst the thermal and moisture properties of the air will be influenced by heat and evaporative fluxes from or across the surface. Its top is marked by a limit on vertical mixing from the surface.

1.2.1 DEPTH OF THE ABL OVER LAND AND THE OCEAN

In the atmospheric context, it has always been quite difficult to mark the top of the ABL. In a wind tunnel, the thickness or depth of the boundary layer is defined as the distance from the surface where mean velocity reaches 99% of its ambient free stream velocity. However, such definitions are of no use in the case of the ABL, because the measurements of mean wind profiles are not that precise and also because such profiles vary rarely monotonically with height. The most direct measures of the ABL depth are provided by upward looking SODAR (acoustic sounder, Sound Detection And Ranging) and LIDAR (Light Detection And
Ranging), which measure the height to which turbulent fluctuations in temperature and refractive index extend. Turbulence sensors mounted on a research aircraft, high tower or tethered balloon also provide other means of measuring the boundary layer depth (Plate, 1982). More often, the ABL depth is also estimated from the measured temperature and humidity profiles obtained from standard radiosonde releases (Arya, 1988; Garratt, 1992; Plate, 1982; Stull, 1988). In daytime unstable and convective conditions, it has been found that the ABL usually extends all the way up to the base of the lowest inversion, which can be easily inferred from the radiosonde-measured temperature and humidity profiles. Above the inversion base, potential temperature increases and specific humidity decreases fairly rapidly with height, while these quantities remain very nearly uniform throughout the mixed layer below the inversion base. Thus, the height of the lowest inversion base is, for all practical purposes, considered equal to the height of the unstable or convective ABL (Arya, 1988; Garratt, 1992; Plate, 1982; Stull, 1988). During nighttime stable conditions, the ABL is generally identified with the surface inversion layer. Based on the time and location of observations, the ABL height at a given location varies over a wide range (several tens of meters to several kilometers) and also depends on the rate of heating or cooling of the surface, strength of winds, the roughness and topographical characteristics of the surface, large scale vertical motion, horizontal advects of heat and moisture, and many other factors.

Over the oceans, the boundary layer depth varies relatively slowly in space and time. The Sea Surface Temperature (SST) changes very little over a diurnal cycle because of the tremendous mixing within the top of the ocean. Also, water has a large heat capacity, meaning that it can absorb a large amount of heat from the sun with relatively little temperature change. Thus, a slowly varying SST means a slowly varying forcing into the bottom of the boundary layer. Most changes in boundary layer depth over oceans are caused by synoptic and mesoscale processes of vertical motion and advection of different air masses over the sea surface. In general, the tendency of the ABL is to be thinner in high-pressure regions than in low-pressure regions over land as well as the oceans. The subsidence and low-level horizontal divergence associated with synoptic high-pressure moves boundary layer air out of the high towards lower-pressure regions. Shallower depths are often associated with cloud-free regions. In low-pressure regions, such as the ITCZ, the
upward motions carry boundary layer air away from the ground to large altitudes throughout the troposphere. It is difficult to define a boundary layer top for these situations. In these kinds of occasions, the cloud bases are quite close to the inversion base and a measurement of the mean height of the cloud base would give a good estimate of the ABL depth (Plate, 1982; Stull, 1988). Thus, the depth of ABL may be thinner in low-pressure regions than in high pressure. The definition of the MABL to be consistent with that used throughout, must relate to the structure of small-scale turbulence and so the top may not necessarily correspond with either a low level inversion or cloud top. In literature, since the MABL is often associated with clouds, several definitions of the MABL top can be found. Thus, the MABL may be taken to be that part of the lower atmosphere that is directly coupled to the surface by turbulent transfer, including fair-weather cumulus and stratocumulus clouds (Garratt, 1992; Stull, 1988). In this case, the boundary layer tends not to be well mixed (as deduced from mean temperature profiles) above cloud base. In contrast, the ABL top may be defined to be at or near cloud base. This usually corresponds both to the top of a relatively well-mixed layer and to the height at which turbulent fluxes of momentum and heat becomes negligible (Garratt, 1992; Nicholls, 1985). For the near-neutral or slightly unstable MABL where surface buoyancy is weak, the mixed layer height \( h \) tends to be much less than the inversion level and, in the presence of cumulus convection, it corresponds approximately to the cloud base height. Only in moderate to strong instability, when buoyancy is strong we find the mixed layer height \( h \) to correspond to the inversion base (irrespective of cloud base), as is usual during the afternoons over land in moderately unstable conditions.

1.2.2 General Structure of the ABL

Following sunrise on a clear day, the continuous heating of the earth's surface by the sun and the resulting thermal mixing in the ABL cause the ABL depth to increase steadily throughout the day and attain a maximum value in the late afternoon. Later in the evening and throughout the night, on the other hand, the radiative cooling of the ground surface results in the suppression or weakening of turbulent mixing and consequently in the shrinking of the ABL depth. Thus, the ABL structure and its depth waxes and wanes in response to the diurnal heating and
cooling cycle. Over land surfaces in high-pressure regions the ABL has a well-defined structure that evolves with the diurnal cycle.

Figure 1.1. Schematic of the vertical structure of the lower atmosphere and its various layers during neutral and unstable conditions (Not to the scale). (Based on Arya, 1982 and Garratt, 1992).

Figure 1.1 shows a schematic of the vertical structure of the lower atmosphere over a land surface and its various layers during neutral and unstable conditions. In figure, several regions of the lower atmosphere are identified, including the interfacial (or roughness) layer, the inner (or surface) layer and the outer (or Ekman) layer. In the diagram, \( h \) is the boundary layer depth, \( z \) is height and \( z_\theta \) is the aerodynamic roughness length. The molecular transport dominates over the turbulent transport within the roughness sublayer. Within this layer, the turbulence and mean profiles are strongly affected by the structure of the roughness elements. In the surface layer, wind and stress exhibit negligible rotation with height. The inertial sublayer is the region within which the velocity profile in neutrally buoyant conditions is logarithmic. In the outer region, the flow shows little dependence on the nature of the surface and, in the atmosphere, the Coriolis force due to the earth's rotation is also important. This region is referred to as Ekman layer. In contrast to the above, flow in the inner layer (or surface layer) is mainly
dependent on the surface characteristics and is little affected by the rotation of the earth. The transition between the surface layer and Ekman layer is not abrupt, but is characterized by an overlap region. The influence of the surface is directly felt in the interfacial sublayer, which is the layer of air within and just above the roughness elements comprising the land or sea surface. In this layer, molecular diffusion is an important process by which heat and mass are exchanged between the surface and the air.

The boundary between the turbulent ABL and the free atmosphere above is usually quite sharp and distinct, particularly during unstable and convective conditions, but is highly variable both in time and space. The ABL depth may be thought of as an average height of this contorted interface over a period of the order of an hour, so that the effects of large eddies, such as horizontal roll vortices and convective cells, are also smoothed out in the averaging processes (Plate, 1982).

1.2.3 Factors Influencing the Structure of the ABL

The ABL is essentially driven by large-scale atmospheric flows such as geostrophic, thermal, and gradient winds. Above the top of the boundary layer the flow is generally assumed to be frictionless and essentially in geostrophic balance. The boundary layer height \( h \) is probably one of the most important parameters, which influence the ABL structure. The dimension of the largest eddies are essentially fixed by \( h \). Wind shears, turbulence intensities and fluxes in the ABL also strongly depend on \( h \). Due to these properties, many investigators use \( h \) as one of the basic length scaled for describing the ABL structure. The other important factor influencing the boundary layer is the surface drag. Surface drag essentially depends on the roughness characteristics of the underlying surface and is primarily responsible for the characteristic wind profile with monotonic increase of wind speed with height in the lower part of the ABL. Thus the surface roughness exerts strong influence on the mean profile and turbulence structure in the surface layer and also in the upper part of the ABL. Besides friction, the underlying surface also influences the ABL structure. On a clear day, the surface absorbs a part of the solar radiation and warms up relative to the air above. This temperature differential usually gives rise to a variety of convective circulations, which transfer heat and other substances from the surface to the atmosphere. The radiative warming of the
surface relative to air results in an unstably stratified ABL in which buoyancy forces generate turbulence in addition to that generated by wind shear. On a clear night, on the other hand, the surface cools down relative to the air above, resulting in a stably stratified or surface inversion layer. It can easily be seen that buoyancy inhibits vertical motions in such a layer. Consequently in a stable ABL, the vertical turbulent transfers are greatly reduced and at times suppressed altogether, especially under light winds.

Thus, the diurnal cycle of heating and cooling of the surface is one of the important factors in determining the thermal stability of the ABL flow and, hence, its mean flow and turbulence structure. Among other factors, which can influence the vertical structure of the ABL, are the presence of fog and stratus layers within the ABL, and advection of heat and moisture. The entrainment of the air from above into the turbulent boundary layer can also influence the structure of the ABL. Due to entrainment at the top of the unstable ABL, significant amounts of heat and momentum flux may occur near the top of the ABL.

1.3 DIURNAL EVOLUTION OF THE ABL

The structure of the ABL turbulence over land surface is strongly influenced by the diurnal cycle of surface heating and cooling, and also by the presence of clouds. Generally, the ABL structure over the land surface in high-pressure regions can be classified into three broad categories (Stull, 1988): (1) Convective Boundary Layer; (2) Residual Layer and (3) Stable Boundary Layer. The unstably stratified ABL, or Convective Boundary Layer (CBL), occurs when strong surface heating produces thermal instability or convection in the form of thermals and plumes, and when upside-down convection is generated by cloud-top radiative cooling. In strongly unstable conditions driven by surface heating, the outer layer in particular is dominated by convective motions and is often referred to as the mixed layer. In contrast, the stably stratified ABL, or Stable Boundary Layer (SBL) occurs mostly at night. The Residual Layer (RL) is often observed after the sunset hours during the transition from CBL to SBL. Figure 1.2 gives the schematic representation of the evolution of ABL with time.
Different regions of the ABL are: SL (Surface Layer); CBL (Convective Boundary Layer); RL (Residual Layer); SBL (Stable Boundary Layer); CL (Cloud Layer); EZ (Entrainment Zone) and FA (Free Atmosphere).

1.3.1 CONVECTIVE BOUNDARY LAYER (CBL)

(A) SURFACE LAYER (SL)

The lowest one-tenth or so of the ABL, which is very close to the earth's surface and in which the earth's rotational or Coriolis effects can be ignored is termed as the Surface Layer. It is characterized by the sharpest variations of wind speed, temperature, and other meteorological parameters with height. It is also characterized by intense small-scale turbulence generated by surface roughness or friction and, in the case of a heated surface, also by the thermal convection. The turbulence in this layer is mainly due to wind shear, which is generated by surface frictional force, generally known as mechanical turbulence. This small-scale turbulence is largely responsible for the vertical exchanges of momentum, heat and mass to and from the surface. Within the surface layer, the vertical fluxes of these quantities are found to remain nearly constant with height and vertical variations of these quantities are observed to be within 10% of their surface values. This layer, therefore, is also referred to as a constant flux layer. The height of the surface layer is typically 50 m, but may vary over a wide range (5 to 200 m) as does the ABL height (Plate, 1982). The ability of surface layer to transport momentum, sensible heat, water vapour and other constituents is of fundamental importance in all studies.
related to land surface/atmosphere as well as ocean/atmosphere interaction processes, including parameterization in global circulation models. Due to the above-mentioned properties of the surface layer, it has received far greater attention from the researchers than has the outer part of the ABL.

(B) **FREE CONVECTION LAYER (FCL)**

In this layer, buoyant convection dominates compared to mechanical turbulence and is said to be in a state of the free convection. This Free Convection Layer (FCL) is formed under conditions of large heat flux and light wind. When the boundary layer reaches a free convective condition, the forced convection condition is almost negligible. Also the gradient of wind speed and potential temperature are almost negligible in this layer. The turbulence in this layer close the surface might feel the influence of the ground much more than that of the influence of the capping inversion.

(C) **MIXED LAYER (ML)**

The Mixed Layer (ML) is a part of the CBL, which is characterized by intense mixing in a statically unstable situation where thermals of warm air rise from the ground. The turbulence in the ML is usually convectively driven; the convective sources include heat transfer from a warm ground surface, and radiative cooling from the top of the cloud layer. Turbulence in the ML tends to mix heat, moisture, and momentum uniformly in the vertical. Even when convection is the dominant mechanism, there is usually wind shear across the top of the ML that contributes to the turbulence generation. The ML reaches its maximum depth in the afternoon. Decrease in the mixing ratios with height in the ML reflects the evaporation of soil and plant moisture from below, and the entrainment of drier air from above. Decrease in moisture across the top of the ML is often used together with potential temperature profiles to identify the ML top.

(D) **ENTRAINMENT ZONE (EZ)**

The Entrainment Zone (EZ) is the region of statically stable air at the top of the ML, where there is entrainment of air from the free atmosphere downward and overshooting thermals upward. Alternatively the EZ can also be defined as that region above the top of the ML, where the buoyancy flux is negative. EZ is
characterized by increased vertical shear in wind speed, potential temperature and humidity. In this region, negative heat flux, increase in momentum flux and positive moisture flux are seen. When the ML is shallow during morning hours over land, the EZ is proportionally shallow. As the ML grows, so does the EZ thickness. Thin EZs are expected for large temperature changes across the ML top, because thermals will not penetrate as far and entrainment will be slow. Thick EZs are expected with more intense ML turbulence when convection is vigorous.

1.3.2 Residual Layer (RL)

After the sunset, the thermals cease to form allowing turbulence to decay. The resulting layer of air is sometimes called the Residual Layer (RL). The RL is neutrally stratified resulting in turbulence that is nearly of equal intensity in all directions. Therefore, the smoke plumes emitted into the RL tend to disperse at equal rates in the vertical and lateral directions. The cooling rate is more-or-less uniform throughout the depth of the RL, thus allowing the RL virtual potential temperature profile to remain nearly adiabatic. The RL does not have direct contact with the ground. The RL often exists for a while in the mornings before being entrained into the new ML.

1.3.3 Stable Boundary Layer (SBL)

The ABL can become stably stratified whenever the air above the earth's surface becomes warmer than the surface. This type of layer often forms at night over land, and is often known as Stable (or Nocturnal) Boundary Layer (SBL or NBL). Such a layer can also form during the day, as long as the underlying surface is colder than that of the air. Such a situation can often occur during warm-air advection over a colder surface. The SBL is characterized by statically stable air with weaker, sporadic turbulence. The statically stable air tends to suppress turbulence, while the enhanced wind shear due to the developing nocturnal jet tends to generate turbulence. With the progress of night, the bottom portion of the RL is transformed by its contact with the ground into a SBL. As opposed to the daytime ML that has a clearly defined top, the SBL has a poorly defined top that smoothly blends into the RL above. The top of the ML is defined as the base of the stable layer: while the SBL top is defined as the top of the stable layer or the height where turbulence
intensity is a small fraction of its surface value. Pollutants emitted into the stable layer disperse relatively little in the vertical; however, they disperse more rapidly in the horizontal.

1.4 Stability of the ABL

As discussed in the previous section, one of the striking features of the ABL is its diurnal evolution with time. One contrasting feature in the nature of the ABL during typical daytime and nighttime is its stability. During daytime the ABL is unstable in nature, while it becomes stable with the advancement of night.

When we talk of atmospheric stability, we are referring to a condition of equilibrium. To explore the behaviour of rising and sinking air, we have to consider a small volume of air, also referred to as the parcel of air. At the earth's surface, the parcel has the same temperature and pressure as the air surrounding it. Suppose we lift the air parcel up into the atmosphere. Since the pressure decreased with increasing altitude, the air pressure surrounding the air parcel lowers, allowing the air molecules inside to push the parcel walls outward, expanding the parcel. Such an expansion of the air parcel causes its temperature to decrease. On the other hand, if we push an air parcel downward, the compression of air parcel due to pressure difference with the surrounding air, the temperature of air parcel increases. Hence, a rising parcel of air expands and cools, while a sinking parcel is compressed and warmed. Thus, we see that the variations of temperature and humidity with height in the ABL lead to density stratification with the consequence that an upward- or downward-moving parcel of air will find itself in an environment whose density will, in general, differ from that of the parcel, after accounting for the adiabatic cooling or warming of the parcel. In the presence of gravity this density difference must lead to the application of a buoyancy force on the parcel, which would accelerate or decelerate its vertical movement. If the vertical motion of the parcel is enhanced, i.e., the buoyancy force accelerates the parcel, the environment is called statically unstable. On the other hand, if the parcel is decelerated, the atmosphere is called stable or stably stratified. When the atmosphere exerts no buoyancy force on the parcel at all, it is considered neutral. In particular, the static stability parameter of the atmosphere $s$ is defined as (Arpe, 1988):
\[ s = \left( \frac{g}{T_v} \right) \left( \frac{\partial T_v}{\partial z} \right) \]  

---(1.1)

where \( g \) is the acceleration due to gravity; \( T_v \) is the virtual temperature. Based on equation (1.1), the atmospheric stability can be qualitatively divided into three broad categories:

1. Unstable, when \( \gamma < 0 \); or \( \frac{\partial T_v}{\partial z} < -\Gamma \)
2. Neutral, when \( \gamma = 0 \); or \( \frac{\partial T_v}{\partial z} = -\Gamma \)
3. Stable, when \( \gamma > 0 \); or \( \frac{\partial T_v}{\partial z} > -\Gamma \)

From equation (1.1), another important feature can be seen that vertical motions are generally enhanced in an unstable atmosphere, while they are suppressed in a stably stratified environment. The lower atmospheric layer can also be characterized on the basis of virtual temperature gradient or lapse rate, relative to adiabatic lapse rate \( \Gamma \) as follows:

- Super adiabatic, when \( \frac{\partial T_v}{\partial z} < -\Gamma \)
- Adiabatic, when \( \frac{\partial T_v}{\partial z} = -\Gamma \)
- Subadiabatic, when \( 0 > \frac{\partial T_v}{\partial z} > -\Gamma \)
- Isothermal, when \( \frac{\partial T_v}{\partial z} = 0 \)
- Inversion, when \( \frac{\partial T_v}{\partial z} > 0 \)

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Figure 1.3. Schematic of the various stability categories on the basis of virtual temperature gradient (After Arya, 1988).
Figure 1.3 gives a schematic of the above stability categories in the lower part of the ABL. In the figure, the surface temperature is assumed to be the same and virtual temperature profiles are assumed to be linear for convenience only; actual profiles are usually curvilinear.

1.5 Features of the ABL over Oceanic Surface

A precise definition of the MABL may vary with the author, and with the application also; generally speaking, however, the MABL is considered as the turbulent region adjacent to the oceanic surface. The MABL is an important element in the influence of the ocean on weather or climate, since heat or water vapour from the ocean's surface must pass through the MABL before escaping to the free atmosphere. The MABL, typically differ from the ABL over the land surface in a number of ways. Some of the important features of the ABL over oceanic surface, which makes the MABL studies more significant, are listed below:

- Air is usually moister over the sea surface, compared with the land, with a typical relative humidity of about 75% to 100%. The diurnal cycle of meteorological parameters tends to be weak (though not negligible), since surface energy fluxes get distributed over a considerable depth (10 m to 100 m+) of water, which has a heat capacity as much as hundreds of times as large as the ABL.

- Air-sea temperature differences tend to be small, except near the coasts. The air tends to be 0 to 2°C cooler than the water. This is because the boundary layer air is usually radiatively cooling, and some of this heat is supplied by sensible heat fluxes off the ocean surface. However, if the air temperature is much lower than the SST, vigorous convection will reduce the temperature difference, and except where there are large horizontal gradients in SST, horizontal advection cannot maintain the imbalance. Hence the surface layer is nearly neutral over almost all of the oceans.

- Diurnal variation in the SST, and hence diurnal forcing of the MABL, are small mainly because of the large heat capacity of the oceanic mixed layer. Due to the small air-sea temperature difference, the Bowen ratio (= sensible heat flux/latent heat flux) tends to be small (typically 0.1 in
the tropical oceans, and more variable in midlatitudes); latent heat fluxes are 50 to 200 Wm\(^{-2}\), while (except in cold air outbreaks off cold landmasses) sensible heat fluxes are 0 to 30 Wm\(^{-2}\). In contrast to the land surface, over much of the ocean, the surface heat flux does not play a large role in determining the boundary layer structure.

- Oceanic surface containing water is very much denser than the air above it. The density of air varies with temperature, pressure, and humidity, typical surface values being 1.2 - 1.3 kg.m\(^{-3}\), whereas the sea is some 800 times more dense (typically 1025 kg.m\(^{-3}\) at the sea surface). Therefore the interface between air and water is very stable because of the strength of the gravitational restoring force when it is displaced from its equilibrium position. Typical displacements observed in surface waves are about a metre or so.

- Because of the stability of the air-sea interface, the two media do not mix in any significant way (whitecaps and spray are only found close to the interface), so transfers of properties between the two media must take place through a well-defined interface. Important exchanges of energy, mass, and momentum occur across these water surfaces, which influence atmospheric and oceanic circulations over a whole spectrum of time and spatial scales.

- Over 95% of the MABL contain clouds. The only exceptions are near the coasts, where warm, dry continental air is advected over a colder ocean, and in some regions (such as the eastern equatorial Pacific cold tongue and some western parts of the major subtropical oceans) in which air is advecting from warmer to colder SST, tending to produce a more stable shear-driven boundary layer which does not deepen to the Lifting Condensation Level (LCL) of surface air. Cloud profoundly affects the boundary layer dynamics over the oceanic surface.

- The MABL is an important element in the influence of the ocean on weather or climate, since heat or water vapour from the ocean’s surface must pass through the MABL before escaping to the free atmosphere. Virtually all water vapour that reaches the free atmosphere is first transported through the MABL turbulent and advective processes.
• Turbulent transport of momentum down through the ABL to the surface is the important momentum sink for the atmosphere. Wind stress on the sea surface is the primary energy source for ocean currents. The latent heat stored in water vapour accounts for 80% of the fuel that drives atmospheric motions.

• The primary energy source for the whole atmosphere is solar radiation, which is absorbed at the earth's surface (ocean as well as land surface) and transmitted to the rest of the atmosphere by ABL processes. Pollutants, water vapour, or cool air can be trapped within the MABL until they escape into the free atmosphere above.

• Over large areas, the oceanic surface is relatively uniform and large excursions from near-neutral conditions are restricted to relatively small regions of the global oceans.

• About 90% of the net radiation absorbed by oceans cause evaporation, amounting to the evaporation of about 1 m of water per year over the earth's ocean area. Dynamic interaction occurs between water waves and surface layer turbulence.

1.5.1 Sea-Surface Temperature (SST)

One of the most important characteristics of the sea surface is its temperature ($T_s$). As distinguished from most land surfaces, open sea and ocean surface are characterized by a remarkable temporal and spatial homogeneity of temperature. It is primarily due to large heat capacity and efficient mixing processes in the upper oceanic mixed layer. The SST depends on a number of oceanic, atmospheric, and other factors. The meteorological factors affecting the SST are the net radiation to or from the sea surface, evaporation and precipitation processes, and the sensible heat exchange with the atmosphere. Among the oceanic factors are the mixed layer depth, the intensity of turbulent mixing, the presence of upwelling or downwelling, and the advective heat transport by oceanic currents. Observations of the diurnal variation of $T_s$ from research vessels indicate that the minimum temperature occurs in the morning hours before sunrise and the maximum value occurs in the afternoon. The actual times of minimum and maximum and the range of diurnal variation of the SST depend on the latitude, season, and also on the prevalent weather conditions at
the location of interest. The diurnal range, \( \Delta T_s = T_{\text{max}} - T_{\text{min}} \), is found to decrease with increasing latitude, wind speed and cloudiness. The typical values of \( \Delta T_s \) are 0.5°C in the tropics, 0.25°C in midlatitudes, and less than 0.1°C in high latitudes (Meyn, 1988).

1.5.2 Surface Waves

The sea surface is distinctively different from the land surface in its mobility. The endless, unceasing, up and down movement in the form of waves remain one of the most interesting and fascinating aspects of the sea surface. It is obvious that part of the momentum transfer from the atmosphere to the ocean is via the sea waves. These surface waves are succession of humps and hollows in ever-changing patterns. They are generated by atmospheric winds, astronomical causes (e.g., high and low tides), seismic disturbance, and, on a small scale, by passing ships and boats. The short waves generated and maintained by local winds are called seas, while long waves generated elsewhere and propagated into the region of interest are called swells. Surface waves can be classified according to water depths (e.g., shallow-water and deep-water waves) and also on the basis of the governing force (e.g., gravity waves and capillary waves). The sea state at any location and time depends on the strength and duration of winds, the fetch or distance over which winds blow in a coherent or persistent manner, and the strength and direction of surface currents. Observers have recorded waves of over 20 and even 30 m in height. But more frequently encountered seas have significant heights of less than 2 m and wave periods of less than 10 sec.

1.6 Oceanographic Field Experiments: An Overview

During the 16-17th centuries there have been many maritime activities in different oceanic regions of the globe. However, most of them were for adventuring and trade. Our present knowledge and understanding of the MABL over various oceanic regions of the globe comes from measurements taken during different field experiments. The great scientific oceanographic expeditions, usually involving several disciplines, essentially included meteorological surveys. In the first half of the 20th century, the scientific community focused their research towards air-sea
interaction processes and the features of MABL and tried to derive a balanced description of the global climate. Among the earliest open-ocean measurements were those of Sheppard and Omar (1952), for low wind speeds. There was a growing need for data in higher wind-speed conditions to determine the dependence of drag coefficient ($C_d$) on the wind speed over the oceans. More experiments were conducted (Brocks, 1959; Fleagle et al., 1958; Hay, 1955), but data were still lacking at speeds above 13 ms$^{-1}$. By the end of the 1950s, there was some agreement on the value of $C_d$ but no consensus on its wind speed dependence. It began to appear that open ocean coefficients were slightly lower than those from semi-enclosed waters. Deacon and Webb (1962) reviewed early determinations of air-sea exchange coefficients. In the 1960s an era of large international field projects began, with an emphasis on tropical seas. An important step from marine meteorology towards air-sea interaction was based on observations of pilot balloon ascents made from ships of opportunity the Atlantic Ocean (Husse, 1990). In the beginning studies pertained to determining bulk transfer coefficients on the local scale. However, with the advancement of science, the emphasis shifted to studies of a more complex nature with multi-ship based international experiments. In the late 1970s, climate and global change issues created a need for more precise modelling of air-sea momentum, heat, and moisture fluxes. Offshore oil platforms have created another need to understand air-sea interface fluxes, with an emphasis on waves and on coastal domains. Stull (1988) summarizes the major field experiments that were either completely devoted to ABL measurements, or had a major ABL subprogram (see Table 10.1, Stull, 1988, pp 417 - 419). In the following sub-sections, we present brief details of some major oceanographic field experiments conducted over different oceanic regions over the entire globe during the last five decades prior to INDOEX (1996 - 1999).

### 1.6.1 INTERNATIONAL INDIAN OCEAN EXPEDITION (IIOE)

Due to lack of coordinated and well-planned scientific field experiments over the tropical Indian Ocean region, for a long time, the Indian Ocean remained one of the least explored regions. Ocean studies received impetus only after India's independence with the launching of the International Indian Ocean Expedition (IIOE) during 1962-1965. The expedition was a large, international program, conceived at
the first International Oceanographic Congress in 1959 as a means of gathering information on the least studied ocean in the world. This expedition co-ordinated by the international agencies like United Nations Educational Scientific and Cultural Organization (UNESCO) & Scientific Community of Oceanic Research (SCOR) involved participation of 20 countries with 40 oceanographic research vessels. The expedition led to the generation of huge amount of data and unravelled many features of the Indian Ocean (Wyrski, 1973; 1971). This field experiment yielded the first large data sets of fluxes based on profiles of wind, temperature, and humidity. During IOE, radiosondes from ships and dropsondes from aircraft provided some valuable data to study the structure of MABL over the tropical Indian Ocean. However, the IOE data were not adequate to delineate the different sublayers of the MABL in this region and bring out their characteristic features (Parasnis and Normal, 1993).

1.6.2 BARBADOS OCEANOGRAPHIC AND METEOROLOGICAL EXPERIMENT (BOMEX)

The Barbados Oceanographic and Meteorological Experiment (BOMEX) was carried out over tropical Atlantic East of Barbados in the summer months of May – July of 1969 (Davidson, 1974; Dunckel et al., 1974). During this field experiment intensive oceanographic and meteorological observations including studies of air-sea interface fluxes were carried out from an array of ships. The field operations for this multiagency national study of the ocean-atmosphere system were divided into four observation periods: May 3 to 15, May 24 to June 10, June 19 to July 2, and July 11 to July 28. The first three were devoted to the Sea Air Interaction Program—the BOMEX ‘Core Experiment’—within a 500-km by 500-km square ship array. During the fourth period, the array was extended southward to incorporate the Inter Tropical Convergence Zone. The ‘Core Experiment’ of BOMEX proposed to determine rates of transfer of water vapor, heat, and momentum from the ocean to the atmosphere. Parameters included in this BOMEX dataset were: boundary layer and surface air temperature, wet bulb temperature, humidity, and winds; clouds; visibility; precipitation; sea surface temperature; and waves. Pond et al. (1971) measured the water vapour eddy flux; a mean $C_e$ of 1.36 X 10$^{-3}$ compared well with the results of Kruspe (1977) from the North and Baltic Seas.
1.6.3 **Atlantic Trade Wind Experiment (ATEX)**

The Atlantic Trade Wind Experiment (ATEX), also known as APEX (Der Atlantische Passat Experiment) took place in the Northeast Atlantic trades during February of 1969. The experiment was basically designed to study the development of the boundary layer in the trade winds on their way towards the ITCZ. The experimental setup was primarily constituted three research vessels to form a drifting equilateral triangle with sides of approximately 750 km. The data from the ship array consisted of radiosondes and radar wind observations from each ship at three-hour intervals. Turbulent fluxes were calculated using the 'profile method' and the 'eddy-correlation technique'. Buoy data were also used to correct for ship effects on the deck-level observations. Air-sea fluxes were measured by the profile and the eddy correlation methods on two buoys: a stable, surface-following buoy for profiles and servo-stabilized buoy for eddy fluxes (Dunckel et al., 1974). The neutral drag coefficient from the eddy correlation data was $C_{DN} = 1.28 \times 10^{-3}$, and the evaporation coefficient from profiles $C_{EN} = 1.39 \times 10^{-3}$. These first flux measurements in the open ocean demonstrated the need to consider the influences on profile measurements of waves (Hasse et al., 1978) and of the flow distortion effects of even a slender mast (Wucknitz, 1977). Planning was under way for experiments making use of fast-response wind and temperature sensors, and of the ability to compute spectra and eddy fluxes from time series data.

1.6.4 **GARP (Global Atmospheric Research Program) Atlantic Tropical Experiment (GATE)**

The purpose of the GATE experiment was to understand the tropical atmosphere and its role in the global circulation of the atmosphere. It was the first major experiment of the Global Atmospheric Research program (GARP), whose goal was to understand the predictability of the atmosphere and extend the time range of daily weather forecasts to over two weeks. The experiment took place in the summer of 1974 in an experimental area that covered the tropical Atlantic Ocean from Africa to South America. The work was truly international in scope, and involved 40 research ships, 12 research aircraft, numerous buoys from 20 countries all equipped to obtain the observations specified in the scientific plan. GATE also included extensive radiosonde measurements and observations of the ocean mixed layer. In the
boundary layer subprogramme (Volkov et al., 1982) both radiative and turbulent fluxes were measured. Turbulent fluxes from boom-mounted sensors on ships of the USSR and USA groups, and from a stable buoy operated by a West German group, were intercompared (Hasse et al., 1975). Neutral bulk coefficients for a 10 m reference level $C_{DN} = 0.00125$, $C_{IN} = 0.00134$ and $C_{EN} = 0.00115$ were obtained (Hasse and Seguin, 1977; Hasse et al., 1978). No wind speed dependence of the coefficients was observed since only low and moderate wind speeds were encountered, but processes associated with the cool skin of the ocean surface were found to play a significant role in determining the rate of evaporation and the value of $C_L$ (Hasse, 1971). The cool skin was also found to complicate bulk estimation of the sensible and latent heat flux, particularly for low wind speeds. GATE saw the first measurements of air-sea fluxes from aircraft (Nicholls and Readings, 1979). The combination of surface and aircraft data provided unique insight into the interaction between surface fluxes and convective elements in the boundary layer (Galushko et al., 1975, 1977; Hasse et al., 1978; Khalasa and Businger, 1977; Muller-Glewe and Hinzepeter, 1975; Volkov et al., 1974, 1976).

### 1.6.5 Air Mass Transformation Experiment (AMTEX)

During 1974-1975, a larger scale field experiment named Air Mass Transformation Experiment (AMTEX) was conducted over the South China Sea near Japan. Progress was made in estimating fluxes from a ship. It was found that the fluxes needed to be understood in terms of cospectra and the eddy processes, and a coupling became apparent among surface waves, surface fluxes and turbulent eddies at the scale of the boundary layer (Mitsuta, 1977-1979).

### 1.6.6 Joint Air-Sea Interaction (JASIN) Experiment

Late in the 1970s, a series of international experiments began over the mid-latitudes of the northeast Atlantic and North Sea, taking advantage of research platforms off Germany and the Netherlands. The first of these was the Joint Air-Sea Interaction (JASIN) experiment, conducted in 1978 mainly to the northwest of Scotland. Both air and ocean boundary layers were studied with the intention of closing the momentum and heat budgets. Surface fluxes were measured only from aircraft; the most notable results showed that roll vortices were able to systematically modulate
Changes even in the neutral MABL (Nicholls, 1985; Shaw and Businger, 1985). In the JASIN experiment, some effects on the MABL due to SST changes were observed, particularly in the wind stress and low-level cloud structure as observed from aircraft measurements and ships (Guymer et al., 1983). Businger and Shaw (1984) used the JASIN data and the MABL models to conclude that mesoscale changes in SST could destabilize initially neutral air sea temperature difference and lead to secondary circulations in the MABL.

1.6.7 **MARITIME REMOTE SENSING (MARSEN) EXPERIMENT**

The Maritime Remote Sensing (MARSEN), conducted in the North Sea in 1979, was an integrated remote sensing and surface flux experiment designed to extend physical understanding of the relation of remote sensing signatures to surface waves and wind stress. The remote sensing emphasized synthetic aperture radar and scatterometry using radars on aircraft and on towers. From stress measurements at an offshore mast, Geernaert et al. (1986) found that $C_d$ was larger for steeper waves, and this result was extrapolated to suggest that shallow water waves produce a larger stress than deep-water waves for the same wind speed.

1.6.8 **COASTAL OCEAN DYNAMICS EXPERIMENT (CODE)**

The Coastal Oceanographic Dynamics Experiment (CODE), a multi-institutional research program was basically undertaken to identify and study the important dynamical processes, which govern the wind-driven motion of coastal water over the continental shelf (Reference: http://www-ccs.ucsd.edu/zoo/code/). The initial effort in this multi-year, multi-institutional research program was to obtain high-quality data sets of all the relevant physical variables needed to construct accurate kinematic and dynamic descriptions of the response of shelf water to strong wind forcing in the 2 to 10 day band. A series of the small-scale, densely-instrumented field experiments of approximately four months duration (called CODE-1 and CODE-2) were designed to explore and determine the kinematics, momentum, and heat balances of the local wind-driven flow over a region of the northern California shelf. High-quality data sets of physical variables to construct kinematic and dynamic descriptions of dynamical processes governing wind-driven motion of water
in the 2 to 10 day band over north of the San Francisco bay continental shelf from Pt. Reyes to Pt. Arena.

1.6.9 Genesis of Atlantic Lows Experiment (GALE)
The Genesis of Atlantic Lows Experiment (GALE) was conducted over the eastern United States to study mesoscale and air-sea interaction processes that affect the development of east coast winter storms (Dirks et al., 1988). Observations were collected from a heavily augmented surface and upper air meteorological network, including 4 towers, 6 ocean buoys, two research ships plus "ships of opportunity", from 15 meteorological radars and 7 research radars, from 9 aircraft, and 4 satellites. In 1985-1986 the Canadian Atlantic Storms Program (CASP), coordinated with the USA GALE, studied the growth of winter storms and their interaction with the upper ocean (Smith and Stewart, 1989). This was followed in 1992-1993 by CASP II, a study of the mature stages of explosive cyclogenesis in winter storms and their influence on ocean circulation and sea ice on the New Foundland continental shelf and the Grand Banks (Smith et al., 1994).

1.6.10 Humidity Exchange Over the Sea (HEXOS) Programme
The HEXOS (Humidity Exchange Over the Sea) Program was a part of the Humidity Exchange Over the Sea Main Experiment (HEXMAX), which took place from October 6 to November 28, 1986, near the Dutch coast on and in the vicinity of the Dutch Research Platform Noordwink (MPN). MPN is an offshore research tower, which served as a central platform during HEXMAX. Comprehensive measurements of fluxes of water vapour, heat and momentum, turbulent statistics, sea state, whitecap coverage and aerosol concentration were collected on MPN during HEXMAX. One of the stated goals of the HEXOS program was to determine the contribution of the evaporation of sea spray to the total water vapour flux from the ocean to the atmosphere. HEXOS provided the largest range of wind speeds (up to 19 m/s²) and largest data set for the values of water vapour exchange coefficient ($C_v$). De Cosmo (1991) mentions that there was no increase in the sensible heat and water vapour transfer coefficients with increasing wind speed. Results revealed from the HEXOS data analysis indicate that any increase in water vapour flux due to evaporation of spray may be offset by a reduction in surface evaporation due to a
decrease in water vapour density gradient. The HEXOS dataset provided a unique opportunity for assessing the relationship between aerosol, sea spray, and sensible and latent heat exchange (De Cosme, 1991).

1.6.11 **FRONTAL AIR SEA INTERACTION EXPERIMENT (FASINEX)**

The Frontal Air-Sea Interaction Experiment (FASINEX), basically designed to evaluate the air-sea interaction processes, took place in the subtropical convergence zone of the North Atlantic, about 1000 km east of the Florida coast. The location was chosen because of the well-defined and relatively long-lived sea surface temperature fronts in the region. Fluxes were measured both from aircraft and from a ship near the Gulf Stream southwest of Bermuda, at approximately 27°N, 70°W, in 1986. The fluxes in the unstable ABL on the warm waterside of surface thermal fronts were expected to be larger than those on the cold-water side, but the difference was larger than predicted using coefficients of Businger et al., (1971). It suggested that additional processes might be acting on the marine surface layer. Differences in radar backscatter cross section across a front were more consistent with the variation in wind stress than with the smaller variation in wind speed, giving support to the hypothesis that radar remote sensing can be used to monitor wind stress over the oceans.

1.6.12 **TROPICAL OCEAN GLOBAL ATMOSPHERE (TOGA) AND TOGA-COUPLED OCEAN ATMOSPHERE RESPONSE EXPERIMENT (TOGA-COARE)**

The Tropical Ocean Global Atmosphere (TOGA) program is a major component of the World Climate Research Program (WCRP) aimed specifically at the prediction of climate phenomena on time scales of months to years and to improve our understanding on the tropical oceans and their relationship to the global atmosphere. Underlying TOGA is the premise that the dynamic adjustment of the ocean in the tropics is far more rapid than at higher latitudes. Thus disturbances emanating from the western Pacific Ocean (such as El Nino) may propagate across the basin on time scales of weeks compared to years for corresponding basin-wide propagation at higher latitudes. The significance of shorter dynamic timescales near the equator is that they are similar to those of highly energetic atmospheric modes. This similarity
allows the formation of coupled modes between the ocean and the atmosphere. In order to achieve the TOGA goals, a strategy of large-scale, long-term monitoring of the upper ocean and the atmosphere, intensive and very specific process-oriented studies, and modeling were planned and enacted through a series of national, multinational and international efforts (e.g., WCRP-2, 1986). Modeling activities have been coordinated by TOGA Numerical Experimentation Group (WCRP-4, 1987). The field experiments were conducted over various tropical oceans over the entire globe during 1985 - 1989.

The Tropical Ocean Global Atmosphere Coupled Ocean Atmosphere Response Experiment (TOGA COARE) was a large international field experiment conducted in 1992-1993 to study the atmospheric and oceanic processes over the region of the western Pacific known as the "warm pool". This is the region of warm ocean and atmospheric clouds and precipitation that is linked to the El Nino climate variation (Webster and Lukas, 1992). Collectively, the goals of TOGA COARE are designed to provide an understanding of the role of the warm pool regions of the tropics in the mean and transient state of the tropical ocean-atmosphere system.

1.6.13 SHELF MIXED LAYER EXPERIMENT (SMILE)

The Shelf Mixed Layer Experiment (SMILE) was designed to study the response of the oceanic surface boundary layer over the continental shelf to atmospheric forcing. The SMILE field program was conducted over the northern California shelf between Pt. Arena and Pt. Reyes from mid-November 1988 to mid-May 1989. The field program consisted of five main components: (1) A long term moored array to obtain current, temperature, and conductivity time series observations in the upper ocean over the shelf, (2) A short-term moored instrument deployment to measure the vertical current shear and stratification in the top 6 m of the water column, (3) Shipboard CTD and Acoustic Doppler Current Profiler (ADCP) surveys over the shelf and adjacent slope to map regional water property and current distributions, (4) A long-term moored and coastal meteorological array including one sounding station to obtain time series observations of the atmospheric surface forcing and monitor the structure of the marine boundary layer and (5) Overflights with an instrumented aircraft to measure the spatial structure of the surface wind, wind stress, and heat flux fields under different atmospheric conditions.
1.6.14 Atlantic Stratocumulus Transition Experiment (ASTEX)

The Atlantic Stratocumulus Transition Experiment (ASTEX) was conducted in June 1992 off North Africa in the area of Azores and Madeira Islands. ASTEX was based on two islands and several ships in an area where the total cloud cover (mostly stratocumulus) ranges from 50 to 60%. From a broader perspective ASTEX was designed to provide improved dynamical, radiative, and microphysical models and an improved understanding of the impact of aerosols, cloud microphysics, and chemistry on large-scale cloud properties. A telescoping approach was used in ASTEX to investigate connections between scales ranging from microns to thousands of kilometers. The boundary layers observed in ASTEX were extremely deep (1000-2000 m) and the most common cloud type observed was cumulus clouds under a stratocumulus layer of variable thickness and extent. On nearly all occasions the thermodynamic structure of the boundary layer was very complex. Only in a few cases was the boundary layer well mixed. More typically, the boundary layer was multi-layered; with a subcloud mixed layer (SML) decoupled (both day and night) from a subcloud layer and the cloud layer itself. ASTEX was able to address issues related to the stratocumulus to trade-cumulus transition and cloud-mode selection. An overview of ASTEX is given by Schubert et al., (1992).

1.6.15 Surface of the Ocean, Flux and Interaction with the Atmosphere (SOFIA)

In June 1992, during the ASTEX field program, the SOFIA campaign (Surface of the Ocean, Flux and Interaction with the Atmosphere) was performed to improve the parameterization of the air-sea interface fluxes and of the MABL turbulence. This experiment was conducted in the Azores region, with an instrumented ship and two-instrumented aircraft. Anticyclonic conditions prevailed during the whole campaign, leading to weak-to-moderate winds, and to the presence of broken stratocumulus at the top of the boundary layer. The turbulence structure was described by Réchou et al., (1995) and Réchou and Durand (1997). SOFIA was, in fact, the first realization of a programme, devoted to the study of the ocean-atmosphere interaction, whose most important phase was the SEMAPHORE (Structure des Exchanges Mer-Atmosphère, Propriétés des Hétérogénéités Océaniques: Recherche Expérimentale) field campaign.
1.6.16 Structure des Exchanges Mer-Atmosphère, Propriétés des Hétérogénéités Océaniques: Recherche Expérimentale (SEMAPHORE)

The SEMAPHORE field campaign was performed in the autumn of 1993 in the Azores region. During the field campaign, two instrumented aircraft were simultaneously used for in-situ turbulence measurement in the MABL (Lambert and Durand, 1998), generally in the middle part of the day. The experimental area was situated to the south of the Azores archipelago. The aircraft were based on the island of Santa-Maria (37°N, 25°W), which is the southernmost island. A general description of the strategy and equipment used during the campaign is given by Eymard et al., (1996). Analysis of SEMAPHORE data in relation to the MABL structure and turbulence are presented in two consecutive papers by Lambert and Durand (1999) and Lambert et al., (1999).

1.6.17 Central Equatorial Pacific Experiment (CEPEX)

In 1993, the Central Equatorial Pacific Experiment (CEPEX) tested the hypothesis that convective-cirrus clouds act as a thermostat of ocean-surface temperature in the tropical Pacific (Ramanathan and Collins, 1992; 1993). The CEPEX field studies were conducted by investigators from 15 institutions between March and April 1993. Research results were published on heat transfer from ocean to air and cloud formation, and discuss why the western Pacific warm pool ocean does not heat excessively and create a runaway greenhouse effect. The experimental data also revealed that the atmospheric solar absorption exceeded model predictions by as much as 8% of the solar insolation.

Table 1.1 summarizes the duration and location of all the major field experiments prior to INDOEX campaign described in this section:
### Table 1.1 Summary of Some Oceanographic Field-Experiments

<table>
<thead>
<tr>
<th>Si. No.</th>
<th>Experiment</th>
<th>Duration</th>
<th>Location</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>IIIOE</td>
<td>1962 – 1965</td>
<td>Tropical Indian Ocean</td>
</tr>
<tr>
<td>2</td>
<td>BOMEX</td>
<td>1969</td>
<td>Tropical Atlantic, east of Barbados</td>
</tr>
<tr>
<td>3</td>
<td>ATEX</td>
<td>1969</td>
<td>North-east Atlantic</td>
</tr>
<tr>
<td>4</td>
<td>GATE</td>
<td>1974</td>
<td>Tropical Atlantic Ocean from Africa to South America</td>
</tr>
<tr>
<td>5</td>
<td>AMTEX</td>
<td>1974 – 1975</td>
<td>South China Sea near Japan</td>
</tr>
<tr>
<td>6</td>
<td>JASIN</td>
<td>1978</td>
<td>Mid latitudes of North-east Atlantic and North Sea</td>
</tr>
<tr>
<td>7</td>
<td>MARSEN</td>
<td>1979</td>
<td>North Sea</td>
</tr>
<tr>
<td>8</td>
<td>CODE</td>
<td>1981 – 1982</td>
<td>North of San Francisco bay</td>
</tr>
<tr>
<td>9</td>
<td>GALE</td>
<td>1986</td>
<td>Atlantic Ocean, east of United States</td>
</tr>
<tr>
<td>10</td>
<td>HEXOS</td>
<td>1986</td>
<td>Dutch Coast</td>
</tr>
<tr>
<td>11</td>
<td>FASINEX</td>
<td>1986</td>
<td>North Atlantic, about 1000 km east of Florida Coast</td>
</tr>
<tr>
<td>12</td>
<td>TOGA &amp; TOGA-COARE</td>
<td>1985 – 1993</td>
<td>Various Oceanic regions</td>
</tr>
<tr>
<td>13</td>
<td>SMILE</td>
<td>1988 – 1989</td>
<td>Northern California Shelf</td>
</tr>
<tr>
<td>14</td>
<td>ASTEX</td>
<td>1992</td>
<td>North Africa, Azores and Madeira Islands</td>
</tr>
<tr>
<td>15</td>
<td>SOFIA</td>
<td>1992</td>
<td>Azores Islands, North Africa</td>
</tr>
<tr>
<td>16</td>
<td>SEMAPHORE</td>
<td>1993</td>
<td>Azores Islands, North Africa</td>
</tr>
<tr>
<td>17</td>
<td>CEPEX</td>
<td>1993</td>
<td>Pacific Ocean</td>
</tr>
</tbody>
</table>
1.7 FEATURES OF THE TROPICAL INDIAN OCEAN

1.7.1 HISTORICAL BACKGROUND

Most of the field experiments described in Section 1.6 were confined to the Atlantic Ocean and the Pacific Ocean. The International Indian Ocean Expedition (IIOE) conducted during 1963 – 1965 proved itself to be a landmark in the oceanographic studies over the data sparse region of the tropical Indian Ocean. The expedition led to the generation of huge amount of data and unravelled many of the hitherto unknown features of the Indian Ocean. Analysis of IIOE data revealed that the northwestern Indian Ocean is dominated by the monsoon (Wyrki, 1971). The expedition came to an end in 1966. However, the IIOE data was not adequate to delineate the different sub-layers of the MABL in this region and bring out its characteristic features (Parasnis and Morwal, 1993). In the last two decades, however, there were few research papers published in various international journals describing the circulation or transport of the major tropical currents in the Indian Ocean. Reverdin (1987) and Molinari et al., (1990) provide climatology with respect to the circulation for the tropical Indian Ocean from ship drifts or drifting buoys. Swallow et al., (1991, 1988) and Schott et al., (1988) are mostly constrained to the western area of the tropical Indian Ocean. Woodberry et al., (1989) simulates large-scale upper circulation and provides estimation of transport. Another (Visbeck and Schott, 1992) compares observed data (mooring and ship drift) with results from a model in the western equatorial area. The third one (McCreary et al., 1993) presents, monthly circulation at the surface and in the mixed layer of the entire tropical Indian Ocean. However, as far as MABL studies are concerned, the tropical Indian Ocean remained one of the least explored regions. The urge to explore the tropical Indian Ocean region was in its peak among the scientific community, before it got materialized through the INDOEX.

1.7.2 TROPICAL INDIAN OCEAN REGION

The entire oceanic region over the entire globe can be divided broadly into three major oceans: Pacific Ocean; Atlantic Ocean and Indian Ocean (Figure 1.4).
The tropical Pacific and Atlantic Oceans share many common climatological features such as easterly trade winds, eastward shoaling thermocline, an eastern cold tongue and a northerly ITCZ (Yue et al., 1999). Deep convective activity and associated rainfall over the eastern Pacific and the Atlantic reach a local minimum at the equator and are displaced into the Northern Hemisphere. While the land-sea distribution appears to be responsible for displacing the Pacific and Atlantic ITCZs into the Northern Hemisphere, the actual dynamics and underlying physics is not fully understood. Though the tropical Pacific and Atlantic Oceans have many common features, a comparison of climate variability between the two oceans reveals more differences also (Yue et al., 1999). The Pacific Ocean is dominated by the equatorially symmetric El Nino/Southern Oscillations (ENSO), while the Atlantic ITCZ is controlled by changes in interhemispheric SST. Unlike the Pacific, the tropical Atlantic does not have a dominant mode of variability, but has several modes (e.g., SST dipole pattern, Okumara et al., 2001), which appear to coexist. The SST dipole pattern observed over the Atlantic Ocean is most pronounced on decadal time scales, and is known for its influence on the Atlantic ITCZ. The most well defined boundary layer in the tropical atmosphere is observed over the eastern and central Pacific, where subsidence and low-level easterly trade winds are dominant. The MABL over the tropical Pacific Oceanic region is maintained by heating or cooling effects due to subsidence, radiation and surface fluxes (Betts and Ridgway, 1989).

The tropical Indian Ocean climatology differs significantly with the tropical Pacific and Atlantic Oceans in many aspects. One of the very important features of the tropical Indian Ocean is the seasonal march of ITCZ over both the hemispheres.
It is a widely accepted fact that the equatorial Indian Ocean is the only place in the world where continental aerosols, anthropogenic trace species and their reaction products from the Northern Hemisphere are directly transported to the pristine air of the Southern Hemisphere by a cross equatorial monsoon flow into the ITCZ. Figure 1.5 gives the climatological features of the surface position of the ITCZ over the entire globe for two different seasons (i.e., the month of January and July, adopted from Asnani, 1993).

Figure 1.5. Surface position of the ITCZ over the entire globe based on climatological values. (After Asnani, 1993).

Averaged around the whole globe, the ITCZ lies to the north of the equator during the northern summer and to the south of the equator during the southern summer (Asnani, 1993). However, it has large regional variations. Over the eastern Pacific and the Atlantic Oceanic regions, between longitudes 160°W and 10°E, the ITCZ lies north of the equator throughout the year. Over this region, ITCZ has a small latitudinal oscillation. Over nearly other half of the globe from 10°E across the Indian Ocean to western Pacific Ocean up to 160°W, it is to the north of the equator during the northern summer and to the south of the equator during southern summer. In its annual oscillation, ITCZ migrates farthest away from the equator into the summer continental regions. As can be seen from Figure 1.5, during the month
of July, the ITCZ penetrates deep into the south Asian continents; during the month of January on the other hand, it penetrates deep south into the African continent and also in the Australian land.

During the month of January to March, the Indian sub-continent experiences the winter monsoon and the prevailing winds are northeasterly. This low-level northeasterly flow carries a significant amount of aerosol such as mineral dust, nitrates, sulphate particles and organic aerosols into the tropical Indian Ocean region. These aerosols scatter sunlight back into space, causing a regional cooling effect, and this cooling effect is not properly represented in the existing global climate models. One of the main objectives of the INDOEX programme is to study the natural and anthropogenic climate forcing by aerosols and its feedback on the regional and global climate (Ramanathan et al., 2001; 1996 and the references cited therein). The tropical Indian Ocean region was selected for this experiment because during the Northern Hemispheric winter monsoon (January - March) the region provides a nearly ideal natural laboratory for observing the direct aerosol radiative forcing. Polluted air flows off the Indian and Southeast Asian subcontinents in the northeasterly flow over the Bay of Bengal, and the Arabian Sea, which for this time of year often have large cloud-free regions. Thus, the Indian Ocean region remains the subject of intensive research because of its large potential for growing pollution emissions, due to the rapid growth of human population and industrial development in South and Southeast Asia.

1.8 INDIAN OCEAN EXPERIMENT (INDOEX)

1.8.1 OBJECTIVES OF THE INDOEX CAMPAIGN

Actually emerging from CEPEX (conducted over the Pacific Ocean in 1993), the INDOEX is the result of concerted efforts of many scientists from India, USA, Europe and the island countries Maldives, Mauritius and Reunion. INDOEX, being a multi-disciplinary international field experiment, had several objectives focused towards developing a comprehensive analysis of the interactive role of radiation, clouds, and anthropogenic and continental aerosols transport for a better understanding of the role of aerosols on natural and climatic forcing and its feedback on regional and global climate from experimental observations over the western
tropical Indian Ocean region (http://www-indoex.ucsd.edu). INDOEX observations started in a relatively small way in 1995, onboard a research vessel (Rhoads et al., 1997), which documented the sharp north-south gradients in the pollutants across the ITCZ, giving strong credence to the INDOEX concept. The experiment was also motivated by the increasing awareness of the importance of both anthropogenic and natural aerosols to the earth's radiation budget. The INDOEX campaign was mainly designed to quantitatively assess the effect of the aerosols on the earth's radiation through measuring the radiation budget, aerosol concentration and clouds over the Indian Ocean. Though the INDOEX was primarily aimed towards the studies of aerosols and radiation, the campaign also had a meteorological component associated with the studies of the ITCZ, MABL and atmospheric modelling.

1.8.2 Cruise Track of INDOEX, IFP-99 Campaign

To address the multifaceted objectives of INDOEX, observations over a fairly large domain were necessary. The Indian component of the INDOEX observations were conducted onboard Oceanic Research Vessel (ORV) Sagar Kanya in four consecutive years 1996 - 1999: with different cruise paths covering a broad oceanic region of the western tropical Indian Ocean during the Northern Hemispheric Winter Monsoon period (Mitra, 1999; 2001). The present study is focused on INDOEX, IFP-99 conducted during January 20 - March 12, 1999 spanning a period of 52 days. After meeting the scientific requirements of various participating groups, the cruise track of IFP-99 campaign was finalized. Figure 1.6 depicts the cruise track of the INDOEX. IFP-99 campaign conducted onboard ORV Sagar Kanya. The approximate position of the ship on various Julian days is marked on the cruise track in the figure.

INDOEX, IFP-99 campaign was an extensive one, which involved a variety of ground-, ship, air- and space-based observational systems. From Figure 1.6, one can see that the INDOEX, IFP-99 campaign covered a broad oceanic region of the Indian Ocean and Central Arabian Sea over a latitude range 15°N to 20°S and a longitude range 63°E to 77°E. The track covered between Goa (India) to Port Louis (Mauritius) during January 20 - February 11, 1999 (indicated by arrows in Figure 1.6) is referred to as the forward track, whereas the track between Port Louis (Mauritius) to Goa (India) during February 17 - March 12, 1999 is refereed to as
In Figure 1.6, two meridional and two zonal tracks can be seen. The first meridional track (track-AB, Figure 1.6) approximately along 17°E longitude (hereafter, referred to as leg-1) was traversed between January 20 – February 04, 1999 during the forward track, whereas the second meridional track (track-CD) approximately along 63°E longitude (hereafter, referred to as leg-3) took place between February 18 – March 01, 1999 during the return track of the cruise.

Similarly, there are two zonal tracks, one during the forward track and the other during the return track of the cruise. The first zonal track (track-BC) along 20°S latitude (hereafter, referred to as leg-2) was conducted during February 04 – 11, 1999 during the forward track, whereas the second zonal track (track-DE) along 15°N latitude (hereafter, referred to as leg-4) was covered between March 01 – 06, 1999 during the return track of the cruise.

Figure 1.6. Cruise track of INDOEX, IFP-99 campaign (SK-141).
The strategy of the Indian component of IFP-99 was different from the earlier phases of the INDOEX programme in two aspects: firstly in terms of the time period of campaign and secondly in the finalization of the cruise track (Mitra, 2001). One of the highlights of the IFP-99 campaign was the inclusion of East-West transects both in the pristine region south of ITCZ along the 20°S latitude towards Mauritius (Leg-2 in Figure 1.6) during onward journey of the cruise and again in the Central Arabian Sea eastward along 15°N, towards the Indian subcontinent (Leg-4 in Figure 1.6) during the return journey of the cruise.

1.9 OBJECTIVES AND SCOPE OF THE PRESENT STUDY

Majority of the research work and results presented in this thesis are outcome of the comprehensive analysis of the meteorological data obtained from a ship-borne field experiment. Besides the meteorological data analysis obtained from ORV Sagar Kanya over tropical Indian Ocean during INDOEX, IFP-99 campaign, availability of meteorological data over a tiny island of Kaashidhoo during the campaign has been utilized to study the vertical structure of the ABL over Kaashidhoo through observations and simulations. In brief, the study presented in this thesis aims at improving our understanding of the structural characteristics of the MABL and its associated dynamics over the data-sparse region of western tropical Indian Ocean. The following aspects of MABL over the tropical Indian Ocean are studied in detail in this thesis:

- Underlying physics of air-sea interaction processes and role of wind speed in the estimates of air-sea exchange parameters
- Dynamics of the vertical structure of the MABL in relation to the degree of convection in the vicinity of the ITCZ
- Observational features in the vertical structure of the ABL over a tiny island of Kaashidhoo in the Indian Ocean and mesoscale modelling simulation studies in one-dimensional vertical column mode over Kaashidhoo and comparison of simulations with the observations

A modification is suggested in the bulk aerodynamic algorithm for the estimation of air-sea interface fluxes. Spatio-temporal variation of air-sea interface flux estimates are discussed in relation to prevailing meteorological conditions.
Emphasis has been laid on understanding the role of wind speed in the estimates of air-sea interaction parameters and also to see how it affects the behaviour of these parameters with varying speeds. Meteorological data obtained from balloon borne GLASS Sonde are being utilized in characterizing the vertical structure of the MABL in the vicinity of ITCZ. With a view to testing the mesoscale model performance over a tiny island of Kaashidhoo, two mesoscale models were used to simulate the profiles of winds, temperature and humidity. Results revealed from the analysis of GLASS Sonde data obtained over the Indian Ocean and Kaashidhoo island explain the thermodynamic structure of the ABL in varying convective regimes over ocean vis-à-vis over an island. This study is the first of its kind to be reported over a large spatial domain of data-sparse region of tropical Indian Ocean. It is a widely accepted fact that the western tropical Indian Ocean is one of the least explored regions, as far as the MABL studies are concerned. Therefore the results obtained from the present study are quite relevant for further studies over this region.

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