1.1 BACKGROUND

The Atmospheric Boundary Layer (ABL) is the lowest part of the atmosphere that is formed due to interaction between atmosphere and land. Studies relating to the ABL are important as most of the human and biological activities are taking place in this layer. Most of the atmospheric pollution is generated at the surface due to natural and anthropogenic activities. Exchange of heat and moisture with the earth surface are taking place in ABL.

The state of the troposphere, the lowest atmospheric layer, is determined by surface forcing. Weather systems are driven by heating and moisture release from the surface and act throughout the depth of the troposphere. However, only a relatively thin layer of this lowest part of the atmosphere is regularly subject to turbulent mixing, the subject addressed here using micrometeorological methods. This layer is known as the atmospheric boundary layer. The defining features of the ABL are that it responds to surface forcing on a time scale of around an hour, and it is turbulent throughout its depth due to interactions with the surface. The layer has practical and scientific importance. Mass, energy and momentum transfer takes place between the ABL and the free atmosphere. The lowest subdivision of the ABL (10%) is known as the surface layer.

The earth’s atmosphere is semitransparent to the incoming solar radiation. The rest of the energy passes through the atmosphere and is absorbed by the surface of the earth. Later this energy is transferred back, primarily to the lowest kilometer of the atmosphere. The flow in the ABL has turbulent character. During the daytime energy gained by the surface is transferred to the atmosphere, to the soil, and is also used in the evaporation processes. This transfer of the heat from the ground surface to the air directly above can generate convection. Convection redistributes heat throughout the ABL.

The atmospheric surface layer is the active link between atmosphere and the surface of the earth. The characteristic of the boundary layer is depending on the surface layer fluxes. Several formulations (Sorbjon, 1986) exit establishing
relationship between the surface layer fluxes and the parameter characterizing the fluxes of the different levels in the boundary layer.

The structure and other characteristics of the boundary layer are strongly influenced by the diurnal cycle of the solar heating. After sunrise on a clear day, the surface immediately becomes warmer than the air above. The sensible heat and evaporating moisture are transferred from ground to air above by convective mixing (larger eddies) and turbulence (smaller eddies) as result the boundary layer in its upper part becomes a mixed layer with uniform temperature, wind direction, wind speed and humidity with height. In the late afternoon before sunset, infrared radiation emission and cooling starts. The ground and air temperatures become equal. The upward flux of sensible heat vanishes. A short lived state of neutral or nearly neutral stability prevails till sunset. As the sun sets, solar radiation absorption at the earth surface stops and energy supply to buoyant convective motions cut off. A brief description of the structure of the atmosphere is given below.

1.2 STRUCTURE OF THE ATMOSPHERE

Earth's atmosphere can be divided into five main layers. These layers are mainly determined by whether temperature increases or decrease with height. These layers are demonstrated in Figure 1.1 as a variation of air temperature with height and are described as under:

1.2.1 Troposphere

The troposphere begins at the surface and extends to between 7 km at the poles and 17 km at the equator. The troposphere is mostly heated by transfer of energy from the surface. Therefore, on average the lowest part of the troposphere is warmest and temperature decreases with height. Due to higher temperature at surface, it promotes vertical mixing. The troposphere contains roughly 80% of the mass of the atmosphere. The tropopause is the boundary between the troposphere and the next layer, stratosphere.

1.2.2 Stratosphere

The stratosphere extends from the tropopause to about 51 km. Temperature increases with height, which restricts turbulence and mixing. The stratopause, which is the boundary between the stratosphere and mesosphere, typically is at 50 to 55 km.
1.2.3 Mesosphere

The mesosphere extends from the stratopause to 80–85 km. It is the layer where most meteors burn up upon entering the atmosphere. Temperature decreases with height in the mesosphere. The mesopause, the temperature minimum that marks the top of the mesosphere, is the coldest place on the Earth and has an average temperature around −100 °C.

1.2.4 Thermosphere

Temperature increases with height in the thermosphere from the mesopause up to the thermopause and then it is constant with height. The temperature of this layer can rise to 1500 °C, though the gas molecules are so far apart that temperature in the usual sense is not well defined. The top of the thermosphere is the bottom of the exosphere called the exobase. Its height varies with solar activity and ranges from about 350–800 km.
1.2.5 Exosphere

The outermost layer of Earth's atmosphere extends from the exobase upward. Here the particles are so far apart that they can travel hundreds of kilometers without colliding with one another. Since the particles rarely collide, the atmosphere no longer behaves like a fluid. These free-moving particles follow ballistic trajectories and may migrate into and out of the magnetosphere or the solar wind. The exosphere is mainly composed of hydrogen and helium.

1.3 ATMOSPHERIC BOUNDARY LAYER

The atmospheric boundary layer (ABL) of the atmosphere is the layer from surface of the earth to a few hundred meters above. This is also called as Planetary Boundary Layer (PBL) or simply boundary layer (BL). The vertical thickness or depth of ABL or PBL changes with time of the day in response to the diurnal heating/cooling cycles. Following sunrise on a clear day, heating of land surface takes place and consequent vertical mixing leads to increase in the thickness of the PBL, reaching maximum in the afternoon. Throughout the night radiative cooling of the surface inhibits thermal mixing which substantially reduces the thickness of the PBL. The thickness of the PBL generally varies from a few tens of meters to two or three kilometers depending on roughness of the ground, topography, nature of the vegetation coverage, intensity of wind, rate of heating/cooling of the surface, advection of heat and moisture, and vertical motion etc.

Over land surfaces in high pressure regions the boundary layer (BL) has a well defined structure that evolves with the diurnal cycle. The three major components of this structure are:

- The mixed layer (ML)
- The residual layer (RL) and
- The stable BL (SBL)

When clouds are present in the ML, it is further subdivided into a cloud layer and a sub-cloud layer. The surface layer is the region at bottom of the BL where turbulent fluxes and stress vary by less than 10% of their magnitude; whether it is a part of a ML or SBL. Finally, a thin layer called a micro layer or interfacial layer exists in the lowest few centimeters of air, where molecular transport dominates over turbulent transport.
Over oceans, the BL depth varies relatively slowly in space and time. Since water has large heat capacity, SST (Sea Surface Temperature) varies slowly and hence forcing into the bottom of the BL also varies slowly. Most of the changes in the BL depth over oceans are caused by synoptic and mesoscale processes of vertical motion and advection of different air masses over the sea surface.

1.3.1 Significance of Boundary Layer

1. People spend most of their lives in the PBL.
2. Daily weather forecasts of dew, frost, maximum and minimum temperatures are really boundary layer forecasts.
3. Pollution is trapped in the boundary layer.
4. Fog occurs within the boundary layer.
5. The primary energy source for the whole atmosphere is solar radiation, which for the most part is absorbed at the ground and transmitted to the rest of the atmosphere by boundary layer processes.
6. Virtually all water vapour that reaches the free atmosphere is first transported through the boundary layer by turbulent and advective processes.
7. Turbulent transport of momentum down through the boundary layer to the surface is the most important momentum sink for the atmosphere.
8. About 50% of the atmosphere’s kinetic energy is dissipated in the boundary layer.
9. Wind stress on the sea surface is the primary energy source for ocean currents.
10. Turbulence transport and advection in the boundary layer move water and oxygen to and from immobile life forms like plants.

Thus the boundary layer processes affect our lives directly and indirectly via its influence on the rest of the weather. Hence it is essential to probe the PBL by observations and numerical modelling. The ABL consists of the surface layer and the transition layer (or Ekman) layer.

1.3.2 The Surface Layer

The surface boundary layer (SL) is the lowest part of the PBL. Under conditions of horizontal homogeneity (e.g., over open oceans or broad, flat prairies) and quasi-steady state, the following approximations are usually applied to SL (Panofsky and Dutton, 1984):
1. The rotation of the earth, that is, the Coriolis effect is probably unimportant in the SL.

2. The SL occupies the lowest 10% of the PBL.

3. Experiments have shown that the vertical variation in stress and heat flux in the SL is within 10% and thus SL is also named as the “constant flux layer”.

4. Wind direction does not change appreciably with height in the SL.

5. The variation of mean variables with height Z, is controlled primarily by three parameters: the surface stress, the vertical heat flux at the surface and the terrain roughness.

6. Transport of atmospheric properties by turbulent diffusion (eddies) is much more important than transport by molecular diffusion.

### 1.3.3 The Mixed Layer

The turbulence in the mixed layer (ML) is usually convectively driven, although nearly well ML can form in regions of strong winds. Convective sources include heat transfer from a warm ground surface, and radiative cooling from the top of the cloud layer. Even when convection is the dominant mechanism, there is usually wind shear across the top of the ML that contributes to the turbulence generation. Starting about a half an hour after sunrise, a turbulent ML begins to grow in depth, and reaches its maximum depth in late afternoon. It grows by entraining the less turbulent air from above. The resulting turbulence tends to mix heat, moisture, and momentum uniformly in the vertical. When clouds are present in the mixed layer, it is further subdivided into a cloud layer and a sub-cloud layer.

### 1.3.4 The Residual Layer

About a half hour before sunset the thermals cease to form, allowing turbulence to decay in the formerly well mixed layer. The resulting layer of air is sometimes called the residual layer (RL), because its initial mean state variables and concentration variables are the same as those of recently decayed ML. The RL is neutrally stratified, resulting in turbulence that is nearly of equal intensity in all directions. The RL often exists for a while in the mornings before being entrained into the new ML. Variables such as virtual potential temperature usually decrease slowly during the night, because of radiation divergence. This cooling rate is of the order of 1°C/day. The cooling rate is more-or-less uniform throughout the depth of
the RL. When the next day’s ML top reaches the base of the RL, the ML growth becomes very rapid.

1.3.5 The Stable Boundary Layer (SBL)

Sometimes the boundary layer can become stably stratified whenever the surface is cooler than that of the air. This SBL often forms over land during night when it is called nocturnal boundary layer. It can also form by advection of warmer air over cooler surface. This is characterized by statically stable air with weaker, sporadic turbulence. Although the wind at ground level frequently becomes lighter or calm at night, the wind aloft may accelerate to super geo-strophic speeds in a phenomenon that is called the low level jet or nocturnal jet. The statically stable air tends to suppress turbulence, while the developing nocturnal jet enhances wind shears that tend to generate turbulence. As a result, turbulence sometimes occurs in relatively short bursts that can cause mixing throughout the SBL.

1.3.6 Coastal Atmospheric Boundary Layer

The coastal ABL is characterized by variations in topography, large temperature gradients and change in surface roughness. The sea near the coast is characterized by large variations in sea surface temperature (SST) and roughness and a non equilibrium sea state. These conditions produce interactions between sea-breeze, mesoscale eddies and terrain-generated winds that cause complex flow patterns (Wilczak et al., 1991; Douglas and Kessler, 1991; Ulrickson, 1992). The flow structure of the coastal ABL depends on the curvature of the shoreline and the topography of the coast like the presence of mountains along the coast (Klipp and Mahrt, 2003).

Coastal atmospheric characteristics are found to be distinctly different from that over an inland with particular reference to the dynamics of local atmospheric circulation that decides the dispersion of air pollutants. The land-sea breeze in coastal region greatly influences the trajectory while the formation of the thermal internal boundary layer (TIBL) affects the turbulent diffusion of the pollutants. One of the important measurements required to characterize the coastal ABL is the vertical profile of wind and temperature at a few specified time intervals representative of the relevant climatology. Sea breeze is a local convective circulation developing as a direct effect of the differential heating between land and sea. The strength, horizontal and vertical extent of the sea breeze depends upon the prevailing synoptic flow, the
land-sea surface temperature difference, and boundary layer temperature and moisture profile.

1.4 EVOLUTION OF ABL

The solar heating causes thermal plumes to rise, transporting moisture, heat and aerosols. The plumes rise and expand adiabatically until a thermodynamic equilibrium is reached at the top of the ABL. The moisture transferred by the thermal plumes forms convective clouds. Figure 1.2 illustrates a typical daytime evolution of the atmospheric boundary layer in high pressure conditions over land.

![Figure 1.2. Schematic fair-weather atmospheric boundary layer structure over land.](image)

The drier air from the free atmosphere penetrates down, replacing rising air parcels. The part of the troposphere between the highest thermal plume tops and deepest parts of the sinking free air is called the entrainment zone. The convective air motions generate intense turbulent mixing. This tends to generate a mixed layer, which has potential temperature and humidity nearly constant with height. When buoyant turbulence generation dominates the mixed layer, it is called a convective boundary layer (CBL). The boundary layer from sunset to sunrise is called the nocturnal boundary layer (NBL). It is often characterized by a stable layer, which forms when the solar heating ends and the radiative cooling and surface friction stabilize the lowest part of the ABL. Above that, the remnants of the daytime CBL form a residual layer.
The ABL is bounded at one end by land or water. This solid or liquid boundary offers resistance to the motion of the atmosphere in its neighborhood. There is no such resistance to the motion at the other end of ABL which constitutes the lower boundary of the free atmosphere. The resistance offered by the underlying surface brings the motion of the air to a stop at the land surface and reduces it by nearly two orders of magnitude at the sea surface. The considerable vertical shear in the air motion, thus created at the underlying surface generates random motion called turbulence, which plays key role in affecting the structure of the ABL. This turbulence is accentuated by favourable and damped by unfavourable temperature gradients. It regulates the rates of transfer of heat, moisture and particulate matter in the ABL.

The ABL contains several component layers distinguishable by their physical properties. Closest to the surface, usually extending no higher than a few centimeters in depth is the interfacial or viscous layer. It is only in this thin envelope of air most directly in contact with the surface that transport of physical fields via turbulence is less efficient than by molecular transport (Stull, 1988). Above the interfacial layer, the lowest 10% to 15% (by height) of the ABL is called the surface, or constant flux layer. In this layer, fluxes of momentum, sensible and latent heat are approximately constant in the vertical, and variations in the wind field are most directly attributable to static stability and local frictional effects. The influence of the Earth’s rotation is insignificant and often neglected in the surface layer (Kaimal and Finnigan, 1994).

The outer, or Ekman layer comprises the bulk (about 90% by depth) of the ABL. Flow in the outer layer usually exhibits less variability in both direction and speed than in the surface layer, hence the Coriolis effect becomes a necessary consideration (Stull, 1988). Moreover, in the outer layer, fluxes of energy and momentum may vary significantly with height. If sufficient lift and water vapor is Present, cumuliform clouds may develop in this layer, which alter the thermodynamic response of the entire ABL (Wang and Liang, 2009). The entrainment zone divides the ABL from the free atmosphere. With few exceptions, this layer defines the upper boundary for turbulence generated in the ABL. The high static stability of this layer often results in a thermal inversion, characterized by an increase in temperature with height.
1.5 STRUCTURE OF THE ATMOSPHERIC BOUNDARY LAYER

When the boundary layer is observed above land with regular cycle of sunlight and darkness, it can be divided into four major parts. During daytime the ground is heated by the sun and a convectively driven, vertically mixed layer grows. It reaches its maximum during the afternoon. In the mixed layer temperature decreases with height, on the bottom unstable stratification, higher up neutral stratification is observed. On top of this layer usually a stable layer acts as a lid for the rising thermals, restraining the domain of turbulence. While heating ceases when the sun goes down, turbulence decays in the formerly mixed layer and the so called residual layer is formed. Figure 1.3 shows that the residual layer is not in contact with ground therefore night processes form the stable nocturnal boundary layer. The underlying surface is colder than the air and provokes an increasing temperature with height. The so called nocturnal or low-level jets develop wind shears that tend to generate shortly bursting turbulences. In contrast to the sharp lid which defines the top of the mixed- and residual layer, the nocturnal boundary layer smoothly blends into the residual layer.

Figure 1.3. The boundary layer high pressure region over land this consists three major parts: A very turbulent mixing layer; a less turbulent residual layer containing former mixed layer air; A nocturnal stable boundary layer of sporadic turbulence (Stull, 1988).
layer. The surface layer is defined as 10% of the bottom of the boundary layer and has to be considered as an independent part, regardless of whether it is part of the mixed or stable boundary layer. Furthermore, the roughness of the surface has to be taken into consideration by talking about the height of the surface layer and its conditions.

Further, the ABL is divided into three regimes due to the mixing mechanism. They are:

1. Skin layer: This layer is a small layer close to the ground having thickness of few centimeters. In this layer, molecular diffusion is taking place to transfer the energy from one air parcel to another.

2. Surface Layer: The surface layer is the lowest part of the boundary layer which is in direct contact with the earth’s surface. Surface layer is typically 50-100 m thick or 10% of the depth of the ABL. In this layer, the energy fluxes (especially the momentum flux) are nearly constant. The surface layer is part of the daytime mixed layer and night time stale layer.

3. The Ekman Layer: This layer is also called as ABL. In this layer, the eddy diffusion is taking place. (Eddy is the concept (like molecule)), but the size of eddies are different and varying with respect to atmospheric stability.

1.6 THE THERMODYNAMIC STRUCTURE OF THE ABL

All the water vapour entering the atmosphere by evaporation enters it through the PBL. Sensible heat can and does enter and leave the atmosphere through the PBL. The fluxes of water vapour and heat, coming into the PBL in association with the meso and synoptic scale motions of the atmosphere, result in establishing a characteristic structure of the PBL. In the tropics, the stratification of the PBL into five distinct layers, viz., (a) super adiabatic layer, (b) well mixed layer, (c) transition layer, (d) cloud layer and (e) trade wind inversion plays a special role in the massive convective transports of heat and moisture (Augstein et al., 1974; Garstang and Betts, 1974).

1.7 ATMOSPHERIC STABILITY

Atmospheric stability is the state of air-mass which can rise above or sink to ground or remain well mixed in the ABL. Generally, it is stable in the night hours due to cut-off of solar radiation and it is unstable/well mixed in the noon hours of the
day due to intense solar heating of the earth surface. The atmospheric temperature and winds are the vital factors to determine the state of the air parcel in the ABL or stability of the ABL.

1.8 MEAN CHARACTERISTICS OF ABL

The stability of the ABL can be assessed with the use of altitude profiles of air temperature, humidity and wind as shown in Figure 1.4. On the basis of these basic parameters, another parameter such as surface heat flux can also be used to assess the stability of the ABL. The mixed layer (ML) is so named because intense vertical mixing tends to leave conserved variables such as potential temperature and humidity nearly constant with height (as shown in Figure 1.4). Even wind speed and direction are nearly constant in the mixed layer.

Mixing can be generated mechanically by shears, or convectively by buoyancy. Shears near the ground are usually more important for generating mixing than shears across the top of MLs, for atmospheric situations. Shears at the ML top, however, can cause a separate layer to form. A mixed layer dominated by buoyant turbulence generation is called a convective boundary layer (CBL) or convective mixed layer.

Figure 1.4. Mean Characteristics of the ABL with the shape of profiles (After Driedonks and Tennekes, 1984)
1.8.1 Stable Boundary Layer (SBL) Characteristics

As the night progresses, the bottom portion of the residual layer is transformed by its contact with the ground into a stable boundary layer. This is characterized by statically stable air with weaker, sporadic turbulence. Although the wind at ground level frequently becomes lighter or calm at night, the winds aloft may accelerate to super-geostrophic speeds in a phenomenon that is called the low-level jet or nocturnal jet. As opposed to the daytime ML which has a clearly defined top, the SBL has a poorly defined top that smoothly blends into RL above (as shown in Figure 1.5). The top of the ML is defined as the base of the stable layer, while 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. This SBL which often forms at night over land is also known as nocturnal boundary layer (NBL).

Figure 1.5 shows typical profiles of mean variables in the SBL for the case of weak turbulent mixing. The greatest static stability is near the ground, with stability decreasing smoothly toward neutral with height. If stability is great enough near the surface to cause temperatures to increase with height, then that portion of the SBL is classified as a temperature inversion. In fact, sometimes the whole SBL is loosely called a nocturnal inversion.

![Figure 1.5](image.png)

**Figure 1.5.** Typical Stable boundary layer of mean (a) absolute temperature (b) potential temperature (c) wind speed and (d) specific humidity. (After Stull, 1991)
1.8.2 Diurnal Cycle of ABL

During a clear day the boundary layer can be divided into several sub layers as shown in Figure 1.6.

1. The roughness sub layer: This is the layer of air in which air flows around individual roughness elements (such as grass, plants, trees or buildings).

2. The surface layer (known as the constant flux layer): In this layer, typically 100 m thick (or 10% of the depth of the ABL), the winds, temperature and humidity vary rapidly with altitude, and the characteristics of turbulence are affected by the surface, vertical fluxes of heat and momentum.

Figure 1.6. Typical profiles of potential temperature, wind and humidity over land in mid-latitudes during cloudless conditions (left panels) and Schematics of the typical ABL circulation and eddy structure of the ABL (right panels) in the day and night (from Kaimal and Finnigan, 1994).
3. The well-mixed layer rising buoyant plumes from the surface layer, and associated turbulence, cause potential temperature and other quantities to be relatively constant with altitude. The earth’s rotation becomes important in this layer, and the wind direction with height.

4. The capping inversion on a day the convective boundary layer is often capped by a temperature inversion, which inhibits mixing, and confines air and pollution below it to within the boundary layer. At night a new stable nocturnal boundary layer grows as air is cooled from the surface. The daytime mixed-layer remains as a residual layer while the capping inversion is eroded. Sometimes the ABL is difficult to define; in the vicinity of fronts there is no obvious capping inversion and the ABL structure is more a response to synoptic forcing.

1.9 BOUNDARY LAYER HEIGHT AND STRUCTURE

The descriptions of the ABL provided by Sutton (1953) separated the layer into two regions: (1) the first 50–100 m, a region of constant shearing stress, insensitive to the Earth’s rotation, with wind structure determined by the surface friction and temperature gradient, (2) above that layer, extending to about 1000 m, a region where shearing stress is variable and wind structure is influenced by the surface friction, density gradient, and the earth’s rotation. The theoretical geostrophic balance is achieved at the top of that layer. An exact definition of the top of the ABL had been elusive. Viewed as the height this is:

\[ z_h = \pi \left( \frac{2 K_m}{f} \right)^{1/2} \]  

(1.1)

Where \( K_m \) is the eddy coefficient for momentum and \( f \) is the Coriolis parameter.

Another boundary layer height, derived from considerations of Rossby number similarity (Tennekes, 1982) is,

\[ z_h = c \left( \frac{u_*}{f} \right) \]  

(1.2)

where \( c \) is an unknown constant (0.2 – 0.6) and \( u_* \) is the friction velocity. The friction velocity is defined as \( (\tau / \rho)^{1/2} \) where \( \tau \) is the surface stress or vertical momentum flux and \( \rho \) is the density of air. This derivation assumes a neutral boundary layer with no heat flux, so its extension to the non-neutral ABL is complicated. A value of \( c = 0.25 \) yields ABL heights close to the observed day time heights.
Observational studies in the laboratory and in the atmosphere show that the height of the lowest inversion base $z_i$ provides the most consistent scaling length for day time boundary layer turbulence:

$$z_h = z_i$$

This approach assumes that a capping inversion exists, acting as a lid and damping out vertical motions. In the middle latitudes, such an inversion is usually present during the day. The height $z_i$ has to be measured directly with slow ascent balloons or with remote sensors like Sodars and Radars. In the tropics, the diminishing Coriolis force and high moisture levels present a different upper boundary for the ABL. The cloud base, which often appears below the trade-wind inversions, is separated from the boundary layer below by a transition layer 10-100 m thick, where the mean profiles of virtual potential temperature and mixing ratio shift from their mixed layer values to their cloud layer values.

Under high pressure, divergence of air masses results in shallow boundary layer height. Typically, only fair-weather cumulus clouds are present. The analysis of the boundary layer structure is not usually that straightforward in the case of low pressure. The air parcels converge in low pressure in connection with updrafts, which transfer boundary layer air parcels high above the ground. The clouds may then grow to the top of the troposphere. This leads to extensive variations in the local boundary layer top. Thus, it becomes difficult to define a larger scale boundary layer depth.

### 1.10 SPECTRAL AND CO-SPECTRAL CHARACTERISTICS OF TURBULENCE IN THE ABL

#### 1.10.1 Spectra and Co-spectra

The one-dimensional power spectrum of boundary layer turbulence exhibits three major spectral regions, namely, (i) the energy-containing range which has the bulk of the turbulent energy produced by shear and buoyancy, (ii) the inertial subrange where energy is neither produced nor dissipated but handed down to smaller and smaller scales and (iii) the dissipation range, where the kinetic energy is converted to internal energy. The frequency dependence of power spectral densities in these three regions is roughly given by $n^0$, $n^{5/3}$, and $n^{-7}$, where $n$ is the cyclic frequency (Hz). The power spectra in general are dependent on height, surface roughness, mean velocity, and thermal stability.
In addition to the above general regions of the power spectra, a “buoyant subrange” has been suggested to exist (Teunissen, 1970) for atmospheric turbulence in stable condition. In this region, a “buoyant length” depending on the stratification is a factor limiting the inertial subrange, rather than the distance to the ground. Busch and Panofsky (1968) suggested the existence of this range, for the vertical spectra with frequency dependence of the order of $n^{-3}$ as suggested by Lumley and Panofsky (1964). Before going into the analysis of the atmospheric spectra, a brief outline of the theoretical development of the spectrum of turbulence is given below.

### 1.10.2 Spectra of Turbulence

Analogous to the dispersion of white light into colors when passed through a prism (spectrum of light), a turbulent atmospheric signal can be analyzed either by using mathematical techniques or measured by using physical devices like spectrum analysers, in order to determine the contribution of eddies of various sizes to the total turbulent kinetic energy.

If we consider the variations of the turbulent stream-wise velocity component of the wind $u(t)$ at a fixed point in time, its variations can be represented by the Fourier integral (Taylor, 1938):

$$U'(t) = 2\pi \int_0^\infty [I_1(n) \cos 2\pi nt + I_2(n) \sin 2\pi nt] \, dn$$

(1.4)

Where $n$ is the cyclic frequency (Hz), $I_1(n)$ and $I_2(n)$ are the Fourier coefficients. Using the theorem in harmonic analysis, the total variance $\overline{u'^2}$ may be expressed as:

$$\overline{u'^2} = 2\pi^2 \int_0^\infty \lim_{T \to \infty} [\frac{(I_1^2(n) + I_2^2(n))}{T}] \, dn$$

(1.5)

where $T$ is the time over which the turbulent variations exit. The quantity

$$2\pi^2 \int_0^\infty \lim_{T \to \infty} [\frac{(I_1^2(n) + I_2^2(n))}{T}] \, dn$$

(1.6)

represents the contribution to total variance by frequencies between $n$ and $n+dn$, and is denoted by

$$\overline{u'^2} \int F(n) \, dn.$$  

Note that $\int_0^\infty F(n) \, dn = 1$

A plot of $F(n)$ versus $N$ denotes the energy or power spectrum.
1.10.3 Correlation and Spectrum Properties

An idea about the size of eddies would be more realistically formulated in terms of the differences in velocity existing instantaneously between one point and another in the fluid. For example, small eddies would impose differences in velocity between two relatively close points, whereas large eddies would often give velocities that would be similar at the two points. The statistical expression of this idea is provided by the space-correlation coefficient $R(x)$ defined as

$$R(x) = \frac{\overline{u'_1 u'_2}}{u'^2} \tag{1.7}$$

Where $u'_1$ and $u'_2$ are instantaneously measured turbulent velocity components at two points, one is fixed and other varied in distance $x$ from it.

In the above expression, turbulence is assumed to be homogeneous which means that the statistical properties of $R(x)$ and $u'^2$ are independent of position. The sharpness of the decrease of $R(x)$ with $x$ is a reflection of the eddy sizes, and can be represented by a length $l$ defined as:

$$1 = \int_0^\infty R(x) \, dx \tag{1.8}$$

provided that the integral converges. Taylor (1938) suggested this quantity as a possible definition of the average size of eddies, now commonly referred to as the scale of the turbulence.

A time correlation coefficient is similarly referred to as the autocorrelation coefficient $R(t)$ and defined in terms of eddy velocities at a fixed point at instants of time separated by $t$. Invoking Taylor’s hypothesis, we write

$$R(t) = R(x) \text{ when } x = ut$$

Adopting this relation, Taylor introduced into turbulence theory, the relation between the correlation coefficient and the spectrum of turbulence. Using an analytical theorem due to Norbert Wiener, Taylor showed that $R(t)$ and $F(n)$ are related by the expression

$$R(t) = \int_0^\infty F(n) \cos 2\pi nt \, dn \tag{1.9}$$

And by comparison with the Fourier integral theorem, it was shown that when equation (1.9) holds, then
\[ F(n) = 4 \int_0^\infty R(t) \cos 2\pi nt \, dt \]  

(1.10)

must be true. Thus knowing \( R(t) \), \( F(n) \) can be calculated and vice versa.

The immediate practical importance of this development is two-fold. First, it brings out the idea of a continuous range of eddy sizes (in so far can be identified with the inverse frequency). It provides a way of identifying those sizes which are of most significance to kinetic energy. Second, it provides a method of calculating spectral distribution.

1.10.4 Isotropic Turbulence and Universal Equilibrium Theory

In the early work on the statistical theory of turbulence, Taylor (1938) introduced, in the representation and analysis of turbulence, the further simplification of isotropy: the condition in which the statistical properties such as \( \overline{u'^2} \) are unaffected by rotation or reflection of the axes of reference. The quantities \( \overline{u'^2}, \overline{v'^2} \) and \( \overline{w'^2} \) then become equal, as well as uniform in space.

The condition of isotropy does not occur as a general rule in the atmosphere, but its existence has often been assumed to an approximate degree. Kolmogorov (1941) put forward the idea of local isotropy i.e. isotropy which is confined to the small scale structure of the motion. The theory variously referred to as the ‘similarity theory’ or theory of universal equilibrium postulates that all turbulent motions, irrespective of their origin and of the form of the mean flow, and subject only to their occurrence with sufficiently large scale, posses local isotropy. The idea introduced by Kolmogorov is that in the process of transfer of energy from large to small scale, the successively concerned eddies gradually loose all the original influence of the energy containing eddies, and ultimately reach a stage when the motion of the succeeding eddies is isotropic. For the small eddies the properties are conditioned firstly by an inflow of energy which, because of the small viscous dissipation in the relatively larger eddies, is equal to the energy transferred from the mean flow. The energy is ultimately dissipated firstly by viscous action at a dissipation \( \epsilon \) per unit mass of the fluid, and secondly the viscosity \( \nu \) of the fluid governing this rate of dissipation. Thus for this equilibrium range, Kolmogorov’s first similarity hypothesis states that “the average properties of the small-scale components of turbulence are determined uniquely by the quantities \( \epsilon \) and \( \nu \)”. When this equilibrium range is sufficiently wide, the larger eddies in the range will contribute so little to the total viscous dissipation that a subrange will exist in which the average properties will be determined purely by
the inertial transfer of energy. This concept lead to Kolmogorov’s second hypothesis: “At the large eddy end of the equilibrium range there is an inertial subrange in which the average properties are determined by the quantity $\epsilon$”.

The energy containing and the dissipation ranges have their own characteristics lengths $\Lambda$ and $\eta$ respectively. The length scale $\Lambda$ is the integral length scale of fluid mechanics derived from equation (1) through Taylor’s hypothesis:

$$\Lambda = \frac{\overline{u^2}}{\epsilon} = \bar{u} \int_0^\infty R(t) \, dt$$

where $\tau$ is the integral time scale, $t$, the lag time and $R(t)$, the autocorrelation function. The length scale $\eta$ in the dissipation range is the Kolmogorov’s microscale, expressed as:

$$\eta = (\nu^3 / \epsilon)^{1/4}$$

In the inertial subrange the transfer of energy from $k = \Lambda^{-1}$ to $k = \eta^{-1}$ is controlled entirely by $\epsilon$, the rate at which the energy leaves $k = \eta$ ($k$ is the wave number vector). Typically $\Lambda$ varies from 10 to 500 m, and $\eta$ is of the order of 0.001 m in the Convective Boundary Layer (CBL).

In most atmospheric measurements only fluctuations along the longitudinal (streamwise) component of the wind are measured. The corresponding wave number for fixed tower measurements is approximated as $k_1 = 2\pi/\lambda$, where $\lambda = \bar{u}/n$ is the wave length, $\bar{u}$ is the mean wind speed and $n$ is the cyclic frequency.

### 1.10.5 Spectrum of Atmospheric Turbulence

In homogenous turbulence the scalar energy spectrum $E(k)$ represents the contribution to the total kinetic energy from Fourier modes of wavenumber between $k$ and $k+dk$ (Lumley and Panofsky, 1964). In the inertial subrange

$$E(k) = \alpha \epsilon^{2/3} k^{-5/3}$$

The $u$-spectrum, in its familiar one dimension form, becomes

$$F_u(k_1) = \alpha_1 \epsilon^{2/3} k_1^{-5/3}$$

where $\alpha_1 = 0.5$ to 0.6 is the Kolmogorov constant.

Assuming local isotropy in the inertial subrange we have the following relationship between the $u$, $v$ and $w$ spectra.
\[ F_v(k_1) = F_w(k_1) = \frac{4}{3} F_u(k_1) \quad (1.15) \]

where \( v \) is the crosswind (lateral) and \( w \) is the vertical component of wind. Another consequence of local isotropy is that all correlation between velocity components and scalars vanish implying that there can be no turbulent fluxes in the inertial subrange.

For the temperature spectrum (\( \theta \)), Corrsin (1951) proposed an inertial subrange form:

\[ F_\theta(k_1) = \beta_1 \epsilon^{-1/3} N k_1^{-5/3} \quad (1.16) \]

where \( N \) is the dissipation rate for half the temperature variance and \( \beta_1 \) is a universal constant with a value around 0.8 (Kaimal et al., 1972). This spectral form appears to be valid for other scalars like humidity. The spectral forms in the energy containing range tend to be different for different variables since the integral scales they respond to are different. A spectral gap is assumed to exist in the range which separates boundary layer turbulence from external fluctuations. A spectral gap can often be found in the CBL at frequency \( 10^{-3} \) to \( 10^{-4} \) Hz, between the convection-driven boundary layer scales and synoptic scales (Van der Hoven, 1957). This end of the spectrum is susceptible to contamination from long term trends, diurnal variations, synoptically induced changes, and drifts in the sensor measuring the fluctuation. In the absence of trend, \( F(k_1) \) levels off to a constant value when \( k_1 \) tends to zero. This is a consequence of the one-dimensional formulation of the three-dimensional turbulence spectrum (Kaimal and Finnigan, 1994). To identify the energy peaks and valleys for a more realistic representation in the distribution of turbulent energy, meteorologists use the frequency weighted form \( k_1 F(k_1) \) of the spectrum. The frequency weighted spectrum is also referred to as a logarithmic spectrum since it represents the variance \( \sigma^2 \) per unit logarithmic interval. Its unit is variance instead of variance per wave number interval \( \Delta k_1 \) in the unweighted spectrum. A plot of \( k_1 F(k_1) \) vs \( k_1 \) on log-log scale yields power relationships (e.g. \( k_1^{-2/3} \)) as straight lines. The frequency-weighted spectrum is defined as:

\[ \sigma^2 = \int_0^\infty F(k_1) \, dk_1 = \int_0^\infty k_1 F(k_1) \, d(\ln k_1) \quad (1.17) \]

Although spectral theories are formulated in wavenumber \( (k_1) \) space, measurements, in practice, are performed in frequency space. Assuming Taylor’s hypothesis, the spatial scales can be converted to frequency scales from the relation:
\[ k_1 = 2\pi n/u \]  

(1.18)

The \( u \) spectrum, for example, is given by

\[ \int_0^\infty F_u(k_1) \, dk_1 = \sigma_u^2 = \int_0^\infty S_u(n) \, dn. \]  

(1.19)

Substituting for \( k_1 \) from equation (1.18) into equation (1.19) and differentiating with respect to \( n \) yields

\[ F_u(k_1) \, dk_1 = n \, S_u(n) \]  

(1.20)

In view of the relationship (1.20), \( n \, S_u(n) \) can be plotted as a function of \( k_1 \) or \( n \) without any conversion of units.

**1.10.6 The Cross Spectra**

The cross-spectrum relates the spectra of two variables, namely \( w' \) and \( \theta' \). We get more information about the spectrum of \( w'\theta' \), such as how the phase of the \( w' \) fluctuations relate to the phase of \( \theta' \) fluctuations as a function of frequency (Stull, 1988). The cross spectrum between \( w' \) and \( \theta' \) is defined as

\[ G_{w\theta} = F_{w^*} F_{\theta} \]  

(1.21)

where \( F_w = F_{wr} + i F_{wi} \) and \( F_{w^*} \) is the complex conjugate of \( F_w \); the subscripts \( r \) and \( i \) denote real and imaginary parts. \( |F_w(n)|^2 \) is defined as the unfolded spectral energy for the variable \( w \). Similarly \( |F_\theta(n)|^2 \) for \( \theta \).

Expanding equation (1.21) and collecting the real and the imaginary parts, the real part is defined as the cospectrum, \( C_O \):

\[ C_O = F_{wr} F_{\theta r} + F_{wi} F_{\theta i} \]

and the imaginary part is defined as the quadrature spectrum, \( Q \):

\[ Q = F_{wi} F_{\theta r} - F_{wr} F_{\theta i} \]

The cospectrum equals the covariance \( \frac{\partial}{\partial n} \) and is frequently used in meteorology. The quadrature spectrum is not used directly. It is equal to the spectrum of the product of \( \theta' \) times a phase shifted \( w' \), where \( w' \) is phase shifted by a quarter period of \( n \). From the quadrature and cospectra, an amplitude spectrum is defined as
\[ A_m = Q^2 + Co^2 \]

A large amplitude at any frequency \( n \) implies that \( w' \) and \( \theta' \) are strongly correlated at that frequency. If the amplitude of the spectrum is small for any frequency \( n \) then the coherence spectrum will not be significant (i.e. unreliable) for that frequency. The coherence spectrum acts very much like a frequency dependent correlation coefficient.

1.10.7 The Spectral Coefficient

When the power spectral density \( nS_u(n) \) is plotted against frequency on a log-log graph we find the inertial subrange characteristics obey a “power law” of the form \( n^i \) where the exponent ‘i’ is called the spectral coefficient (Cambel, 1993). In such plots the one dimensional turbulence spectrum normally depicts \( i = -2/3 \) characteristics in the inertial subrange. The inertial subrange region, as stated above, appears to get influenced by the stratification or other factors which results in the spectral coefficient depicting different power laws; for example \( n^{-2} \) dependence in an \( nS(n) \) vs \( n \) plot showing the existence of buoyant range limiting the -2/3 region of the inertial subrange.

The \((1/n^i)\) law serves as a powerful way of describing music, speech, and noise. In the \( S(n) \) or \( nS(n) \) versus \( n \) plot, one can identify different types of noise depending on the values of \( i \). These noises have some physical connotation. The nomenclature used to describe the noise in a \( S(n) \) vs \( n \) plot is: \((1/n^0)\) noise is called the white noise; \((1/n)\) noise is called the pink noise used in acoustic research; \((1/n^2)\) noise is called the Brownian noise and is very dependent on the frequency (Gardner, 1978); and finally the \((1/n^3)\) noise is called the black noise. White noise is independent of frequency and hence independent of its past. In acoustic research the sounds having frequency generated by \((1/n)\) sources were found to be pleasing, while those generated by \((1/n^2)\) sounded too correlated, and those sounds generated from white noise, namely \((1/n^0)\) sources, sounded too random (Voss and Clark, 1975; 1978). Black noise is encountered in natural and artificial catastrophes like floods and electric power outages (Schroeder, 1991).

Thus the so called noise of different types could have different physical connotations for the observed spectrum from different sources. The black noise for example given by \((1/n^3)\) slope possibly defines the existence of “buoyant subrange” which could be encountered for atmospheric turbulence in stratified conditions. The
stable boundary layer spectrum of the real atmosphere must be affected by both waves and turbulence. The spectrum in such cases exhibits three subranges (Stull, 1988). Similarity arguments give “-1 power law” in the S(n) vs n spectrum at the large eddy end. Buoyancy subrange is characterised by “-3 power law” in the S(n) vs n spectrum. In buoyant subrange, eddies are of middle-size and are quasi-two-dimensional. A buoyancy wave number is defined as the inverse of the Ozmidov scale: \((\epsilon/N_{BV}^3)^{1/2}\), where \(\epsilon\) is the turbulence dissipation rate and \(N_{BV}\) is the Brunt-Vaisala frequency. Buoyancy wave number separates the buoyancy subrange at middle wave numbers from the inertial subrange at high wave numbers. The eddies are three dimensional at high wave numbers and follow the “-2/3 law” in the inertial subrange \((nS(n) vs n)\). In the potential temperature spectrum of Nai-Ping (1983) a gap exists between the buoyancy and the inertial subranges. Finnigan et al. (1984) explains that the energy transfer from the inertial subrange towards lower wave numbers is blocked at the “-2 power law” \((nS(n) vs n)\) region of the spectrum, leading eventually to a build-up of energy near the buoyancy wavenumber.

In the spectrum of atmospheric turbulence one should look for the presence, if any, of these noises, because one cannot rule out the possibility of these under disturbed weather conditions like the summer monsoon.

1.11 ENERGY BALANCE IN ABL

Flow in the ABL is controlled by the diurnal cycle of the surface energy budget. The energy balance at the surface can be expressed as,

\[
R_n + G + H + E = 0
\]  

(1.22)

where \(R_n\) is the flux of net radiation (Global Solar radiation received by the surface plus atmospheric radiation minus terrestrial radiation), \(G\) is the vertical heat flux into the soil, \(H\) is the sensible heat flux and \(E\) is the latent heat flux (resulting from water phase changes) to the atmosphere.

During the day the energy gained by the surface is transferred to the atmosphere, to the soil and is also used in the evaporation processes. This transfer of heat from the ground surface to the air directly above can generate convection. Convection redistributes heat throughout the atmospheric boundary layer. The influence of the surface sensible heat flux decreases with height. As a result, the temperature amplitude also decreases with height, becoming zero on the top of the
atmospheric boundary layer. Resulting temperature stratification can be classified into three distinct categories; stable, unstable and neutral.

In the case of stable stratification, the temperature of the atmosphere is distributed such that any fluid particle moving vertically is impeded by Archimedean force. This suppresses turbulence and can lead to laminar flow. Stable cases usually occur at night. In unstable stratification, the parcel of air moving vertically is accelerated by Archimedean force. This mechanism increases turbulence. Unstable cases usually occur during day light hours. In the neutral case, observed during transitions between day and night, the buoyancy forces vanish.

1.12 THE RICHARDSON NUMBER

Richardson number (Ri) is defined as the relative importance of buoyancy and shear on producing turbulence.

\[
R_i = \frac{\left( \frac{g}{T} \right) \left( \frac{\partial \bar{\theta}}{\partial z} \right)}{\left( \frac{\partial \bar{u}}{\partial z} \right)^2}
\]  

(1.23)

where \( g \) is the acceleration due to gravity, \( T \) the mean ambient temperature in degree Kelvin, \( \bar{\theta} \) the mean potential temperature and \( \bar{u} \) the mean wind at height \( z \).

The advantage in using this number is that it contains gradients of mean quantities that are easy to measure. \( R_i \) is positive for stable stratification, \( R_i \) is negative for unstable stratification and \( R_i \) is zero for neutral stratification.

1.13 OBJECTIVES OF THE THESIS

This thesis presents some important results on atmospheric boundary layer characteristics over the semi arid region of western India, West and East coast and peninsular regions of India by using meteorological tower, radiosonde, tethersonde observations and also from high resolution measurements using Eddy covariance systems.

The detailed objectives of this thesis are:

- To understand the evolution of boundary layer in various terrain and environmental conditions.
• To understand the boundary layer characteristics during pre-monsoon, active and break monsoon conditions.
• To examine the diurnal and seasonal variation of boundary layer parameters at different stations.
• To investigate the diurnal, biweekly and seasonal variability in different parameters such as wind speed, temperature, humidity, water vapor, carbon dioxide and heat fluxes in the atmospheric surface layer.

The thesis has been mainly divided into nine chapters. **Chapter 1** forms the introduction on the atmospheric boundary layer and its significance, boundary layer characteristics and the spectral characteristics of turbulence. The **Chapter 2** describes the instrumentation and different methods of measurement techniques used to study the boundary layer characteristics in various field experiments under different terrain. The spectra of zonal, meridional and vertical components of wind and temperature component over the semi-arid region of Anand and west coast station, Vasco-da-Gama are presented and discussed in **Chapter 3**. In **Chapter 4** the turbulence characteristics of wind and temperature in the atmospheric surface layer in different terrain condition are described. The vertical profiles of wind speed, air temperature and relative humidity over Vasco-da-Gama in the west coast and Anupuram in the east coast are discussed in **Chapter 5**. **Chapter 6** of the thesis discusses the variation of wind speed, temperature and relative humidity over Anand and Khandha. **Chapter 7** explains the Doppler SODAR observations of three dimensional wind fields and thermal structure of convective boundary layer (CBL). **Chapter 8** presents and discusses the simulation of net radiation, skin temperature, and soil temperature at various depths and surface sensible and latent heat fluxes in wet and dry surface conditions which are compared with observations from LASPEX. A summary of the major results of the investigations made and future scope of research work are provided in **Chapter 9** of the thesis.