CHAPTER 2

INSTRUMENTATION, MEASUREMENT TECHNIQUES AND ANALYSIS METHODS

2.1 INTRODUCTION

The Indian Institute of Tropical Meteorology (IITM), Pune, has been involved in conducting and participating in several national and international meteorological field experiments like the International Indian Ocean Expedition (IIOE-1963), the Indo-Soviet Monsoon Experiment (ISMEX-1973), the Monsoon Experiment (MONEX-1979), Monsoon Trough Boundary Layer Experiment (MONTBLEX-1990), the Land Surface Processes Experiment (LASPEX-1997), the Indian Ocean Expedition (INDOEX-1999) and the Bay of Bengal Monsoon Experiment (BOBMEX-1999), to mention a few. IITM has good experimental facilities and expertise for conducting micro-meteorological measurements on wind, turbulence, the thermal structure of ABL and also cloud microphysics, aerosols and atmospheric electricity.

In tropical latitudes also a few boundary layer experiments were conducted. Because of the dominance of moist processes and the smallness of the coriolis force, the atmospheric boundary layer in the tropics differs from that in the mid and high latitudes. Boundary layer observations near the surface were conducted on a tower at the agrometeorological observatory, Poona over eight decades back by Ramdas and his colleagues (Ramdas and Atmanathan 1932; Ramdas 1932). One of the most intriguing findings of these early studies was the presence of a lifted temperature minimum on calm clear winter nights over bare surfaces, which has been theoretically explained by Vasudeva Murthy et al. (1993). The IITM launched a special experiment in 1966-67 over the Palghat region off the Kerala coast to study the peculiar characteristics of the local atmospheric boundary layer (ABL) as the flow rushes through a large gap in the western Ghat (Ramachandran et al., 1980). Monostatic sodar was installed at National Physical Laboratory, New Delhi in 1980 (Singal et al., 1982) to study stratified layers for application to line-of-sight microwave communication links in the country. The monsoon boundary layer over the landmasses of India, and particularly over the monsoon trough region, has been hardly studied in any organized manner. The problems connected with moist and dry
convection within the convective boundary layer in the monsoon trough are unique in some respects. The monsoon trough is the most important feature in the lower troposphere over India during the summer monsoon season (June to September) and has been known over a century. There are major differences in the dynamical and convective characteristics of the eastern and western end of the trough. The western end corresponds to a heat low with very shallow ascent and convergence limited to the lowest half kilometer or so and clear skies above, whereas the eastern end is a so-called dynamic trough with convergence up to the mid-troposphere and moist ascent throughout the troposphere with intense convection and clouding.

The Boundary Layer Division of the IITM, Pune, has been conducting several field experiments and has experimental facilities like Sound Detection and Ranging (SODAR), Tethered balloon, Sonic anemometers, Solar radiation measuring equipment and meteorological masts, to mention a few, which can generate high resolution data on the atmospheric boundary layer parameters.

2.2 OBSERVATION INSTRUMENTATION AND DATA ACQUISITION

2.2.1 ABL Field Experiments (National)

In India, two major national boundary layer field experiments were conducted over land during the last decade. They are Monsoon Trough Boundary Layer Experiment (MONTBLEX) in 1990 and Land Surface Processes Experiment (LASPEX) in 1997. The objectives of these experiments were to understand the coupling mechanisms among sub-surface, surface and planetary boundary layer and their role in the monsoon dynamics. A lot of data were generated using the state-of-the-art in situ as well as remote sensing instruments.

The radiosonde, Sodar and Kytoon (tethered balloon payload) system in MONTBLEX were used to study the thermodynamics of ABL. LASPEX-1997 aimed to study the seasonal variation of land surface parameters in terms of different vegetation and soil characteristics. Low level stability of the ABL at Anand (one of the LASPEX sites) was studied using radiosonde and micro-meteorological tower data.

To understand the characteristics of the coastal ABL and interaction of meso-scale with micro-meteorological scales, measurement of meteorological parameters and turbulence was made on the west coast of India at NCAOR, Goa during another national experiment called the ARabian Sea Monsoon EXperiment (ARMEX) using
a 9 m high micro-meteorological tower. The data collected from the tower during ARMEX 2002-03 were analyzed to study the characteristics of coastal atmospheric surface layer like development of Internal Boundary Layer (IBL), turbulent fluxes of heat, momentum and water vapor.

The phase-I of ARMEX was conducted during June 15 - August 15, 2002. The year 2002 was a weak monsoon year with little rainfall over India in July. The phase-II of ARMEX during March 15 – May 15, 2003 was the pre-monsoon period with instability building up in the atmosphere due to intense solar heating of the surface.

### 2.2.2 Instrumentation during MONTBLEX

In deciding which instruments to use in measuring a particular quantity, the selection will depend, in most cases, on the problem to be studied. In this context, the aim of the MONTBLEX was to study and understand the ABL over dry, mixed and moist convective regions through which the monsoon trough passes in Indian summer monsoon. Accordingly for probing the surface layer, 30 m towers (with meteorological sensors at logarithmic intervals were mounted on it) were erected at Jodhpur, Varanasi and Kharagpur. Kytoon and Sodar were used to probe the ABL up to a height of 1 km. Details of the experiment and the results are compiled by Narasimha and Sikka (1997).

### 2.2.3 Instrumentation during LASPEX

LASPEX was carried out over watershed region of Sabarmati river basin in Gujarat State to study role of land surface processes on the sub-regional climate in an area known for extreme climate variability on the annual cycle as well as on sub-regional scale. The experimental site covered an area of 100 x 100 km² with different soil types and vegetation cover. The IITM, Pune served as a nodal agency for the experiment with active collaboration from the Gujarat Agricultural University (GAU), Anand. Several scientific organizations also contributed to the logistics of the experiment in different ways.

This land surface processes experiment in India was aimed at (i) improving the understanding of land surface processes over the semi-arid tropical region and (ii) to improve land surface process parameterization in the global model used for medium range weather forecasting.
During LASPEX experiment five micrometeorological towers of 10 m height were installed at Anand (22°35′N, 72°55′E), Sanand (23°04′N, 72°22′E), Derol (22°40′N, 73°45′E), Khandha (22°02′N, 73°11′E) and Arnej (22°40′N, 73°45′E) in addition to one 30 m tower at the central station, Anand. Different sensors to measure temperature, humidity, wind speed and direction were installed at 1, 2, 4 and 8 m height. Incoming and outgoing shortwave and long wave radiation and net radiation sensors were fixed at approximately 2 m height along with pressure and rainfall measuring systems. Soil temperature sensors were placed at the surface and at the depths of 5, 10, 20, 40 and 100 cm. At Anand, all the three components of the wind were measured by the Sonic and Metek anemometer systems fixed respectively at the height 4 and 9 m AGL. The data from these slow sensors were collected on a data logger at the sampling rate of 1 Hz whereas from fast sensors (Sonic and Lyman α Hygrometer systems) were collected through PC at 10 Hz sapling rate. Figure 2.1 shows the experimental setup on 9 m micrometeorological tower at Anand station during LASPEX.

2.2.4 Instrumentation during ARMEX

The objective of the ARMEX is to measure the wind, temperature and humidity profiles and study the surface fluxes of momentum, heat and moisture at Goa and also to study the time series of the measured variables for understanding the intra-seasonal and diurnal oscillations during the Indian summer monsoon. During the ARMEX phase I and phase II the IITM, Pune, installed a 9 m high micrometeorological tower, to measure the surface heat and momentum fluxes, 30 m away from the west of the coast in the premises of the National Centre for Antarctic and Ocean Research (NCAOR), Vasco-da-Gama (15°21′ N, 73°51′E), Goa, India. Measurements of turbulent fluxes of heat, momentum, water vapor, carbon dioxide and profiles of wind, temperature and humidity were made to study the characteristics of coastal atmospheric surface layer. The tower had instrumentation at 1, 2, 5 and 8 m above ground level (AGL).
Figure 2.1. The experimental setup on 9 m micrometeorological tower at Anand during LASPEX.

The various meteorological sensors exposed on the 9 meter tower were: (a) Wind direction (2 and 8 meter only), (b) Wind speed at 1, 2 and 8 meter, (c) Air Temperature at 1, 2, 5 and 8 meter, (d) Relative Humidity at 1, 2, 5 and 8 meter, (e) Rain Gauge at the surface and (f) Radiation sensors at 2 meter (Short wave and Long wave radiometers). Figure 2.2 shows the experimental setup on 9 m micrometeorological tower at Vasco-da-ma, Goa during ARMEX.
Apart from the above mentioned slow sensors, other fast sensors installed on the tower at 5 m AGL were:

(i) Sonic Anemometer  
(ii) CO₂ / H₂O Analyzer

The data from these slow sensors were collected on a data logger at the sampling rate of 1 Hz whereas from fast sensors (Sonic and CO₂ / H₂O Analyzer systems) were collected through PC at 10 Hz sampling rate. Table 2.1 shows details of the tower instrumentation operated during various field experiments.
### Table 2.1. Major Meteorological sensors/instruments operated during field experiment.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Sensor/Instruments</th>
<th>Range and accuracy</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wind speed</td>
<td>Cup anemometer</td>
<td>0 - 65 m/s, +/- 2% of FSR</td>
</tr>
<tr>
<td>Wind direction</td>
<td>Wind vane</td>
<td>0 - 357 deg., +/- 2 deg.</td>
</tr>
<tr>
<td>Air temperature</td>
<td>Platinum resistance Thermometer (RTD)</td>
<td>-40 – 60°C, +/- 0.1°C</td>
</tr>
<tr>
<td>Relative humidity</td>
<td>Solid state capacitance, IR Hygrometer (LICOR, USA)</td>
<td>0 - 99%, 2% of FSR, 0 – 42 gmm⁻³, 0.5 gmm⁻³</td>
</tr>
<tr>
<td>Solar radiation (incoming and outgoing)</td>
<td>Spectral pyranometer (Eppley, USA)</td>
<td>0.3 to 3 μm, 0.1 μm</td>
</tr>
<tr>
<td>Long wave radiation (incoming and outgoing)</td>
<td>Pyrgeometer (Eppley,USA)</td>
<td>3 to 60 μm, 0.1 μm</td>
</tr>
<tr>
<td>Precipitation</td>
<td>Tipping bucket rain gauge</td>
<td>unlimited, 1 mm</td>
</tr>
<tr>
<td>Wind</td>
<td>Sonic anemometer (Metek, Germany)</td>
<td>0 - 65 m/s, +/- 0.05 m/s</td>
</tr>
<tr>
<td>Virtual temperature</td>
<td>Sonic anemometer (Metek, Germany)</td>
<td>-20 to +50°C, +/- 0.05°C</td>
</tr>
<tr>
<td>CO₂/H₂O Concentration</td>
<td>LI -7500 CO₂/H₂O(Open – Path gas Analyzer)</td>
<td>0-3000 ppm for CO₂, 0-60 ppt for H₂O, within 1% of reading</td>
</tr>
</tbody>
</table>

#### 2.2.5 Instrumentation during IGOC

Various types of instruments were installed during Integrated Ground Observation Campaign (IGOC) at Mahabubnagar (16.44°N, 77.59°E) on 20 meter micro-meteorological tower at various levels like slow sensors for wind, temperature and humidity at 5 levels (2, 4, 8, 12, and 18 m) etc. Eddy covariance sensor measuring the turbulent fluctuations in wind, temperature, water vapor and CO₂ (using sonic anemometer and infrared gas analyzer LI-7500) was installed on the
tower at the height of 6 meter. A sonic anemometer is also installed at 16 meter heights to capture wind components u, v and w. The soil temperature and moisture at various depths up to 1 m were also recorded using soil temperature and moisture sensors. A typical set up of the tower during IGOC experiment at Mahabubnagar is shown in Figure 2.3. In addition to these measuring instruments mounted on tower, a short platform near the tower was installed to measure radiation components viz. shortwave and long wave radiation both incoming and outgoing. The data from the slow sensors were collected on a data logger at the sampling rate of 1 Hz whereas from Eddy co-variance system (Sonic and CO₂ / H₂O Analyzer) the data collected through Laptop PC at 10 Hz sampling rate. The set of sensors/instruments used along with their sampling frequency are listed in Table 2.2 shown below:

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Sensor/ Instruments</th>
<th>Sampling frequency</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wind Speed</td>
<td>Cup anemometer</td>
<td>1s</td>
</tr>
<tr>
<td>Wind direction</td>
<td>Wind vane</td>
<td>1s</td>
</tr>
<tr>
<td>Relative humidity</td>
<td>Humicap</td>
<td>1s</td>
</tr>
<tr>
<td>Air temperature</td>
<td>Platinum RTD element (PT1000)</td>
<td>1s</td>
</tr>
<tr>
<td>Soil temperature</td>
<td>Encapsulated platinum wire</td>
<td>1s</td>
</tr>
<tr>
<td>Net Radiation</td>
<td>Thermopile (Kipp and Zonen)</td>
<td>1s</td>
</tr>
<tr>
<td>Sonic anemometer</td>
<td>3-axis ( Metek Inc. Germany )</td>
<td>0.1s</td>
</tr>
<tr>
<td>CO₂/H₂O Concentration</td>
<td>L1 -7500 CO2/H2O(Open –Path gas Analyzer)</td>
<td>0.1s</td>
</tr>
</tbody>
</table>

The sensors that were mounted at various levels and platforms were connected to data logger to store observations into the dynamic range of the data system. Instruments on tower are cup anemometer and temperature and humidity sensors at 1, 2, 4, 8, 12 and 18 meter, wind vane at 2, 4 and 18 meter, and two Eddy co-variance (EC) systems (fast response sonic sensor and water vapour/CO₂ analyzer) at 6 and 16 meter installed on 20 meter micrometeorological tower (figure 2.3). In addition to above sensors, the soil temperature sensors at the depth of 10, 20, 40, 60, 100 cm and soil moisture sensors at the depth of 20, 40, 100 cm were also installed.
2.2.6 *Eddy covariance instrumentation*

Eddy covariance measurements are made at a single height, meaning that only turbulent transport contributes to the measured flux. The second advantage of eddy covariance is that since the measurements are made at high frequency and resolve the effects of individual eddies, spectral, co-spectral and probability density analyses can be employed to determine the mechanisms responsible for the flux. For eddy...
covariance the three-dimensional wind field is measured rapidly, and the value of the scalar of interest (e.g. temperature, water vapour density, CO₂ concentration) is measured at the same frequency and location.

The covariance of the vertical wind speed and the scalar value is equal to the flux at the measurement point. The covariance is calculated over periods of fifteen to thirty minutes depending on stationary issues. The following sections define the eddy covariance flux and discuss some requirements and limitations. The instrumentation shown in the image (Figure 2.4) is typical of an eddy covariance installation with the 3-dimentional Sonic Anemometer and H₂O/CO₂ Infra Red (IR) gas analyzer. It is given by

1) Eddy covariance system consisting of an ultrasonic anemometer and infrared gas analyzer (IRGA).

2) The Eddy covariance technique is a key of atmospheric flux measurement technique to measure and calculate vertical turbulent fluxes within atmospheric boundary layer.

3) It is a statistical method used in meteorology and other applications that analyzes high frequency wind components (u, v and w) and temperature and scalar atmospheric data series, and yields values of fluxes of these properties. Such flux measurements are widely used to estimate momentum, heat, water, and carbon dioxide exchange.

4) Vertical flux can be presented as a covariance of the vertical wind velocity and the concentration of the entity of interest.

Figure 2.4 shows Sonic Anemometer with Open-path Infrared gas H₂O/CO₂ Analyzer installed at 6 m height over Mahabubnagar during IGOC experiment. Sonic Anemometer measures 3 dimensional wind velocity and temperature and open-path infrared gas analyzer measures the water vapor and carbon dioxide density and the concentration with fast response.
2.3 METHODS FOR DETERMINING MOMENTUM, HEAT AND MOISTURE FLUXES IN THE ATMOSPHERIC SURFACE LAYER

Determination of turbulent exchange taking place between the earth and the atmosphere near their interface (surface) is of primary concern in micrometeorology. The emphasis here is on principles and techniques rather than on the details of instrumentation and measurements (Lenschow, 1986).

The various methods of determining momentum, heat and moisture fluxes are,

1. Profile method
2. Bulk transfer method
3. Eddy correlation method
4. Energy balance method
5. Gradient method
6. Geostrophic departure method
7. Thermodynamic energy equation method
8. Variance method

The eddy correlation and profile methods only are discussed in detail below since the results described in the thesis are mainly based on these two methods.
2.3.1 Eddy correlation Method

The eddy correlation method is the most reliable and direct measurement of turbulent exchanges of momentum and heat in the atmosphere are usually made with sophisticated fast response turbulence instrumentation. If all fluctuations of velocity and temperature that contribute to the desired momentum and heat fluxes are faithfully sensed and recorded, one can determine their co-variances simply by averaging the products of the appropriate fluctuations over any desired averaging time.

In particular, the vertical fluxes of momentum, heat and moisture over a homogeneous surface are given by,

$$
\tau_0 = -\rho u'w' \tag{2.1}
$$

$$
H_0 = \rho C_p \partial w' \tag{2.2}
$$

$$
E_0 = -\rho C_w u'q' \tag{2.3}
$$

where $\rho$ - density of air and
$C_p$ - specific heat at constant pressure
$C_w$ - heat transfer coefficient for water

The above fluxes are expected to remain constant, independent of height in a horizontally homogeneous surface layer. Eddy correlation measurements also provide a means of determining the surface fluxes. Although the eddy correlation method of determining fluxes is simple, in practice, it requires expensive research grade instrumentation, such as sonic, laser or hot wire anemometers and thin wire resistance thermometers, as well as rapid (sampling rates of 10-100 s\(^{-1}\)) data acquisition systems. The closer to the surface the turbulence measurements are made, the more severe become the instrument response problems. At heights greater than 10 m or so, light cup anemometers may also be adequate for measuring variances and fluxes.

Larger averaging times may be required, however, with increasing height of measurement, because the characteristic size of large eddies usually increases with height in the PBL. The requirements of instrument leveling, orientation, calibration and maintenance are also quite severe for accurate eddy correlation measurements (Lenschow, 1986). The above mentioned instrumental requirements have kept the eddy correlation method from being widely used, except in special research.
expectations. This is the only method that has the advantage of measuring turbulent exchanges directly, without too many restrictive assumptions about the nature of the surface (such as uniform, flat and homogeneous) or of the atmosphere. It is the only method available for measuring turbulent fluxes inside plant canopies, or in the wakes of hills and buildings.

2.3.2 Profile method

A graph of atmospheric parameters such as air temperature, humidity, and wind speed data are available from instrument setup at any two levels versus time scale. Data captured with the frequency of 1 second is averaged to half hour mean. This averaged ‘aerodynamic profile’ is further useful in quantifying vertical transfers of sensible heat energy, moisture, and momentum (e.g. Arya, 2001). If biases are within manufacturers’ specifications, relative calibration adjustments may not be needed. The logarithmic wind profile is valid and the wind gradient \( \Delta \bar{u}/\Delta z \) is found to be inversely proportional to the height above the surface \( z \). Since the constant proportionality can be equated to the slope of the neutral wind profile the equation follows that:

\[
\bar{u}_* = k z \left( \Delta \bar{u}/\Delta z \right)
\]  

(2.4)

After substituting \( \bar{u}_*^2 = \tau/\rho \) in the above equation, the vertical flux of horizontal momentum \( \tau \) in terms of the wind speed differences alone can be obtained by the equation:

\[
\tau = \rho K^2 Z^2 \left( \Delta \bar{u}/\Delta z \right)^2
\]  

(2.5)

With the measurement of momentum flux \( \tau \) and invoking the principle of similarity theory we can use ratio of fluxes by use of \( \tau \). For computation of the sensible heat flux under neutral conditions the formula is given by:

\[
Q_H = -\rho C_p K^2 Z^2 \left( \Delta \bar{u}/\Delta z, \Delta \bar{T}/\Delta z \right)
\]  

(2.6)

For computation of the latent heat flux under neutral conditions the formula is given by:

\[
Q_L = -\rho L_v K^2 Z^2 \left( \Delta \bar{u}/\Delta z, \Delta \bar{q}/\Delta z \right)
\]  

(2.7)
To use profile methodology one needs empirical parameterization between fluxes and mean profiles. Moreover this technique involves estimating fluxes that the best fit of the mean profile data. These similarity relations are semi-empirical functions which relate wind and temperature profiles to the surface fluxes of momentum, sensible and latent heat. In addition, turbulence in the atmospheric surface layer can be conveniently described in terms of surface layer similarity parameters.

2.4 MONIN-OBUKHOV SIMILARITY THEORY

Monin-Obukhov similarity theory is the basis of much of the micrometeorology used here. It applies to the surface layer (approximately the lowest 10% of the boundary layer) and is often referred to as surface layer similarity. In situations where the mechanisms determining the behaviour of an atmospheric parameter are incompletely understood, similarity can be used to derive empirical relationships, typically between height and the parameter in question. The approach is to non-dimensionalise groups of variables using scales such as the Obukhov length, and to look for common (‘similar’) behaviour in the non-dimensionalised variables. Other parameters often used to normalise variables include wind speed, measurement height, and friction velocity and roughness length.

Monin-Obukhov hypothesis states that, statistics of various surface layer parameters such gradient, variance and fluxes, when normalized by their appropriate power of the scaling velocity, scaling temperature and scaling humidity becomes universal functions of $z/L$. The Obukhov length represents the height in the stable surface layer below which shear production of turbulence exceeds buoyant consumption. Further interpretation can be made by dividing $L$ into two clear categories: stable and unstable condition. If $L < 0$, then it means that the temperature gradient is directing upward (the ground is warmer then the adjacent air with various turbulence and light wind), implying that the condition tend to happen rather daytime: - i.e. unstable conditions. On the other hand, if $L > 0$, then the temperature gradient is now downward (the adjacent air is warmer than the ground with week turbulence and moderate to strong wind), which tends to happen at night: - i.e. stable conditions. The third condition are neutral conditions, in which there is little heating and cooling at the boundary of the ground and the surrounding air , usually being from windy conditions. ‘$L$’ can be calculated as;
\[ L = -u^3_* / \left[ K(g / \overline{\theta_v})w' \overline{\theta_v} \right] \] (2.8)

Where \( \overline{\omega \theta_v} \) - Temperature flux at the surface

- \( g \) - gravitational acceleration,
- \( \theta_v \) - virtual potential temperature,
- \( u^* \) - friction velocity,
- \( L \) - Obukhov Length,
- \( K \) - von Karman constant (0.4)

Turbulence in the atmosphere is also measured by Richardson Number.

### 2.5 REMOTE SENSING TECHNIQUES

Remote sensors that measure the wind and map the distribution of aerosol and turbulence have contributed a lot to our understanding of the atmospheric processes over homogeneous and complex terrains. They have provided new insight into the initiation, evolution and destruction of nocturnal drainage winds, and into the complexities of transport and dispersion in an environment in which synoptic, mesoscale, and microscale processes interact strongly. Remote sensing instruments mimic conventional instruments in many respects and also provide much more extensive temporal and spatial coverage. Since they depend on the presence of aerosols or turbulence-produced refractive index fluctuations, they see the atmosphere differently than conventional instruments. For this reason, their data often provide a visualization of atmospheric processes that is not easily available from those of in-situ instruments.

The use of acoustic, optical, and radar waves to probe the lower atmosphere has received considerable attention over the past two decades. Lenschow (1986) provided the most recent comprehensive reference on remote sensors and their application to studies of the lower troposphere. Sodar, lidar, and radar (sound-, light-, and radio-detection and ranging) are all active remote sensing devices that emit a short pulse of radiation into the atmosphere within a narrow beam. A small fraction of the radiation scatters towards a receiver where it is amplified and isolated in a frequency band pass centered on the carrier frequency. The intensity and/or frequency shift is then analyzed as a function of range from the transmitter (using the time of flight to and from the scattering volume). The mean motion of scatterers
within the scattering volume produces a mean frequency shift from which one component of the wind can be deduced. Given a sufficiently strong scattered signal from which to deduce the frequency shift, suitable averaging, and radial measurements in several discreet directions (or in a continuous conical scan), the u, v, and w components of the wind can be obtained as a function of height.

Underlying each of the remote sensing techniques is the need for scatterers; aerosol for lidar techniques, and refractive index variations at small scales (usually less than 1 m) for sodars, clear-air radars, and optical crosswind sensors. In the case of radars and sodars, the tracers, namely turbulence-induced refractive index fluctuations, depend almost instantaneously on wind shear and stratification. In convective conditions, strong refractive index variations develop near the surface because of surface heating and evaporation of moisture from the surface. Subsequent vertical mixing then results in small-scale refractive index fluctuations detectable by radars and sodars. In statically stable conditions, an increased wind shear is required to produce turbulence and, consequently, refractive index fluctuations. This can occur either near the surface or in free shear layers. As a result, delineation of these flow-dependent tracers by remote sensors can define the surface mixing-layer depth as well as identify regions of dynamical instability and turbulent mixing aloft (Chadwick and Gossard, 1986; Neff and Coulter, 1986). Thus, properly interpreted, the spatial distribution and temporal evolution of turbulence microstructure should provide direct insight into the dynamics of many complex terrain processes.

Understanding the planetary boundary layer (PBL) structure is necessary in any atmosphere environmental study. To investigate this structure, knowledge of the wind field and the thermal structure becomes necessary. Its measurement can be made directly by meteorological towers, or by radiosonde or instrumented planes, or indirectly by means of remote sensors, such as radar, sodar or lidar. Direct exploration of the PBL has added much to the understanding of the PBL structure and processes. Remote techniques, however, have become more and more popular because they can scan volumes of the atmosphere making volume averages of selected variables possible. It has been virtually impossible to use direct sensors to explore the PBL due to the difficulty of deploying these sensors at all locations and altitudes throughout the layer (Stull, 1988). Neff (1975), Asimakopulos et al. (1983) and Singal et al. (1985) have used sodar to study the thermal and dynamic structure of the PBL.
2.5.1 Acoustic Scattering Theory

The turbulent velocity is a random, three-dimensional, time-dependent vector field. Its evolution in space and time is governed by the Navier-Stokes and continuity equation, which express momentum and mass conservation respectively. There are several concepts which are useful when dealing with planetary boundary layer (PBL) turbulence. We call turbulence stationary if its statistical properties are independent of time. In other words, stationary implies statistical invariance with respect to translation of the time axis. In the PBL this is often the case near mid-day in fair weather. If the field is statistically invariant to translation of the spatial axes we call it homogeneous. While PBL turbulence is near homogeneous in the vertical (it is strongly affected by the lower surface and the capping inversion) it can be horizontally homogeneous, to a good approximation. An isotropic field is statistically independent of translation, rotation and reflection of the spatial axes. This is clearly not the case in the PBL, since the upper and lower boundary conditions, and the effects of buoyancy, make properties different in the horizontal and vertical. However, experimental data and theoretical arguments suggest that the small-scale (meters and smaller) structure of the PBL is effectively isotropic. This is called ‘local isotropy’, meaning isotropy confined, or localized, to the smallest-scale structure.

There are three important scales associated with turbulence energy in the PBL. They are integral length scale (l), Taylor microscale (λ) and Kolmogoroff microscale (ηk). The eddies carrying the bulk of the turbulent kinetic energy and doing most of the turbulent transport are of integral length sized in magnitude. Taylor microscale is defined by the turbulence dissipation rate per unit mass (ε), the turbulence kinetic energy per unit mass (q²), and the molecular kinematic viscosity (ν) given as \( \lambda = (\nu q^2/\varepsilon)^{1/2} \). This scale indicates the eddy size where viscous effects begin to become significant. It roughly marks the small-scale end of the inertial subrange, that broad range of scales between the anisotropic, energy-containing eddies and the isotropic, dissipative eddies which convert kinetic energy into internal energy through their viscous friction. The Kolmogoroff scale ηk indicates the size of the viscous eddies. A turbulent fluid, having a kinematic viscosity ν and needing to dissipate kinetic energy at a rate ε, establishes a dissipative range of eddies scaling in size with \( \eta_k = (\nu^3/\varepsilon)^{1/4} \).

In convective PBL, the eddy size, l can range from several hundred meters to a few kilometers, \( \lambda \) can be about 0.1 m and the dissipative eddy size, \( \eta_k \sim 0.001 \) m. Spectral
region between 0.1 m and 1/λ constitutes inertial subrange which is about three decades (Kaiman and Finnigan, 1994). Tatarskii (1971) treats acoustic and electromagnetic wave propagation and scattering in a turbulent medium in detail. The scattering of sound in a turbulent atmosphere (locally isotropic and homogeneous) is described using the equation for the differential scattering cross-section:

\[
\sigma(\theta_s) = \frac{1}{8} k^4 \kappa_b^{-1/3} \cos^2(\theta_s) \left[ \frac{0.33 C_T^2}{T_o^2} + \frac{\cos^2(\theta_s/2)}{\pi C_o^2} 0.76 C_v^2 \right]
\] (2.9)

which represents the fraction of the incident acoustic power scattered by refractive index inhomogeneities, per unit distance into a unit solid angle oriented in the direction \(\theta_s\) from the direction of transmission and is written in terms of the temperature structure function

\[
C_T^2 = \frac{\langle (T(x+r) - T(x))^2 \rangle}{r^{2/3}}
\] (2.10)

and the velocity structure function

\[
C_v^2 = \frac{\langle (V(x+r) - V(x))^2 \rangle}{r^{2/3}}
\] (2.11)

where \(x\) and \(x+r\) are two measurement locations and where \(\langle \rangle\) indicates an ensemble average (but for practical purposes normally derived from a time average). In equation (2.9), \(k\) is the acoustic wave number; \(\kappa_b\) the Bragg wavenumber, \(\kappa_b = 2 k \sin(\theta_s/2)\); \(T_o\) the local temperature; and \(C_o\) the local speed of sound.

Equation (2.9) introduces an interesting angular dependence in the acoustic scatter; namely, temperature microstructure alone determines the backscatter cross-section (\(\theta_s=180^0\)) and neither temperature nor velocity microstructure contributes to scatter at right angles (\(\theta_s=90^0\)). In principle, measurements at two angles should allow determination of both \(C_T^2\) and \(C_v^2\). Velocity inhomogeneities contribute to the scattering of sound waves but not electromagnetic radiation because of the relatively slow speed of sound and the fact that typical velocity inhomogeneities in the atmosphere can significantly affect the phase speed of sound, producing scattering (Batchelor, 1957).

Equation (2.9) is written in terms of structure functions \(C_T^2\) and \(C_v^2\) because they provide a conceptually and experimentally useful description of the small-scale
refractive index structure that produces acoustic scattering. Underlying the derivation that leads to equation (2.9), however, are the results that refractive index variations at spatial scales corresponding to one-half an acoustic wavelength (or the Bragg scale for off-axis scattering) contribute most strongly to the acoustic scatter and that $C_T^2$ can represent the isotropic three-dimensional spectral density of temperature $\phi_T(k)$ through $C_T^2 = \phi_T(\kappa_b) / (0.033 \kappa_b^{-1/3})$. In addition, $C_V^2$ can similarly represent the turbulent kinetic energy spectral density $E(\kappa_b)$ through $C_V^2 = E(\kappa_b) / (0.70 \kappa_b^{-5/3}) = 2 \varepsilon^{2/3}$ where $\varepsilon$ is the dissipation rate of turbulent kinetic energy ($m^2/s^3$) and $\kappa_b$ indicates the spectral wave number sensed by the sodar. It is important to note that the above equations relating $C_T^2$ and $C_V^2$ to the respective spectral densities only apply to the portion of the spectra within the inertial subrange (Stull, 1988). Because the portion of the spectra at larger scales, outside of the inertial subrange, can be highly variable in magnitude and extent in wavenumber space and because variances are often obtained by integrating arbitrarily over some portion of the spectra, it is not always possible to relate the structure function $C_T^2$ and $C_V^2$ directly to the temperature variance and the total turbulent kinetic energy.

For most sodars, the scale of refractive index variation most important for scattering is a small fraction of a meter. Facsimile displays of echo intensity (involving either $C_T^2$ or $C_V^2$) reveal regions defined in height and time in which there is a significant cascade of temperature or velocity variance from large to small scales (Stull, 1988). This provides a pictorial representation, although qualitative, of turbulence generation/dissipation regions in space and time. Neff and Coulter (1986) provide a short review of the literature on acoustic scattering. The winds and other data collected from the phased array Sodar system reveals the signature of thunderstorm events during pre-monsoon and monsoon season over Pune. Table 2.3 below provides the detailed specifications of the Sodar system which is in operation at IITM, Pune.
Table 2.3. The detailed specification of Sodar system.

<table>
<thead>
<tr>
<th>Specification</th>
<th>Operating range</th>
</tr>
</thead>
<tbody>
<tr>
<td>No. of piezoelectric transducers</td>
<td>52</td>
</tr>
<tr>
<td>Frequency</td>
<td>1800–2500 Hz</td>
</tr>
<tr>
<td>Acoustic power (output)</td>
<td>100 W</td>
</tr>
<tr>
<td>No. of beams</td>
<td>3 (zenith, north, east)</td>
</tr>
<tr>
<td>Beam angle</td>
<td>16°</td>
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<tr>
<td>Maximum range</td>
<td>1000 m</td>
</tr>
<tr>
<td>Pulse width</td>
<td>180 milli second</td>
</tr>
<tr>
<td>Pulse interval</td>
<td>18 second</td>
</tr>
<tr>
<td>No. of FFT points</td>
<td>4096</td>
</tr>
<tr>
<td>Transmission type</td>
<td>Reflecting mode</td>
</tr>
<tr>
<td>Beam width</td>
<td>5°</td>
</tr>
<tr>
<td>Range resolution</td>
<td>30 m</td>
</tr>
</tbody>
</table>