DYNAMIC AND THERMODYNAMIC STRUCTURE OF THE MONSOON BOUNDARY LAYER OVER INDIA

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IN

ATMOSPHERIC SCIENCE

By

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CERTIFICATE

This is to certify that the thesis entitled '**DYNAMIC AND THERMODYNAMIC STRUCTURE OF THE MONSOON BOUNDARY LAYER OVER INDIA**' is a bonafide record of the research work done by Ms.Leena P in the Department of Atmospheric Sciences, Cochin University of Science and Technology. She carried out the study under my supervision. I also certify that the subject matter of the thesis has not formed the basis for the award of any Degree or Diploma of any University or Institution.

Certified that Ms. Leena P has passed the Ph D. qualifying examination conducted by the Cochin University of Science and Technology in October 1994.

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Contents

Preface

The atmospheric boundary layer is the portion of the atmosphere next to Earth's surface that is strongly influenced by interaction with the surface on time scales less than a day. This layer has depth and stability characteristics that vary temporally on the diurnal, meso, and synoptic and seasonal scales and spatially due to changing characteristics of the surface and intensity of weather patterns. This layer is responsible for exchanging heat, moisture and momentum with the underlying land or ocean surface and is thus of crucial dynamical and thermodynamical importance. Turbulence is the important physical process in the boundary layer. Turbulence is generated as a result of the instabilities caused either by wind shear or by thermal convectivity. The transport of heat, moisture and momentum in the vertical is by turbulence. The depth of the atmospheric boundary layer may range from a few tens of meters in highly stable night time conditions to even 3km in thermal convective conditions during summer afternoons over the Indian land area, when the inversions and stable layers are absent. For average conditions in the tropics the layer extends upto a depth of about 1km. The surface layer, which is the lowest part of the boundary layer, comprises of 1/10th of the total atmospheric boundary layer. Sharp variations of meteorological parameters occur in this layer. Solar energy, the primary source of energy for the atmosphere is absorbed at the surface and transmitted upward by turbulent processes. Turbulence is a continuous process in this layer. Also this is the layer in which human beings, vegetation and animals live and where most human activities take place.

No systematic study was carried out to quantify the variations that occur in the surface layer during the pre-monsoon, onset and post-monsoon periods and during the occurrence of systems over the Indian region. The objective of this study is to examine the dynamic and thermodynamic structure and the variations that occur in the surface layer during the pre-monsoon, onset and post-monsoon periods over the Indian region. Also the variations caused during the occurrence of micro and mesoscale systems are studied. The structure and variation in the marine boundary layer over the Indian region is also investigated.

The doctoral thesis consists of six chapters. In chapter 1 apart from general introduction on the topic, brief descriptions on the major physical processes and general climatology in the atmospheric boundary layer, international and national scientific experiments conducted in the boundary layer and a review of literature carried out on the topic is given. Chapter 2 consists of the structure and variations that occur during the pre-monsoon, onset and post-monsoon periods in the atmospheric boundary layer over India. The data used and computational procedure is also described. Also the boundary layer variations during the occurrence of thunderstorms and depressions are also given. It was found that the turbulent fluxes increase during the systems and onset. The depression affects the boundary layer characteristics of the entire subcontinent as soon as it is formed. The atmosphere tends become less unstable by the monsoon onset and it tends more unstable conditions after the withdrawal of monsoon.

The variations in the drag coefficients during the various periods and its spatial variations were given in chapter 3. The computational procedure is also given. The drag coefficient values were found to increase with instability and decrease with wind speed. Steep increase was noticed in the case of low winds. In chapter 4 the study on marine boundary layer over the Bay of Bengal is given. In this study slow data as well as fast response data from BOBMEX Pilot Experiment was used. Surface fluxes are computed using various methods and are compared. The surface fluxes were found to be very high during systems and thunderstorms. The boundary layer characteristics computed using the various methods are found to have similar trend. The thermodynamic structure and their variations during the different periods are given in chapter 5. Detailed method of computation is given in the chapter. Considerable variations were noticed in the thermodynamic structure during various seasons. It also was found to vary during thunderstorms. The zonal wind structure shows that the westerly wind depth and speed increases as the monsoon reaches each station in the subcontinent. The precipitable water content over the Arabian Sea and adjacent Indian region also increases by the onset.

An overall summary of the research results from the study is given in chapter 6, which is the conclusion chapter of the thesis. The list of references in the thesis is included at the end of the thesis in an alphabetical order. The outcome of the study is expected to provide a better understanding of the structure and variations in the boundary layer over India, which is useful for many applications especially for numerical modelling studies.

CHAPTER 1

Introduction

The Atmospheric Boundary Layer (ABL) or the Planetary Boundary Layer (PBL) is usually defined as the lowest portion of the atmosphere that is strongly influenced by the thermal and dynamical interactions of the underlying surface. This layer has depth and stability characteristics that vary temporally on the diurnal, meso, synoptic and seasonal scales and spatially due to changing characteristics of the surface and intensity of weather patterns. Significant exchange of momentum, moisture, heat and mass between the underlying surface and the atmosphere and sharp variations in velocity, temperature and mass concentration occur in this layer and is thus of crucial dynamical and thermodynamical importance. The main physical feature of the ABL, which is also sometimes used as an alternative definition is turbulence. This turbulence is due to the instabilities, which are induced by horizontal and vertical shears of wind close to the surface. The influence of surface friction and surface heating is quickly transmitted to the ABL by turbulent transfer. The vertical transport of momentum, heat, and pollutants are also by turbulence.

The ABL depth varies over a wide range from a few tens of meters to several kilometers. It depends on the rate of heating of the surface, surface winds, surface topographical characteristics and roughness, large-scale vertical motion, horizontal advection of heat and moisture and other factors. When the boundary layer is neutrally stable its depth may be defined as the height where the frictional effects are marginal or the wind is given by the geostrophical balance provided the horizontal acceleration is insignificant (Holton 1992). Under convectively unstable conditions the top of ABL is marked by a capping inversion and under highly stable conditions the height of the inversion at surface gives the boundary layer depth. The depth of ABL is found to reach even up to 3km in thermal convective conditions during summer afternoons over India when the inversions and stable layers are absent (Ananthakrishnan and Rangarajan 1963). For average conditions in the tropics the ABL extends up to a depth of about 1km (Sikka and Narasimha 1995). The ABL in the tropics differ from those in the middle and high latitudes due to the smallness of coriolis force and dominance of moist processes. The

ABL also differ over land and ocean. Over oceans the depth of ABL varies relatively slowly in space and time. The sea surface temperature changes little over a diurnal cycle because of mixing within the ocean. Also the water has large heat capacity causing little changes in temperature. Hence the surface forcings vary slowly. Most changes in the oceanic boundary layer are due to synoptic and mesoscale disturbances which cause vertical motion and advection of different air masses over the sea surface. In regions of low pressure the upward motions carry the boundary layer air to greater altitudes through the troposphere. This cause difficulty to define the boundary layer height and often the cloud base is taken as the height of the ABL.

Significance of the Atmospheric Boundary Layer

Although the ABL comprises only a tiny fraction of the atmosphere it is the layer in which human beings, animals and vegetation live and in which most human activities takes place. The small-scale processes occurring within the boundary layer are useful to various human activities and are important for the well being and even survival of life on earth. Solar radiation which is the primary source of energy is mostly absorbed at the ground and transmitted to the atmosphere by boundary layer processes. The radiation absorbed by oceans causes evaporation. The latent heat stored in water vapour accounts for 80% of the fuel for atmospheric motions. The water vapour when condensed on aerosols produces cloud nuclei, which are stirred into the air from surface by boundary layer turbulence. Exchange of carbondioxide between plants and animals occurs with in the ABL as a result of turbulence. The atmospheric turbulence also reduces the fatal poisoning of the life on earth by efficient diffusion and mixing of the pollutants released near the ground. The boundary layer processes also distribute pollen and other seeds of life to wide areas. Turbulent exchange of latent heat and sensible heat is essential for radiation balance and heat energy budget of the earth. The turbulent transfer of momentum from the atmosphere occurs through the boundary layer to the earth's surface, which is the sink for the atmospheric momentum. Wind stress on the sea surface is responsible for the generation of waves and ocean currents. When accompanied by strong wind turbulence can be even harmful to life and vegetation. Exchange processes in the boundary layer have significant influence on local weather. Boundary layer friction is responsible for low level convergence and divergence of flow in low pressure and high pressure regions. The frictional convergence in moist boundary layer air causes low level moisture convergence in low pressure systems. Even though the ABL comprises of a tiny fraction of kinetic energy of the atmosphere about 50% of the atmospheric kinetic energy is dissipated in the boundary layer by small-scale turbulence. Thus it is evident that the boundary layer processes affect our lives directly and indirectly by influencing the weather. Hence it is important to examine the dynamical and thermodynamical processes that occur with in the ABL. In this study the dynamic and thermodynamic structure and the variations that occur in the surface layer during the pre-monsoon, onset and post-monsoon periods over the Indian region are investigated. The variations caused during the occurrence of synoptic scale and mesoscale systems were also studied. The structure and variation in the marine boundary layer over Bay of Bengal was examined.

Structure of the Atmospheric Boundary Layer

The boundary layer has a well-defined structure. Region up to a few millimeters from the earth's surface where vertical transports are mainly by viscosity and molecular diffusion is the molecular sublayer. Above this layer is the surface layer which is approximately 10% of ABL. In this layer the variation of turbulent fluxes and stress is very less. Above this layer is the mixed layer where the vertical gradients of mean properties are very small as a result of turbulent mixing. The turbulence in the mixed layer is usually convective driven hence it is also known as convective boundary layer. At the top of the mixed layer an inversion layer may exist which is the interface between the turbulent boundary layer and free atmosphere where the flow is near geostrophic balance. At nighttime sometimes stable boundary layer where temperature drops less rapidly with height is seen above the surface layer. The height of the top of this stable boundary layer is usually taken as the height of the surface inversion.

Although the ABL is evolving continuously in response to the heating and cooling of the earth's surface it does have distinct states. Following the sunrise the mixed layer begins to grow in depth. The surface inversion, which prevailed during nighttime, evolves as the capping inversion layer. It reaches its maximum height in late afternoon. This layer is characterised by intense mixing where thermals of warm air rise from the ground. It grows by entrainment process or by mixing the less turbulent air above the layer. By sunset the capping inversions weaken and shallow inversion form below it. As the ground cools quickly the air above it cools and mixes upward by turbulence generated by wind shear. The surface inversion layer grows steadily to form a stable boundary layer. This layer is characterised by strong wind shear, small eddies and occasionally by internal gravity waves. This sequence of boundary layer evolution is typical for land surface in the midlatitudes. In the tropics the base of the trade wind inversion serves as the top of boundary layer. The subcloud layer act as the inversion if clouds are present and the base of this subcloud layer is taken as the top of the boundary layer.

Energy balance at the earth's surface

The important energy fluxes at the earth's surface are the net radiation to or from the earth's surface, the sensible heat flux and latent heat flux to or from the atmosphere and the heat flux into or out of soil or water. The net radiation flux is as a result of the radiation balance between the shortwave (R_S) and longwave (R_L) radiation at or near the surface, which can be written as

$$R_N = R_S + R_L \qquad (1.1)$$

where $R_S = R_S\downarrow + R_S\uparrow$

and $R_L = R_L\downarrow + R_L\uparrow$

where the downward and upward arrows denote incoming and outgoing radiation components respectively. Therefore the overall radiation balance can be written as

$$R_N = R_S\downarrow + R_S\uparrow + R_L\downarrow + R_L\uparrow \qquad (1.2)$$

The incoming shortwave radiation ($R_S\downarrow$) consists of both the direct and diffuse solar radiation, which is also called insolation and can be easily measured. The outgoing shortwave radiation ($R_S\uparrow$) is the fraction of $R_S\downarrow$ that is reflected by the surface, i.e.,

$$R_S\uparrow = -\alpha\, R_S\downarrow \qquad (1.3)$$

where α is the surface albedo. Thus for a surface the net short wave radiation $R_S=(1-\alpha)R_S\downarrow$.

The incoming longwave solar radiation ($R_L\downarrow$) from the atmosphere in the absence of clouds depends on the distribution of temperature, water vapour and carbondioxide. It does not have any diurnal variation. The outgoing longwave radiation ($R_L\uparrow$) is proportional to the fourth power of surface temperature and shows strong diurnal variation with its maximum value in early afternoon and minimum at dawn. The two components are usually of the same order of magnitude therefore the net longwave radiation R_L is generally a small quantity. Therefore under clear skies during daytime $R_L << R_S$ and the radiation balance

$$R_N \cong R_S = (1 - \propto) R_S\downarrow \qquad (1.4)$$

At night $R_S\downarrow = 0$ and the radiation balance becomes

$$R_N = R_L = R_L\downarrow + R_L\uparrow \qquad (1.5)$$

Usually at night $R_L\downarrow < - R_L\uparrow$. Thus during the daytime the net radiation is almost directed towards the earth's surface while at night it is directed away from the surface. As a result the surface warms up during the daytime while it cools during the evening and night hours especially under clear sky and undisturbed weather conditions.

The sensible heat flux at and above the surface arises as a result of the difference in temperatures of the surface and the air above. The temperature gradient in the vertical usually decreases with height in the atmospheric surface layer. While the heat transfer is by conduction through molecular exchange at the interface between the atmosphere and earth's surface, the transfer above this molecular sub layer is by convection through mean and turbulent motions in the air. The sensible heat flux is usually directed away from the surface during daytime when the surface is warmer than the air above and it is directed towards the surface during evening and night.

The latent heat flux is as a result of evaporation, evapotranspiration or condensation at the surface. Evaporation occurs from water surfaces and moist soil and vegetative surfaces whenever the air above the surface is drier. Evaporation occurs in the daytime and condensation may occur on colder surfaces at nighttime. Heat exchange takes place only when there is phase change between liquid water and water vapour. Evaporation results in cooling of the surface. The heat exchange through the soil surface is due to conduction. Through water the heat is transferred by conduction in the top few millimeters from the surface and then by convection in the deeper layers.

The energy budget at the surface, which is not flat and horizontal and is having changes of energy fluxes in the horizontal, can be written as

$$\overline{R}_N + \overline{H} + \overline{H}_L + \overline{H}_G + \Delta H_S \qquad (1.6)$$

where R_N is the net radiation, Hand H_L are the sensible and latent heat fluxes to or from the air, H_G is the ground heat flux to or from the ground or water and ΔH_S is the change in energy storage per unit time per unit volume V of the surface. The overbar over the flux quantity denotes the average value over the area A. The rate of energy storage is given by

$$\Delta H_S = 1/A \int \partial/\partial t\ (\rho\ C\ T)\ dV \qquad (1.7)$$

where ρ is the mass density C is the specific heat and T is the absolute temperature of the region. The energy storage term can be interpreted as the difference between the energy coming in (H_{in}) and energy going out (H_{out}), where H_{in} and H_{out} are combinations of R_N, H, H_L and H_G depending on their signs. When the energy input exceeds the outgoing energy there is flux convergence which results in warming of the region. When the energy going out exceeds that coming in the region cools as a result of flux divergence.

Meteorological properties in the planetary boundary layer

Surface and subsurface temperature:

The surface temperature at a given location is given by the surface energy balance, which in turn depends on radiation balance, exchange processes in the atmosphere, vegetation and thermal properties of the subsurface medium. The surface temperature is determined often by extrapolation of air and soil temperatures or by using remote sensors. The satellite monitored surface temperature was found to be reliable in the case of oceanic surfaces but not for land surfaces. The maximum value of surface temperature is noticed one or two hours after the time of maximum insolation, while the minimum is observed in the early morning hours. The surface temperature varies according to the texture of the soil. The presence of moisture at the surface and subsurface reduces the surface temperature due to evaporation and also due to increased heat capacity and thermal conductivity of the soil. The ground heat flux is also reduced by evaporation. The surface temperature is reduced under vegetation also. The incoming solar radiation is intercepted by vegetation thus reducing the surface temperature during daytime. The vegetation also cause increase in latent heat flux due to evapotranspiration and increase in latent and sensible heat flux occurs due to turbulence increase. The combined effect of all these reduce the range of surface temperature. The subsurface soil temperature decreases exponentially with depth and becomes insignificant at a depth of about 1m. The soil temperature may vary with depth depending on the season and time of the day. It also depends on the radiation, soil texture, moisture content, ground cover and surface weather conditions.

Air temperature and humidity:

The vertical profile of air temperature in the atmospheric boundary layer is influenced by the air mass above the PBL; thermal characteristics of the surface and subsurface medium, which influences the surface temperature; net radiation at the surface and its variation, which determines the radiative warming and cooling of the surface and PBL; sensible and latent heat fluxes; warm or cold air advection and height of PBL to which turbulent exchange of heat occurs. Similarly the factors influencing the specific humidity or mixing ratio of water vapour in the PBL are specific humidity of airmass above the PBL; type of surface, its temperature and availability of moisture for evaporation or evapotranspiration; variation of water vapour flux with height; advection of water vapour in the PBL; vertical motion and cloud formation and precipitation and the depth of PBL to which water vapour is mixed.

The vertical exchange of heat and moisture in the PBL is mainly by turbulent motions. The convergence or divergence of sensible heat flux causes warming or cooling of air. Similarly the convergence or divergence of moisture flux cause increase or decrease of specific humidity. The horizontal advection of heat and moisture is enhanced when there is a variation in surface characteristics or in air mass characteristics in the horizontal. Topographical features can also cause variation to temperature and humidity in the vertical and also to cloud formation and precipitation processes.

The variation of temperature and humidity with height in the PBL leads to density stratification. Thus an upward or downward moving parcel of air will find itself in different environment with different density. This difference in density may result in buoyancy force in the presence of gravity. This buoyancy force will accelerate the parcel when the environment is statically unstable. If the parcel is decelerated then the environment is statically stable and no buoyancy force is exerted on a parcel in neutral atmosphere. In moderate or extreme unstable conditions especially during midday or afternoon hours over land intense mixing takes place in the boundary layer so that conservable parameters such as θ and θ_v are distributed uniformly independent of height. This is the mixed layer. Specific humidity, momentum and concentrations of pollutants also have uniform distribution to some extent in this layer. The mixed layers occur more over tropical and subtropical oceans.

As a result of the diurnal variations of the net radiation and other surface fluxes air temperature and humidity show large diurnal variations over land and negligible variations over oceans. The presence of vegetation and evaporation reduces the diurnal range of temperature. It even reduces during the occurrence of cloud and strong winds and with height in the PBL and disappears at the maximum height of PBL. The diurnal variation of specific humidity depends on evapotranspiration, condensation, turbulence, mean winds, surface temperature and PBL height.

Wind:

The wind distribution in the PBL is influenced by so many factors. They are i) large scale pressure and temperature gradients in the PBL, ii) the diurnal heating and cooling, iii) surface roughness characteristics, iv) earth's rotation, v) large scale horizontal convergence and divergence, vi) PBL depth, vii) horizontal advection of momentum, heat and moisture exchange, viii) entrainment of air above PBL, ix) topographical features, x) presence of clouds etc. Within the PBL the actual winds are in geostrophic balance because of surface friction. The influence of the earth's surface on atmospheric winds exists through out the PBL. Hence the wind speeds decreases gradually and mean momentum is transferred downward when going towards the surface. Turbulence is the mechanism for the transfer of momentum and this momentum transfer in the vertical results in frictional force. At the surface the coriolis force vanishes and the frictional force balances with the pressure gradient force. Wind speed increase with height with little change in wind direction in the surface layer. Then the frictional force decreases and rotates anticyclonically with height. Near the top of PBL frictional force becomes small in magnitude and at the top the balance is between pressure gradient force and coriolis force. In a barotropic PBL where there is no temperature gradient wind veers with height. In the presence of thermal winds, wind may veer or back with increasing height according to frictional veering and veering or backing of geostrophic winds. The frictional veering can be taken as the difference between geostrophic wind veering and actual wind veering. The cross-isobaric angle between the actual wind and isobars decreases with height and vanishes at the top of PBL. The surface cross-isobaric angle is found to vary according to the surface roughness, stability, depth of PBL, latitude and baroclinicity of PBL. The angle is found to increase with increasing stability and roughness and decreasing latitude. In the absence of thermal winds the cross-isobaric angle is found to have a range between 0° - 45° and the range becomes wider in the presence of thermal winds which is from -20° to 70° (Arya 1988).

On clear days surface warms up giving rise to convective circulation, which transfer heat and momentum in vertical direction. The vertical heat transfer causes convective or buoyancy generated turbulence, which mixes the momentum thus reducing the wind shears in PBL. That is why the wind speed and wind direction remains uniform in the convective boundary layer. During nighttime the surface cools due to longwave radiation and surface inversion develops. Momentum exchange is reduced in the inversion layer hence wind shear and wind speed increase in this layer. Thermal stability of the PBL influences the PBL height, which in turns affects the wind. During morning hours the PBL depth increase as a result of surface heating and wind shear decreases and strong wind shear is found near the surface layer only. During evening the PBL height decreases and maximum wind shear is observed near the inversion layer. The magnitude of wind shear in the PBL during night is found to be larger than that during the day for a given wind speed at the top of the PBL. This is because of the difference in PBL height.

However the wind shear values very close to the earth's surface during daytime are larger than that at night. The vertical profiles of wind are having strong gradients in surface layer and uniform distribution in the convective boundary layer above the surface layer. In the presence of clouds considerable wind shear exist in the subcloud layer whereas the vertical potential temperature and humidity remains uniform throughout. The diurnal variation of wind from various observation shows that surface wind speed increases after sunrises reaches maximum around afternoon and then decreases whereas the wind above the surface layer decreases in the morning hours and increases in the evening hours reaching maximum in the nighttime.

Turbulence:

Turbulence is the chaotic nature of a flow, which is characterised by fluctuations of variables such as velocity, temperature etc. from their mean values in time and space. The atmospheric boundary layer is almost always turbulent. While turbulence is continuous in the surface layer it is intermittent in the upper part of PBL. Turbulence is responsible for the exchange of heat, momentum and moisture through the PBL especially the surface layer turbulence causes the exchange of these properties between atmosphere and the earth's surface. Turbulence can occur suddenly in a non turbulent flow through an initial breakdown of streamline flow at some localized regions. This breakdown can be either due to static instability or due to dynamic instability upon naturally occurring disturbances in the flow. Once turbulence is generated and fully developed there is no need of the instability mechanism for the maintenance of turbulence. In shear flows the shear will provide the mechanism to convert the mean flow energy to Turbulence Kinetic Energy (TKE). In the unstable stratified flow buoyancy force will convert potential energy into TKE. Thus the two mechanism for generation of turbulence are shear and buoyancy. The TKE equation can be written as

$$\frac{d}{dt}(TKE) = S + B + T - D \tag{1.8}$$

where S is the shear term, B the buoyancy term, T is the transport of energy and D is the energy dissipation term. In the daytime turbulence is generated by shear and buoyancy whereas during night the buoyancy term is negligibly small and sometimes negative buoyancy in the upper troposphere reduces the turbulence so that generation of turbulence is by shear term. In the surface layer since the velocity at the surface vanish and the air in the immediate vicinity flows through various surface inhomogeneties the surface shear is intense. Hence the turbulence generation by shear is always present in the surface layer. In the stably stratified flows turbulence is destructed by buoyancy and to maintain the turbulence Richardson (1920) proposed that in the absence of viscous dissipation of energy the rate of production of turbulence by shear must equal the rate of destruction by buoyancy. That is the ratio of turbulence production by buoyancy to shear production must be equal to or less than unity. This ratio is called the Richardson number. Theoretical and experimental studies found that if Richardson number is less than its critical value 0.25 the flow becomes turbulent. The observations on the transition from turbulent flow to laminar flow have indicated significant viscous dissipation of energy and much lower value of critical Richardson number which is in the range 0.2 to 0.5 (Arya 1988).

Turbulence is characterised by irregularity or randomness. The turbulent flow is three dimensional and highly variable in time and space. Diffusivity or ability to mix properties is the most important characteristic of turbulence. Dissipativeness is another important characteristic of turbulence. The turbulent kinetic energy is continuously dissipated by viscosity. The transfer of energy from mean flow to turbulence occurs in large eddies whereas in small eddies the turbulent kinetic energy is dissipated by viscosity. Therefore there is a continuous transfer of energy from the largest scale larger eddies to small scale small eddies.

International boundary layer experiments

Most of the field experiments done for investigating the boundary layer processes in the PBL are carried out either on land in the mid latitudes or on the marine boundary layer over tropical Atlantic and Pacific oceans. The boundary layer flux coefficients were obtained in the earlier days from computations using the theories and hypotheses proposed by pioneering works of Reynolds (1883), Taylor(1915), Prandtl (1924), Richardson (1920), Von Karman(1930) and Monin and Obukhov (1954). Large international field projects began during 1960s. The first large data sets of fluxes based on profiles of wind, temperature and humidity were obtained from the International Indian Ocean Expedition (IIOE) in 1964. In the year 1967 during July and August Wangara field experiment was conducted at Hay in Australia to study the land boundary layer feature. In the experiment the ground and soil data, temperature, humidity and wind profiles and radiation data were collected. The first major experiment to test the similarity theory of Monin and Obukhov(1954) was conducted in Kansas in 1968. This experiment known as Kansas experiment was conducted during the months of July and August. In the experiment the instruments were mounted on a 32m mast. The surface layer data, boundary layer fluxes, temperature and wind profiles and turbulent kinetic energy were obtained from this field experiment. The profile coefficients which are determined empirically using these observations (Businger et al, 1971) are used widely in the boundary layer studies. In 1969 from May to July intensive oceanographic and meteorological observation were carried out in the Caribbean sea at Barbados which is known as the Barbados Oceanographic and Meteorological Experiment (BOMEX). In the experiment almost all the boundary layer parameters including the air sea fluxes are measured. The Atlantic Trade Wind Experiment(ATEX) was conducted in 1969 to study the boundary layer in the trade wind region which flows towards the Intertropical convergence zone (ITCZ). Air sea fluxes were measured by direct eddy correlation method using hot wire and sonic anemometers as well as indirectly from profiles of temperature and humidity. Large number of field experiments were conducted in 1970s to understand the ABL processes. The Venezuelan International Meteorology and Hydrology Experiment (VIMHEX) was conducted during May-September 1972 at Carrisal , Venezuela to study the boundary layer characteristics with strong emphasis on the hydrological features. The Minnesota experiment carried out at Minnesota, USA during September 1973 employed a 32m mast to obtain the observation in the boundary layer. Emphasis was also given to the surface layer processes. Boundary layer fluxes, turbulent kinetic energy, wind and temperature profiles and radiation were obtained from the observations. The Cabauw/C during 1973-1984 and Cabauw/E during 1977-1979 were conducted at Cabauw tower, Netherlands to study boundary layer features of the region. GATE, the GARP Atlantic Tropical Experiment was the first large scale international field experiment of GARP conducted during 1974 to study the energetics and dynamics of cloud clusters that drift from the African continent over the Atlantic ocean to cause the ITCZ convection. The experiment was organised by the WMO and conducted by various nations. Each nation used their own method of estimating the boundary layer fluxes. Air-Sea fluxes were also measured using aircraft during the experiment. The combination of surface and aircraft data provided a better insight into the interaction between surface fluxes and convection in the boundary layer. The GATE observations also included the radiosonde observation and ocean mixed layer observation.

The Air Mass Transformation Experiment (AMTEX) was another large-scale experiment conducted during 1974-1975 in the East China Sea near Japan. Turbulent fluxes and other boundary layer parameters were measured. The boundary layer parameterisation scheme of Kondo(1975) was developed from the AMTEX observations. The Boundary Layer Structure 77 (BLS 77) during May-June 1977, Severe Environmental Storms and Mesoscale Experiment (SESAME) during April-June 1979 and Boundary Layer Experiment 83 (BLX 83) during May-June 83 are conducted at Oklahoma to study the boundary layer structure of the region during occurrence of severe storms. PHOENIX and PHOENIX II conducted at Boulder Colorado during September 1978 and May-July 1984 respectively are the experiments to measure the boundary layer parameters to study the boundary layer evolution of the region. ABLE the Amazon Boundary Layer Experiment conducted during July-August 1985 and April-May 1987 to study the special characteristics of the boundary layer of the Amazon forest is one of the field experiment carried out in the tropical land. The other important field experiments are MESO-GERS conducted at SW France in September 1984, Preliminary Regional Experiment of Storm central (PRE-STORM) at Oklahoma in 1985, Genesis of Atlantic Low Experiment (GALE) during 1986 at New England, USA and Atlantic, Frontal Air Sea Interaction Experiment (FASINEX) to evaluate processes in the subtropical convergence zone in Western North Atlantic in 1986, HAPEX the Hydrologic Atmospheric Pilot Experiment at SW France in 1986, the Humidity Exchange over Sea (HEXOS) in 1986 at North Sea, Netherlands, Taiwan Area Mesoscale Experiment (TAMEX) at Taiwan in 1987, FIFE, the First ISLSCP Field Experiment at Manhattan USA in 1987 etc. Apart from these many other regional flux experiments by smaller groups are also conducted to obtain a better picture of the boundary layer processes. A recent regional flux experiment was the CASP II 1992-93 (Canadian Atlantic Storms Prediction). This was in continuation of CASP conducted in 1985-86 to study the growth of winter storms and their interaction with upper ocean. The objective of CASP II was to study the mature stages of cyclogenesis in winter storms and their influence on ocean circulation and sea ice on Newfoundland Continental shelf and the Grand Banks.

Other important recent international marine boundary layer experiments are TOGA Coupled Ocean Atmospheric Experiment (COARE) and Marine Boundary Layer Project (MBLP). The existence and influence of warm pool in the western tropical Pacific on atmospheric convection was a result of COARE. Large quantity of air sea flux data over open ocean of both suppressed periods and active westerly period was made available by COARE. The MBLP was launched in 1994. The first experiment was conducted offshore of north of Holland, Denmark and it was called the Riso Air – Sea Experiment (RASEX). The instruments were deployed in a mast to measure bulk quantities of fluxes, aerosol and chemical concentrations. Oceanographic measurements are made by a University of Kiel ship. Second MBLP experiment was conducted in 1995 at west off Monterey California. Apart from the instrumentation used in RASEX, two ships, airplanes, lidar and radar measurements are also taken to have better resolution of data. After the First ISLSCP Field Experiment (FIFE) of 1987 ISLSCP has become a component of the Global Energy and Water Cycle Experiment (GEWEX) and conducted two field experiments; the Boreal Ecosystem – Atmospheric study (BOREAS) and the Large scale Biosphere Atmospheric Experiment in Amazonia (LBA). BOREAS is conducted during 94 –96 over a boreal forest area near Thompson, Manitoba. The experiment is conducted for a better data, which can be used for a good representation in model parameterisations of boreal forest in climatic models. LBA is an international experiment led by Brazil and is planned for the period 1999 – 2000. Climate and Hydrology components will be collected over various sites in the Amazonian Basin.

Boundary layer experiments over Indian region

The International Indian Ocean Experiment (IIOE) during 1963-65 which was the first field experiment to yield the first large boundary layer data was the first experiment that conducted over the Indian region. During IIOE observations are taken from surface and upper air observatories, ships, aircrafts, buoys and satellites. The boundary layer measurements were made by Bunker (1965) and provided some insight into the temperature and moisture stratification and transport of moisture and sensible heat in the region, particularly north of 10°N in Arabian Sea. In the year 1973 the Indo-Soviet Monsoon Experiment (ISMEX) was conducted over West Indian Ocean by four Russian research vessels from 19th May to 8th July. Radiosonde and Radar data were taken onboard ships at 6 to 12 hourly intervals. Very few observations are taken below 100m height. The ISMEX data provides vertical thermodynamic and kinematic structure of the boundary layer in the West Indian Ocean during Indian summer monsoon. MONEX (Monsoon Experiment) was designed as the part of the First GARP Global Experiment (FGGE) from 1st December 1978 to 30th November 1979 to cover the Arabian Sea, Bay of Bengal, Indian Ocean and Southeast Asia. Its two main component programs are a) the summer MONEX programme to study the summer monsoon over India and neighbourhood during 1st May to 31st July 1979 and b) the winter MONEX programme to study the winter Monsoon over Indonesia - Malaysia region during 1st December 1978 to 5th March 1979. The summer MONEX was conducted in two phases. The first phase covered the Arabian Sea and the second phase covered the Bay of Bengal. The experiment provided a better data set than that had ever been collected before in the Indian Monsoon region. The boundary layer in the large-scale monsoon air stream encounters strong interactions across the highly complex surfaces of the ocean and solid landmass over which it flows. The earlier experiments focussed on the marine ABL. To study the ABL over the Indian monsoon trough region wherein for four months from June to September every year organised moist convection prevails over the sub-continental scale, a large-scale field experiment was conducted. The experiment was called Monsoon Trough Boundary Layer Experiment (MONTBLEX) and was conducted from 1987 to 1990. A pilot experiment was carried out at Kharagpur during 1-7 July 1989. Slow and fast response tower instruments, minisonde and humidity instruments are used in the experiment. The full field experiment began from March 1990 and made use of four towers at Varanasi, Delhi, Jodphur and Calcutta, doppler sodar and monostatic sodar, tethered balloons, minisonde, surface and upper air observations, Aircraft flights, special cruise of Ocean Research vessel ORV Sagarkanya during August-September 1990 over north Bay of Bengal, INSAT pictures and weather radar observations in the IMD network. Observations were taken during all types of weather situations. The experiment collected vast amount of data over a vast region, which will continue to be analysed for a long time revealing various features of Indian monsoon.

A recent field experiment called the Bay of Bengal and Monsoon Experiment (BOBMEX) was conducted under the Indian climate Research Program (ICRP) in the Bay of Bengal in 1999. The aim was to study the air-sea interactions and intraseasonal oscillations in the Bay of Bengal during the summer monsoon using new types of measurements from ships and met-ocean buoys. A pilot experiment was conducted from 25th October to 11th November 1998. Phase I of the BOBMEX was carried out during July17 to 10th August 1999 and Phase II from August 13 to 31, 1999. This was a national program and used several advanced sensors and instruments onboard ship. The measurements made during the experiment are 1) direct air-sea fluxes using fast response data 2) surface meteorological parameters including radiation and rain, 3) upper air parameters using radiosondes and minisondes. 4) upper ocean temperature, salinity and current profiles and 5) turbidity in the upper ocean. The pilot experiment was carried out on board ORV Sagarkanya. The experiment arrangement consisted of sensors mounted on a micrometeorological tower fixed on a boom, raingauge placed on the upper deck of the ship, a GPS receiver, minisonde system signal conditioners, data logger and pentium based data acquisition and processing system. The full field experiment also used the same instruments and the data collected during the pilot experiment was used in this study.

Literature Survey

The atmospheric boundary layer characteristics are studied from theoretical and modelling considerations and through field observations and laboratory measurements. Parameterisation of sub-grid scale boundary layer processes has become necessary for large-scale weather forecasting models and for simulation of climate by general circulation models. This increased the importance of boundary layer studies. Most observational studies for investigating the boundary layer processes carried out on land are in the midlatitudes. The simulation of tropical circulation and convection are very sensitive to the parameterisation of boundary layer processes. Very few experiments are carried out to study the structure and characteristics of boundary layer over the land locked tropics especially the monsoon boundary layer over the Indian monsoon region. The monsoon flow in its course over India passes over areas of torrential rain and over mountain complexes where its properties are modified due to planetary, regional and local weather systems and the varying topographical features.

Earlier monsoon experiments (IIOE, ISMEX-73, Monsoon –77, Monsoon-88 and MONEX-79) focused on the study of Arabian Sea marine atmospheric boundary layer. The monsoon boundary layer over Arabian Sea has been critically reviewed by Young (1987) and extensively studied using the data from the field experiments by Colon (1964), Sikka and Mathur (1965), Pant (1978, 1982) etc. These studies revealed the existence of low level atmospheric inversions over the west and central Arabian Sea, which disappeared near the West Coast of India. Ramanathan (1978) found that the buoyancy effects are permanent in May compared to June when the mechanical processes are significant over Arabian Sea, using ISMEX data. During active rainfall period the inversion disappears and has saturated mixed layers, positive sea-air virtual temperature difference and larger specific humidity. However these studies cannot provide information on the structure of the boundary layer over Indian subcontinent. MONTBLEX-90 was the first large-scale field experiment conducted over the Indian land area to study the boundary layer characteristics during monsoon with special reference to the monsoon trough region. The Indian summer monsoon trough region is having organised moist convection for four months from June to September every year. Strong westerly to southwesterly wind flow to the south of the trough and weak easterly to northeasterly wind flow to its north. Also there are major differences on the dynamical and convective characteristics of the eastern and western points of the trough. The western end is a heat low with shallow ascent and convergence up to lowest half kilometer whereas eastern end is having convergence up to mid troposphere and moist ascent throughout the troposphere with intense convection. The eastern end lies on the warm waters of Bay of Bengal, which is a heat and moisture source (Sikka and Narasimha 1995). Therefore the atmospheric boundary layer over the region requires special study and MONTBLEX-90 was conducted. Extensive literature is available on the monsoon trough boundary layer as a result of the field experiment.

Sivaramakrishnan et al (1992) studied the characteristics of turbulent fluxes of sensible heat and momentum in the surface boundary layer using sonic anemometer data at 8m from a 30m micrometeorological tower at Indian Institute of Technology, Kharagpur (22.3°N, 87.2°E), India during MONTBLEX-90. Heat fluxes show a diurnal trend whereas the momentum flux shows variability but no particular trend. The daytime atmospheric condition during monsoon season was unstable. Near-neutral condition occurs during evening hours and sometimes at night stable conditions with downward transport of heat flux and reduced momentum flux. Narahari Rao (1995) estimated the surface temperature at different stations using the temperature data at three levels below the surface collected during MONTBLEX-90 with the one dimensional heat conduction equation for the soil. Temperature of the soil varies with depth and the variation is sinusoidal with a period of a day. The equation gives amplitude and phase variation of temperature wave with depth. Using kytoon and doppler sodar data at Kharagpur during MONTBLEX Verneker et al (1995) studied the turbulent characteristics of the boundary layer and found that the neutral boundary layer height decreases in the early morning hours and then increases followed by a decrease and a routine increase. A decrease in turbulence intensities and turbulent energy dissipation at inversions is noticed. Pradhan et al (1996) found that the synoptic features and surface parameters and the interrelation of low cloud formation with the boundary layer are having close relations. Whenever low clouds penetrate inversion height sensible heat flux increased. If there is a convergence in a synoptic or sub-synoptic scale, momentum flux rises because of high wind shear. The inversion height is modulated by the convergence due to synoptic or sub-synoptic disturbances. Therefore it was concluded that the synoptic and sub-synoptic scale convergence modulates the inversion height and the cloud base within the inversion height which in turn modulates the sensible heat and momentum fluxes. Chatterjee et al (1996) attempted to study the turbulent structure of the atmospheric boundary layer at Kharagpur diagnostically using a first order non local closure model based on turbulence theory and a second order based closure model as part of the MONTBLEX program. Distinct zones where TKE production exceeds consumption which supports the turbulent activity was noticed by the first model. The early morning boundary layers are characterised by downward turbulent transport over upward transport while the reverse is true for afternoon case. The second model also provided evidence of distinct zones of TKE production. Kusuma et al (1996a) studied the sensitivity of the mean and the turbulence structure of the monsoon trough boundary layer to the choice of constants in the dissipation equation using a one-dimensional model with e- ε turbulence closure. The performance of e-ε closure model in simulating the mean structure of the boundary layer is estimated by comparing with e-l model simulations and observations. The study suggested that the modifications of dissipation equation by approximating the TKE generation by maximum shear production or shear and buoyancy production made usually could be avoided by suitable choice of constants, which are suggested in their study. With the suggested constants the observed turbulence structure is better simulated. The turbulence structure simulation with e-ε model is better than e-l model simulations. Temperature profiles during neutral and unstable conditions are better simulated but not the stable profile. Moisture profiles are also simulated well using the models but the wind profiles simulation is more complicated.

Gera et al (1996) investigated the boundary layer dynamics at the western end of monsoon trough using sodar and tower data as part of MONTBLEX-90. Stable layers are noticed in association with monsoon. Morning inversion layer disappeared and a rising layer with growing thermal plumes is rarely noticed during winter. Kusuma (1996) estimated the roughness lengths and drag coefficients at Jodhpur and Kharagpur using MONTBLEX-90 mean velocity data. The roughness at Jodhpur, which is not having uniform surface, is found to be 1.23cm where wind is between 200° and 230° and at Kharagpur it is 1.94cm. The drag coefficient at Jodhpur varied with roughness and wind. It increased with wind speed decrease according to a power law $C_D=0.05U_{10}^{-1.09}$ in the range $0.5 < U_{10} < 7ms^{-1}$. At Kharagpur the drag coefficient is found to be smaller than at Jodhpur by 50% for same range of wind speeds.

Sadani and Murthy (1996) computed the surface heat flux during convective activity in the morning hours using acoustic sounder derived vertical velocity variance and inversion height. Surface heat fluxes computed are of same magnitude as obtained by tower measurements. The height of the boundary layer obtained during early morning hours of active monsoon are higher than during pre-monsoon and onset phase though the heat flux decreased from May to August. Kusuma et al (1996) found that at Jodhpur for parameterisation of sensible heat flux and momentum flux using bulk aerodynamic formula the dependence of exchange coefficients on wind speed must be taken into account because the coefficients increases rapidly with low winds. The sensible heat flux is found to be proportional to the temperature difference between 1 and 30m and the momentum flux is linearly dependent on wind speed. Rajkumar et al (1996) investigated the thermal and wind structure of the monsoon trough boundary layer and found that during wet conditions the atmosphere was predominantly stable during morning and evening whereas during afternoon stable layer exists underneath a neutral layer. During a dry situation the atmosphere is stable during morning and during afternoon and evening neutral or unstable conditions prevail. It was concluded that the monsoon boundary layer is similar to nocturnal boundary layer rather than to well mixed layer especially when the ground is wet and sky is overcast.

Pradhan et al (1994) obtained surface sensible heat flux using a new methodology having a concept of isolated layers. Using the heat flux obtained its interplay with synoptic and mesoscale features are examined. High downward flux is noticed in relation with strong stability of the atmosphere after an intense thunderstorm. But one to one relation between downward heat flux and thunderstorm can not always be established. Viswanadham and Satyanarayana (1996) computed the surface layer parameters during different phases of monsoon over Varanasi and found that the sensible heat flux shows little diurnal variation when the monsoon is active and the values are low. The flux is more during moderate or weak monsoon. The turbulence intensity parameters do not vary very much with monsoon activity. But the stability parameter indicates that during active phases of monsoon the atmosphere tends to become less unstable from highly unstable conditions or from stable to less stable condition. Potty et al (1996) studied the PBL characteristics over the monsoon trough using a numerical experiment with improved boundary layer physics. The PBL characteristics of the western sector and eastern sector of the trough was studied and compared. The results show maximum turbulence in the western sector than eastern sector.

Raman et al (1990) obtained the boundary layer structure and turbulence during the pre-monsoon period. Monsoon boundary layer height was smaller with super adiabatic lapse rate, well-defined mixed layer and sharply defined elevated inversion. The sensible heat fluxes are more in pre-monsoon boundary layer. Parasnis and Goyal *(1990) studied the variations in thermodynamic characteristics during contrasting* synoptic weather conditions. Stable stratification was observed up to higher heights during break monsoon period with an inversion. More convective instability was observed during break period. The zonal transport of moisture is more in the active period in the lowest layers. Parasnis (1991) studied the convective boundary layer during active and break condition of summer monsoon. It was found that during the break conditions the cloud layer is marked with subsaturation and inversion was observed. In the active period the upward buoyancy exists to higher levels even up to 600hPa level.

A few studies are carried out using the observations taken over the Bay of Bengal region during MONTBLEX-90. Mohanty and Mohankumar (1990) examined the link between air-sea interface fluxes of heat and moisture and monsoon activity over the Indian subcontinent. It was found that during the active convective period of summer monsoon there is a net oceanic heat loss, which produces a positive feed back for the maintenance of deep convection above the marine boundary layer. The time variation of surface fluxes of heat, moisture and momentum over a sea station in North Bay of Bengal during the passage of a depression was studied by Sivaramakrishnan et al (1996). Marked synoptic and diurnal variations in fluxes are noticed during the depression. Surface fluxes are high during depression. It was found that the flux calculation by the assumption $C_D=C_H=C_E$ normally used was not appropriate in the disturbed weather condition. Singh (1992) studied the relation of the surface fluxes to low pressure systems over Bay of Bengal. It was found that the fluxes increase steeply during the development and intensification of a system. It decreases after the system moves away. Murthy et al (1996) studied the variability of the ocean surface boundary layer characteristics on daily time-scale in the northern Bay of Bengal. It was noticed that active air-sea interaction with positive feed back from sea to air prevailed during the development of deep depression. The heat loss from the surface layer is utilised to enhance the latent heat flux. Lateral advective processes affecting the ocean boundary layer characteristics by bringing in heat and removing it occurs due to the presence and movement of warm core eddies during the southwest monsoon.

CHAPTER 2

Land boundary layer characteristics

The thermal and dynamic interaction between the atmosphere and the underlying surface occurs through turbulent exchange of momentum, heat and moisture at their interface. The variation of fluxes of momentum heat and moisture in the surface boundary layer and their distribution in the rest of the planetary boundary layer play a vital role in the energy transport mechanism of the ocean-atmosphere land system. Over the Indian region the meteorological features exhibits wide variability during pre-monsoon, onset and post-monsoon periods. Also there is a region to region variation due to the topography of the region. Therefore the boundary layer characteristics over the region will also have wide variability spatially and during various seasons. Thus the boundary layer over the region during various periods requires special study. Several studies were done over the region giving special reference to the monsoon trough region using the MONTBLEX-90 data. All those studies bring out the variation in the surface layer parameters, structure of boundary layer etc., during the southwest monsoon period. The variations that occur in the boundary layer characteristics during various seasons were not examined well. It was also well known that the synoptic and mesoscale features and the atmospheric boundary layer have close linkage. Mishra and Salvekar (1980) produced unstable disturbances by explicit high-resolution integration of the boundary layer equations within the framework of quasi-geostrophic baroclinic theory. Shukla (1978) found CISK as an important mechanism for the instability during monsoon. These studies show the importance of boundary layer on the formation of disturbances. Only a very few studies are done so far to understand the variations in the characteristics in the boundary layer during synoptic scale or mesoscale disturbances (Pradhan et al 1994; Pradhan et al 1996; Singh 1992; Sivaramakrishnan et al 1996 etc.). These studies found that the surface flux increases steeply during development and intensification of the system and it decreases as it decays or passes away. The boundary layer variations that occur at different stations when a depression or cyclone is formed either over Bay of Bengal or over Arabian Sea was not investigated. All the earlier boundary layer studies *associated with the cyclonic system were done as the system passes through the area of* study. According to Klinker (1997) in the tropics convection is an important process and the sensible heat flux shows a high correlation to convective rainfall. Hence deep convection may influence the surface fluxes. The interplay between thunderstorm and the boundary layer over the Indian region is not studied well by the earlier studies. Therefore in this chapter the variations in the boundary layer fluxes and stability during the pre-monsoon, onset and post-monsoon seasons at some selected stations over the subcontinent are examined. Also the variations in association with the formation of thunderstorms are also obtained. To understand the variations in boundary layer characteristics at different stations over the subcontinent during the formation of depressions or cyclones either over Bay of Bengal or over Arabian Sea the surface fluxes and stability parameter are computed at various stations during the occurrence of a monsoon depression.

Materials and Methods

The daily values of surface momentum, sensible heat and latent heat fluxes and stability parameter during April, May, June and October, November, December are obtained indirectly by the profile method for five years from 1984 to 1988 for three west coast stations Bombay, Mangalore and Trivandrum. Computations are performed using the daily surface observations of wind and temperature taken at 00UTC by the India Meteorological Department. The thunderstorm data and the associated rainfall in millimeters during the pre-monsoon and post-monsoon at Trivandrum are taken from the Indian Daily Weather Report to study the variations in surface layer characteristics during thunderstorms. The variations in surface layer features in association with the depression formed during August 2- 6 over the Bay of Bengal at some selected stations are also obtained. The stations selected are (fig. (2.1)) Calcutta (CAL), Masulipatnam (MPT), Madras (MDS), Amini (AMN), Minicoy (MNC), Mangalore (MGL), Goa (GOA) and Bombay (BMB).

The surface fluxes of momentum can be obtained using the profile method which is an indirect method based mainly on Monin- Obukhov similarity theory. The wind profile, temperature profile and specific humidity profile in the surface layer can be described as (Dyer and Hicks 1970; Businger et al 1971)

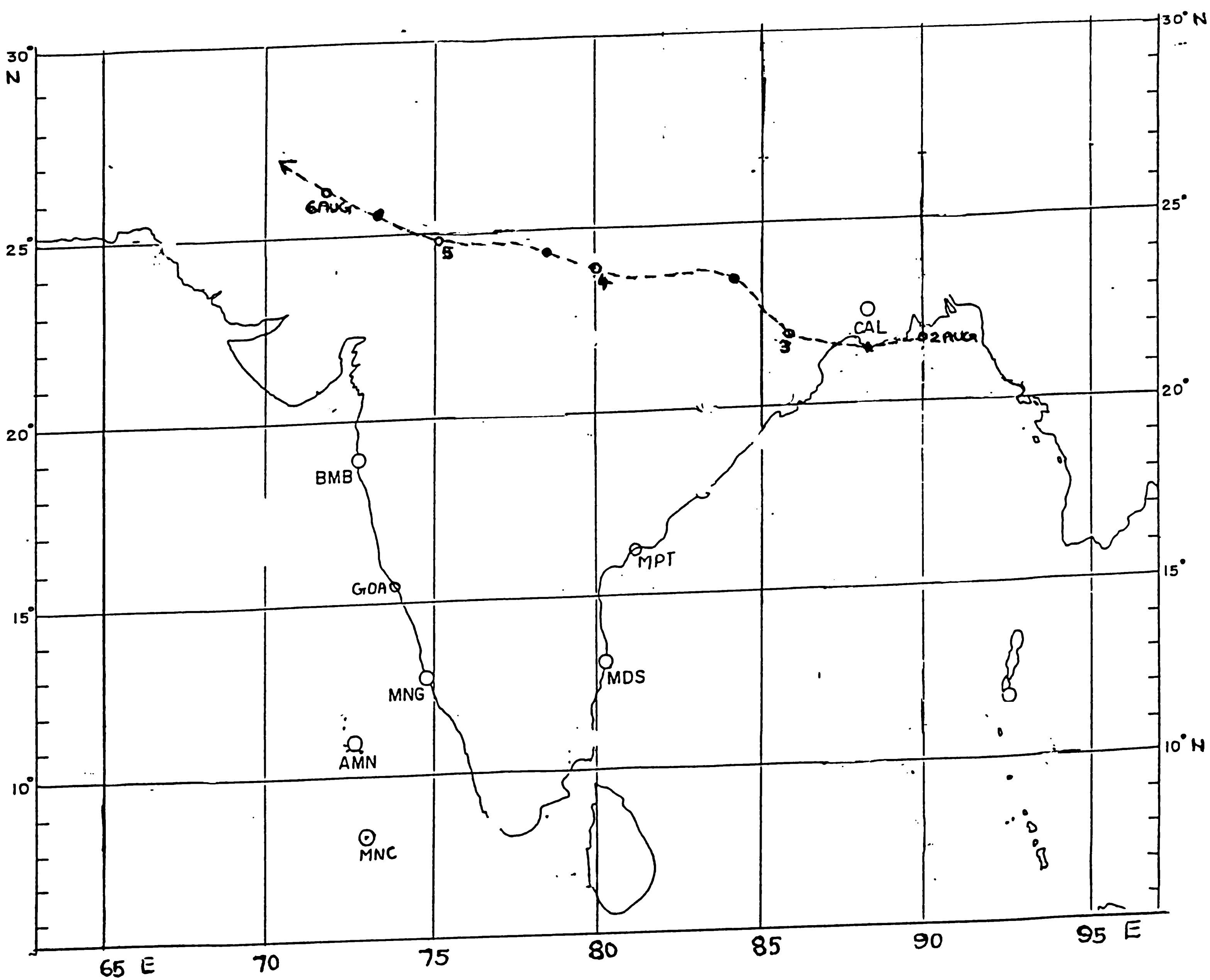

Fig. 2.1 Track of Depression

$$\Delta \bar{u} = (u_* / k)(\ln z_2/z_1 - \psi_m(\zeta)) \qquad (2.1)$$

$$\Delta \bar{\theta} = (R\theta_* / k)(\ln z_2/z_1 - \psi_h(\zeta)) \qquad (2.2)$$

$$\Delta \bar{q} = (Rq_* / k)(\ln z_2/z_1 - \psi_h(\zeta)) \qquad (2.3)$$

where $\Delta u = u_2 - u_1$, u_1 and u_2 are winds at levels 1 and 2 respectively, $\Delta\theta = \theta_2 - \theta_1$ is the difference in temperature at levels 1 and 2 respectively and $\Delta q = q_2 - q_1$ is the difference in specific humidity at levels 1 and 2 respectively. u_* is the frictional velocity, θ_* is the temperature scale and q_* is the humidity scale, ψ_m and ψ_h are the stability functions associated with wind and temperature respectively. $\zeta = z/L$ is the stability parameter, where $z = z_2 - z_1$ is the difference in height of the two levels 1 and 2. z_2 is taken at 10m height which is the height at which surface winds are measured and z_1 is the roughness length z_o where u is zero. The roughness length is obtained following Delsol et al (1971) using the expression

$$z_o = 0.15 + 0.2(236.8 + 18.42h)^2 10^{-8} \qquad (2.4)$$

The temperature at the earth's surface is obtained by solving the energy balance equation and that at 10m is obtained by linearly interpolating the observed surface temperature to that level.

The surface energy balance can be expressed as

$$G_o = (1 - \alpha) S_W\downarrow + L_W\downarrow - \sigma T_s - F_s\uparrow - F_l\uparrow \qquad (2.5)$$

where $S_W\downarrow$ is incoming short wave radiation, $L_W\downarrow$ is longwave radiation flux into the ground, α the surface albedo which is taken as 0.15, σT_s is the outgoing longwave radiation where σ is the Stefan-Boltzman constant. F_s and F_l are the fluxes of sensible and latent heat respectively which are obtained using stability dependent bulk aerodynamic formula and G_o is the soil moisture flux. The surface temperature is obtained by solving the equation using the software by Krishnamurty (1986). The first guess for T_s is assumed to be the air temperature at the lowest observation level. The data used for the computation of ground temperature is the upper air radiosonde data at

00UTC obtained from the India Meteorological Department. The saturated specific humidity at the surface is taken as the specific humidity at the surface. The computed ground temperature is compared with the NCEP reanalysis ground temperature data and both are found to be in good agreement.

L is the Monin Obukhov length given by

$$L = \bar{T} u_*^2 / g k \theta_* \quad (2.6)$$

R=0.74 is the ratio of eddy diffusivities in the neutral limit, k=0.4, the Von Karman constant.

The stability functions can be written as (Paulsen 1970, Barker and Baxter 1975)
In the unstable condition ($\zeta<0$)

$$\psi_m(\zeta) = \ln((1+x^2)/2) + 2\ln((1+x)/2) - 2\arctan x + \pi/2 \quad (2.7)$$

and

$$\psi_h(\zeta) = 2\ln((1+y)/2) \quad (2.8)$$

where $x = (1 - 15\zeta)^{1/4}$ and $y = (1 - 9\zeta)^{1/2}$ (2.9)

In stable condition ($\zeta > 0$)

$$\psi_m = -4.7\zeta \quad (2.10)$$

and

$$\psi_h = -4.7\zeta / R \quad (2.11)$$

The frictional velocity u_*, the temperature scale θ_* and humidity scale q_* are computed iteratively. First a large value was assumed for L and u_* and θ_* are computed in neutral limits. L is then recomputed using the computed u_* and θ_* according to equation (2.6). Then u_* and θ_* are recomputed using equation (2.1) and (2.2) with the new L. Again L is recomputed with the recomputed u_* and θ_* and this process is repeated until L does not change in desired accuracy limits.

The surface momentum flux is computed as

$$\tau = \rho u_*^2 \quad (2.12)$$

Surface sensible heat flux is given by

$$H = -\rho C_p u_* \theta_* \quad (2.13)$$

And surface latent heat flux can be expressed as

$$E = -\rho L_e u_* q_* \quad (2.14)$$

where ρ is the density of air which does not have much variation in the layer considered and is taken as 1.25×10^3 kgm^{-3}, C_P is the coefficient of specific heat capacity at constant pressure and L_e is the latent heat of evaporation/condensation.

Results and discussion

Variation in surface fluxes during pre-monsoon, onset and post-monsoon periods

a) Momentum flux

The surface momentum flux during April, May and June for five years from 1984-1988 at Bombay, Mangalore and Trivandrum are given in figures (2.1a-2.1c) respectively. The momentum flux is much less during April and May when compared to that during June at Bombay. During the pre-monsoon period the flux remains almost constant throughout the period with occasional rise in value. But as the monsoon is reached the surface momentum flux increases and the values reach $0.2Nm^{-2}$ and occasionally more. At Mangalore the momentum flux is less than $0.2Nm^{-2}$ during most of the days in April and early May. Occasional increase in flux is noticed during the period. But by late May and June the flux increases and is about $0.2Nm^{-2}$ or more during

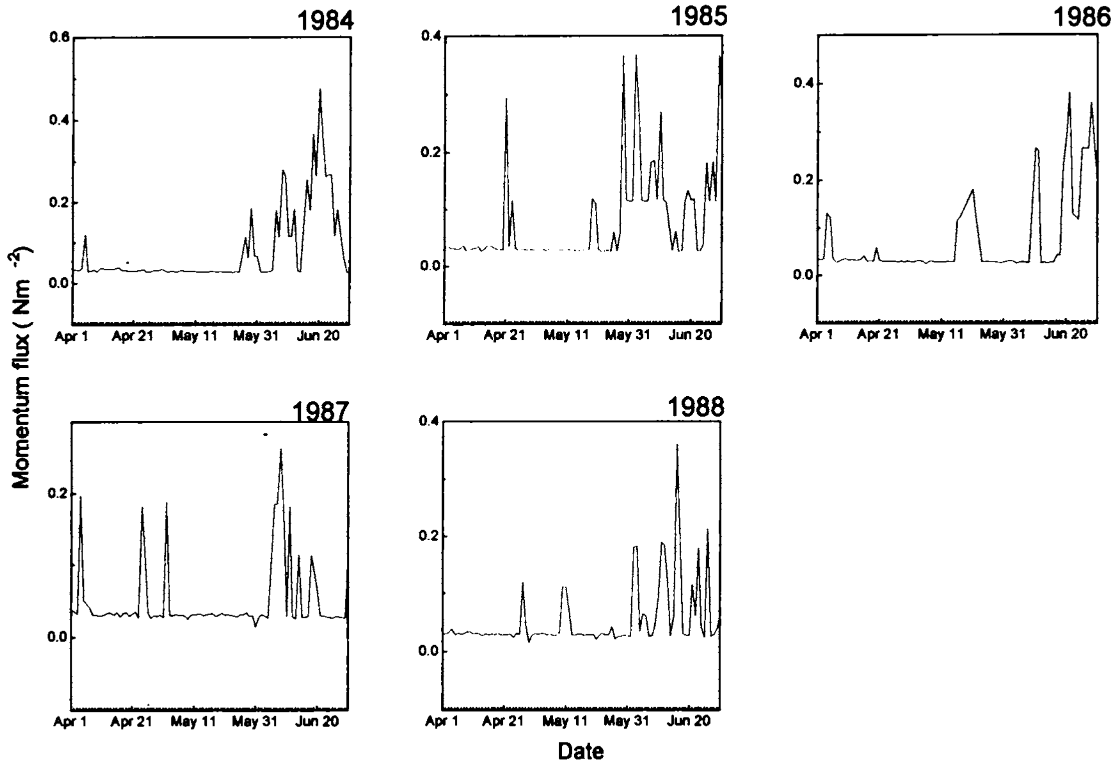

Fig. 2.1a Momentum flux during April, May and June 1984 - 1988 at Bombay

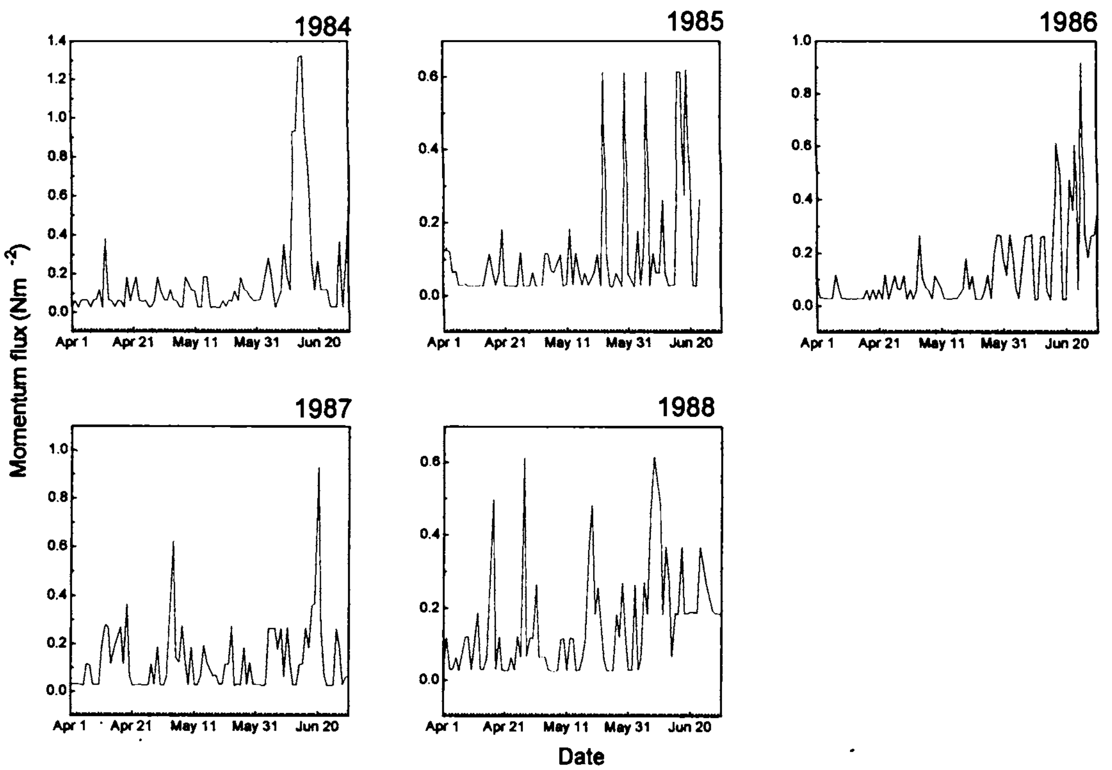

Fig. 2.1b Momentum flux during April, May and June 1984 -1988 at Mangalore

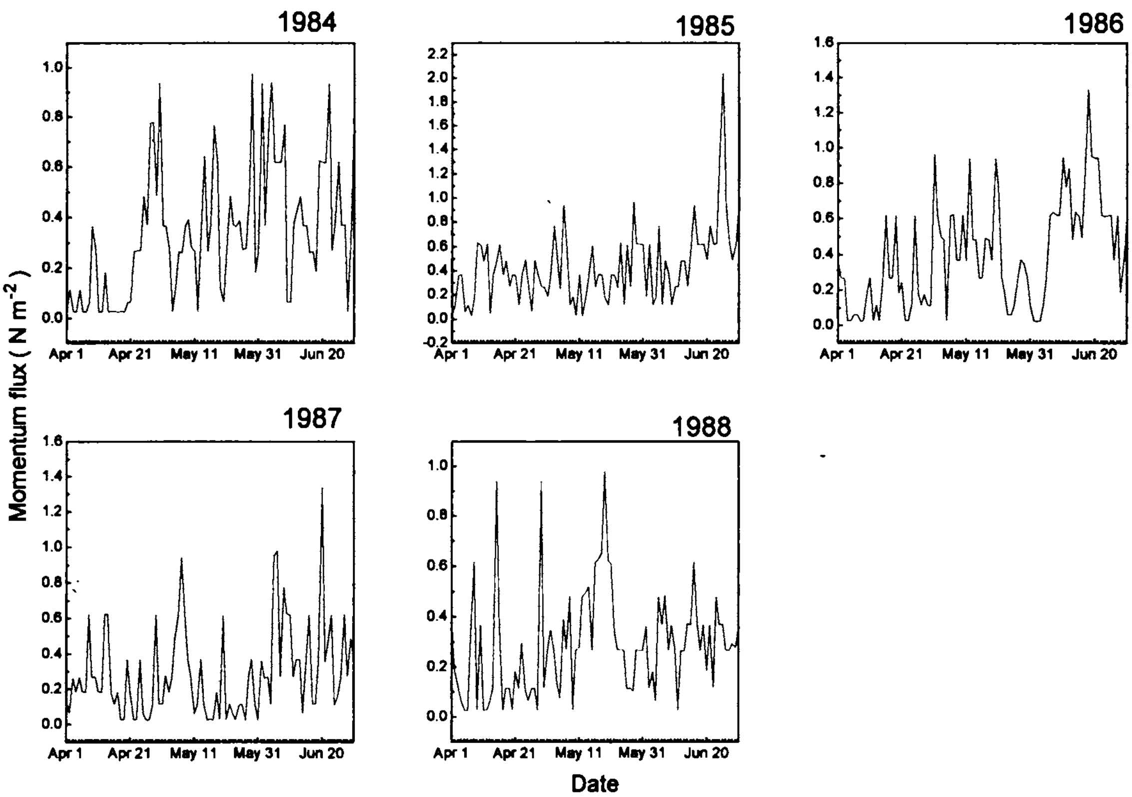

Fig. 2.1c Momentum flux during April, May and June 1984 - 1988 at Trivandrum

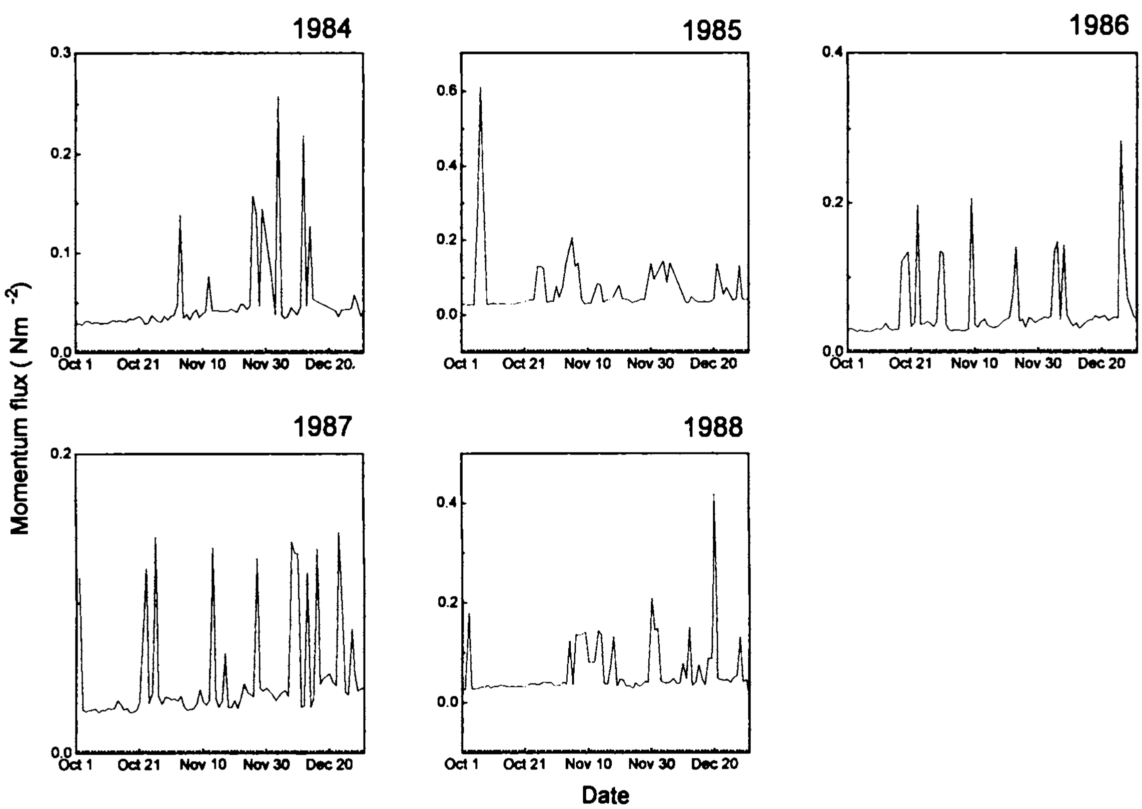

Fig. 2.2a Momentum flux during Oct.-Dec. 1984 - 1988 at Bombay

most of the days. At Trivandrum high flux is noticed in most of the days much before the onset of monsoon and very high flux transport occurs after the monsoon is reached over the station in June. Comparing the magnitudes of the surface momentum flux at the three stations, high flux values are noticed at Trivandrum. Mangalore is having higher fluxes when compared to that at Bombay. Therefore as the monsoon flow approaches the station there is an increase in surface momentum flux. The increase is due to the increase in surface winds in association with the monsoon onset. The observed wind speed at Trivandrum is higher than the other two stations hence higher fluxes are noticed at the station.

Surface momentum flux during October, November and December for five years from 1984-1988 at Bombay, Mangalore and Trivandrum are shown in figures (2.2a-2.2c) respectively. The surface flux was much smaller during October and it increased in most of the days during November and December at Bombay. The increase was in association with the increase in the northeastern wind speed. The maximum flux values are found to be less than $0.2Nm^{-2}$ mostly. At Mangalore the surface flux remained less than $0.2Nm^{-2}$ in October which increased during November and December and reached even up to $0.6Nm^{-2}$ or more occasionally. At Trivandrum higher surface momentum flux was noticed during October. High fluxes are also noticed during certain days in November. The momentum flux in general is having a decreasing trend from October to December at Trivandrum. High flux during October is due to the high wind speeds in association with the withdrawal of southwest monsoon from the subcontinent. Also during the post-monsoon period cyclonic systems will form over Bay of Bengal or Arabian Sea which causes an increase in wind speed. Therefore surface momentum flux will be higher during the occurrence of these systems.

Comparing the flux values during the various periods higher values are noticed during the pre-monsoon and onset periods than the post-monsoon period at all the three stations. This is due to higher wind speed during the pre-monsoon and onset than during post-monsoon. High winds blows across Arabian Sea into the subcontinent much before the onset of monsoon. Because of this high winds higher fluxes are noticed during the period. The northeasterly winds are not much stronger as the monsoon winds at the three west-coast stations considered.

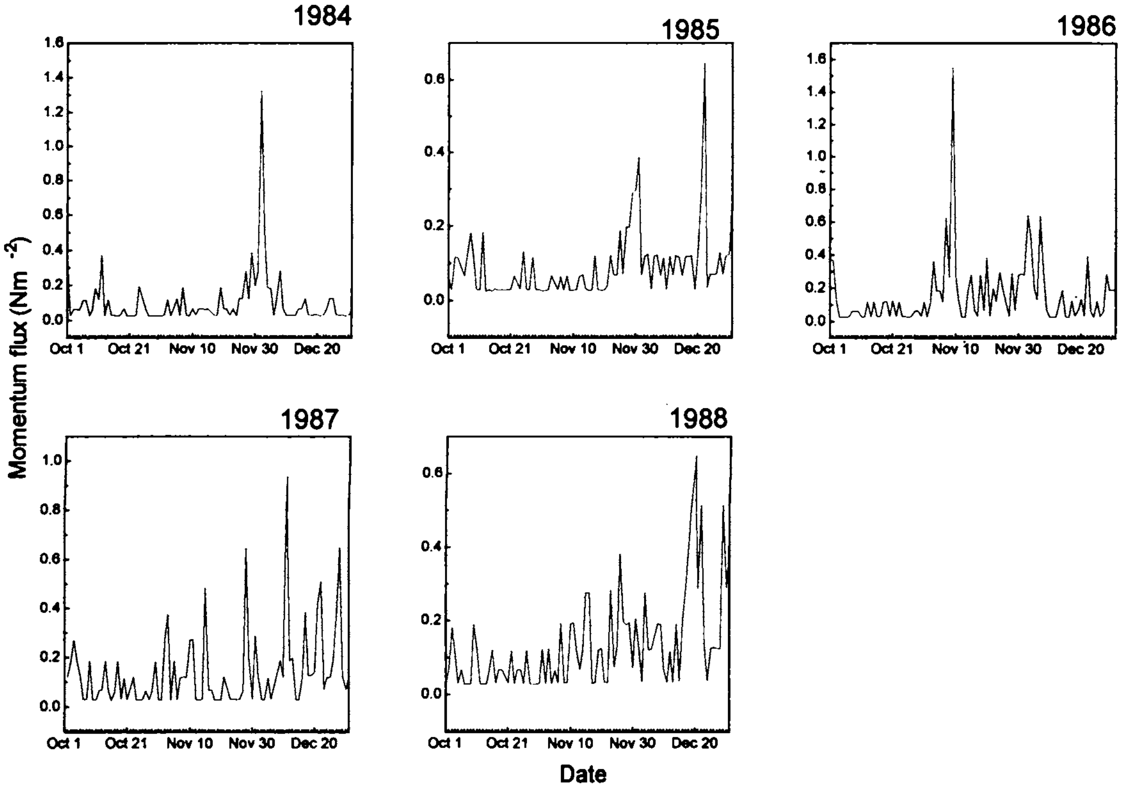

Fig. 2.2b Momentum flux during Oct.- Dec. 1984 -1988 at Mangalore

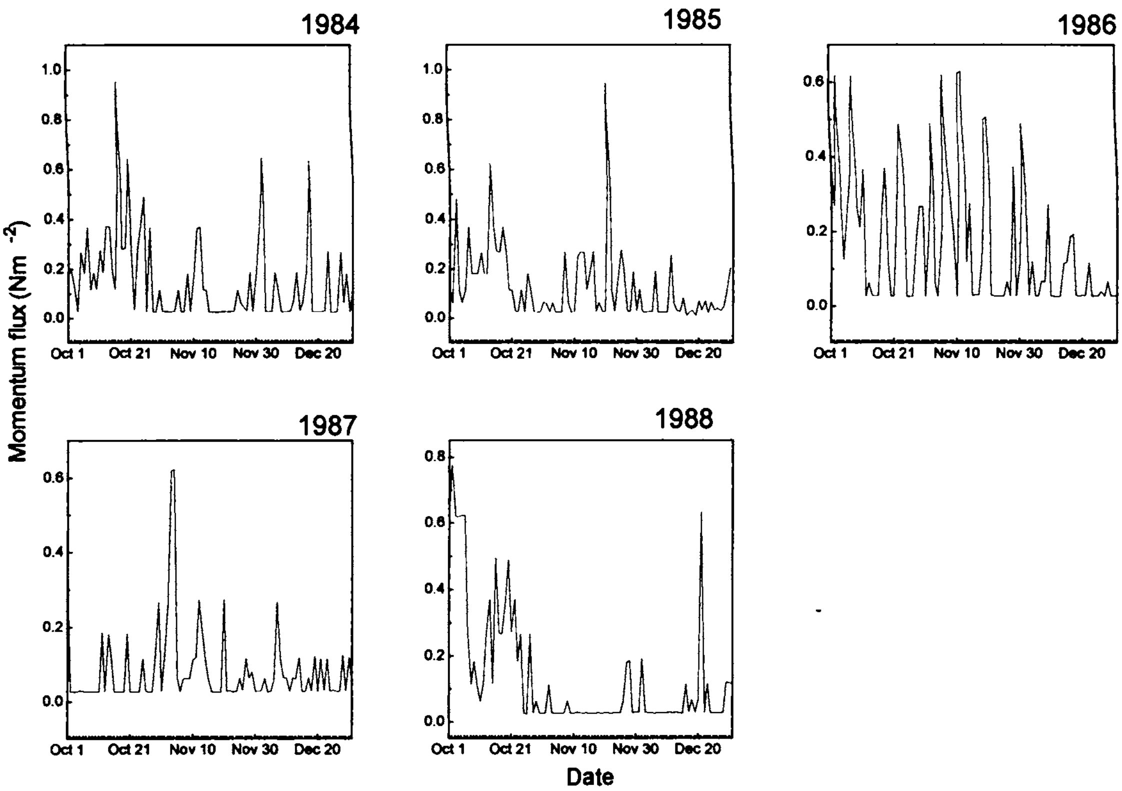

Fig. 2.2c Momentum flux during Oct.- Dec.1984 - 1988 at Trivandrum

b) Sensible heat flux

The time series of surface sensible heat flux during April, May and June from 1984-1988 at Bombay, Mangalore and Trivandrum are given in figures (2.3a to 2.3c) respectively. At Bombay very high sensible heat flux was noticed during April which decreased by May. The fluxes reached even more than $300Wm^{-2}$ in certain days in April. In May it was less than $200Wm^{-2}$. It then increased during the onset period. High flux during April is due to the high temperature in the surface layer whereas the increase in June is in association with the increase in wind speed during the onset. At Mangalore the sensible heat flux remains between $100\text{-}200Wm^{-2}$ during the pre-monsoon period with occasional rise to about $300Wm^{-2}$. The flux increased by the onset and is found to be greater than $200Wm^{-2}$ in most of the days. Very high sensible heat flux occurs in certain days during pre-monsoon period and onset period at Trivandrum. It even reached up to 300 Wm^{-2} or more. High flux during the pre-monsoon is due to the high temperature during the period whereas at the time of onset the high wind speed contributes towards the increase in flux.

Figures (2.4a to 2.4c) show the time series of sensible heat flux during October, November and December from1984-1988 at Bombay, Mangalore and Trivandrum respectively. The sensible heat flux was much less during October at Bombay and it increased gradually and very high fluxes are noticed during December. It was found to reach even up to $600Wm^{-2}$ or more. At Mangalore also the sensible heat flux gradually increased from October to December. During October the flux was less than $200Wm^{-2}$ which gradually increased in November and December. Higher flux values of about $400Wm^{-2}$ are noticed in certain days. At Trivandrum the sensible heat flux is found to be between $100\text{–}200Wm^{-2}$ most of the days throughout the period. In certain days the flux increased and reached even more than $300Wm^{-2}$. Higher sensible heat flux is noticed at Bombay when compared to the fluxes at Mangalore and Trivandrum. The lesser values during October may be due to the radiative cooling of the surface layer because of the overcast skies during the withdrawal period. After that the fluxes increase due to increase in temperature in surface layer during the period. The flux values are found to be very high but Pradhan et al (1994) has reported high sensible heat flux greater $350Wm^{-2}$ at Kharagpur during MONTBLEX-90.

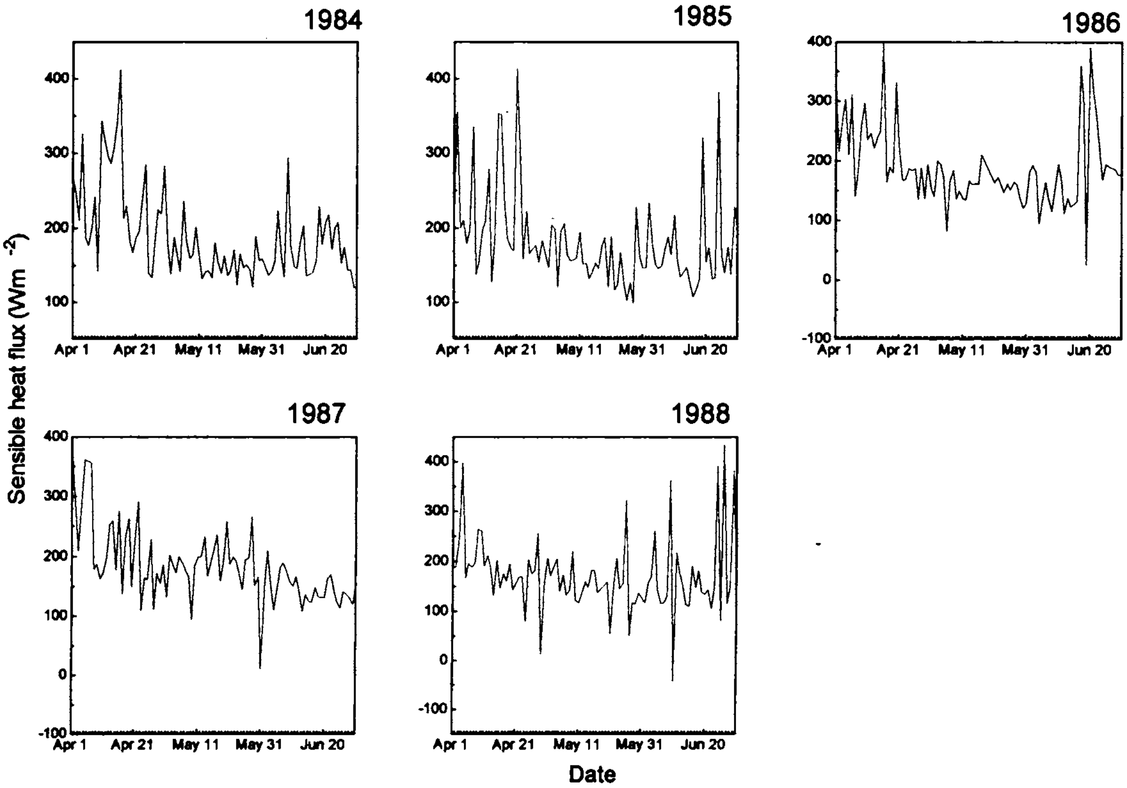

Fig. 2.3a Sensible heat flux during April, May and June 1984 - 1988 at Bombay

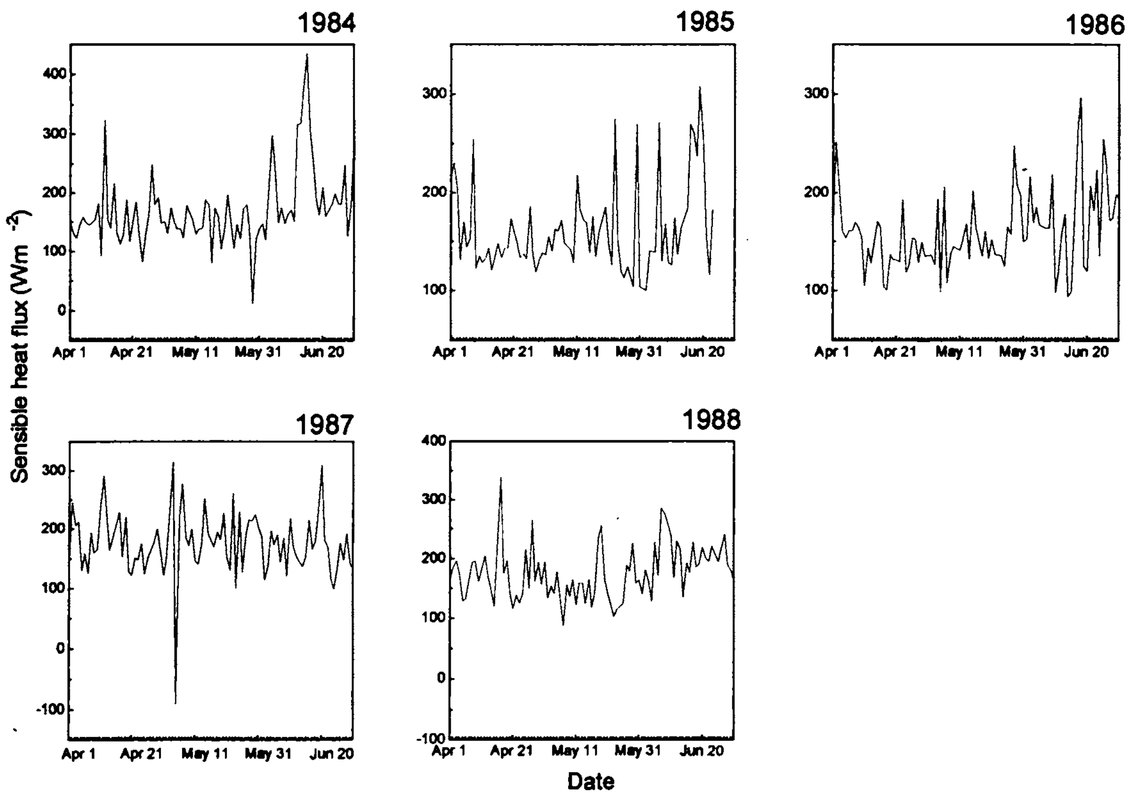

Fig. 2.3b Sensible heat flux during April, May and June 1984 -1988 at Mangalore

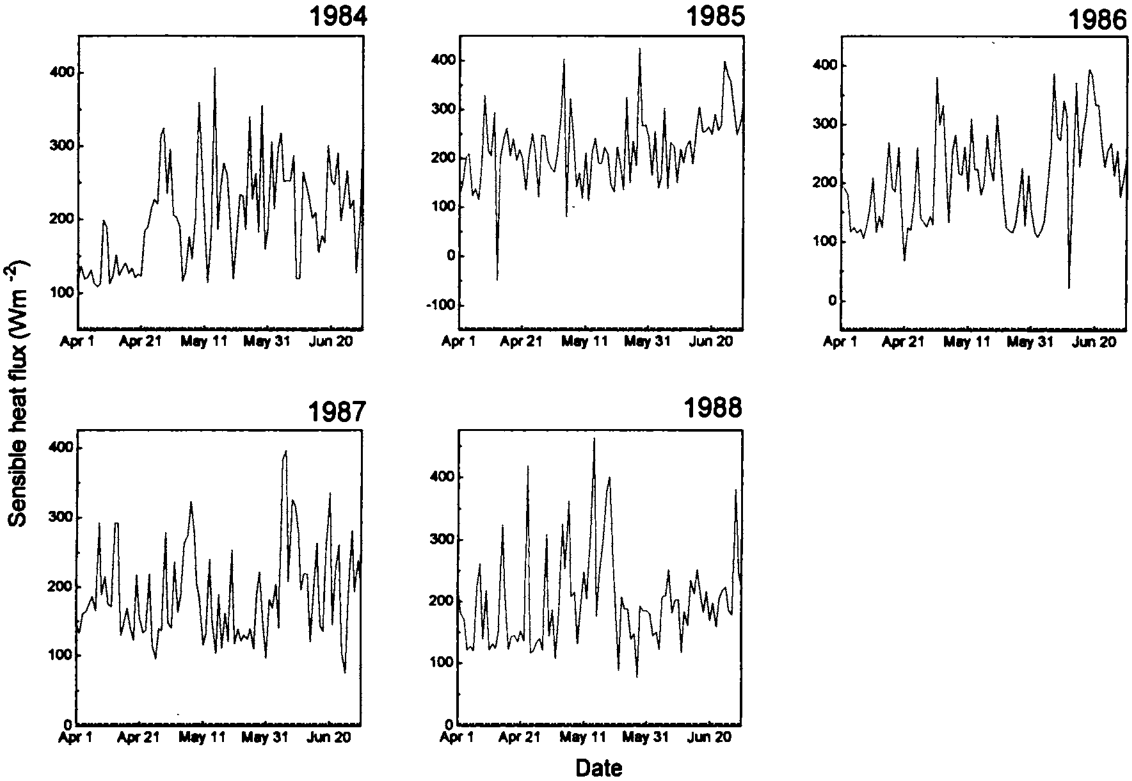

Fig. 2.3c Sensible heat flux during April, May and June 1984 -1988 at Trivandrum

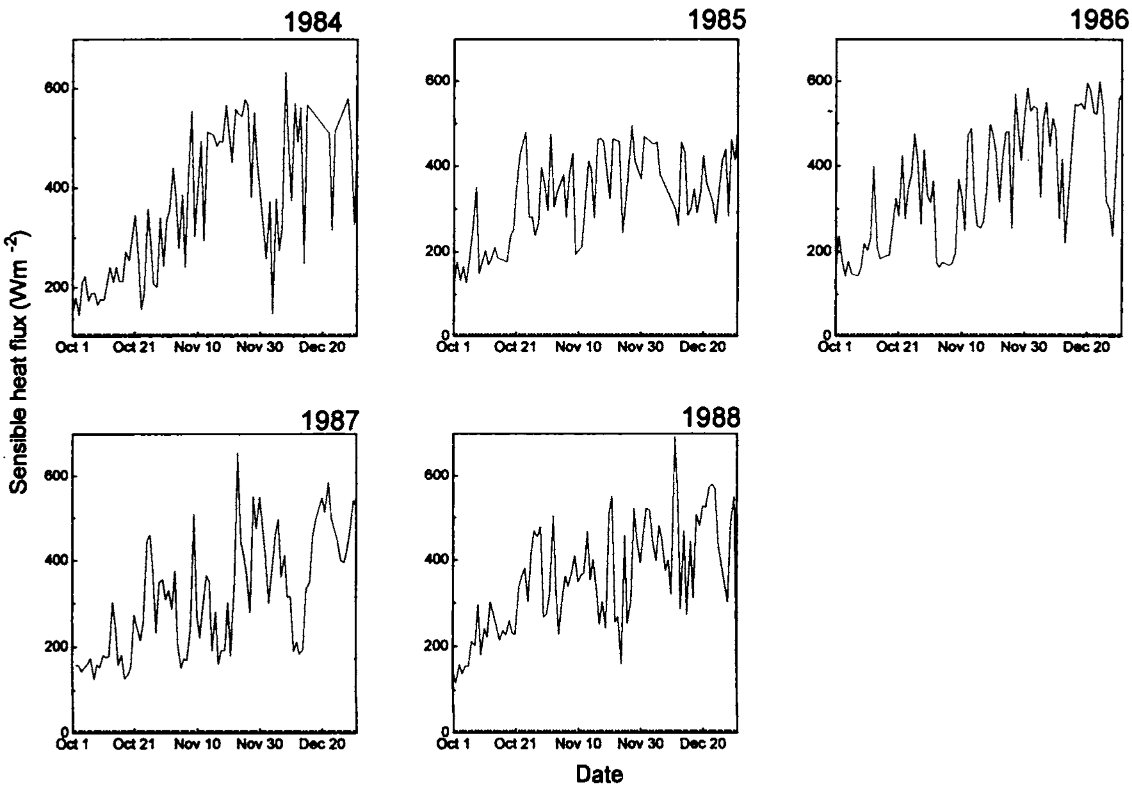

Fig. 2.4a Sensible heat flux during Oct. - Dec.1984 - 1988 at Bombay

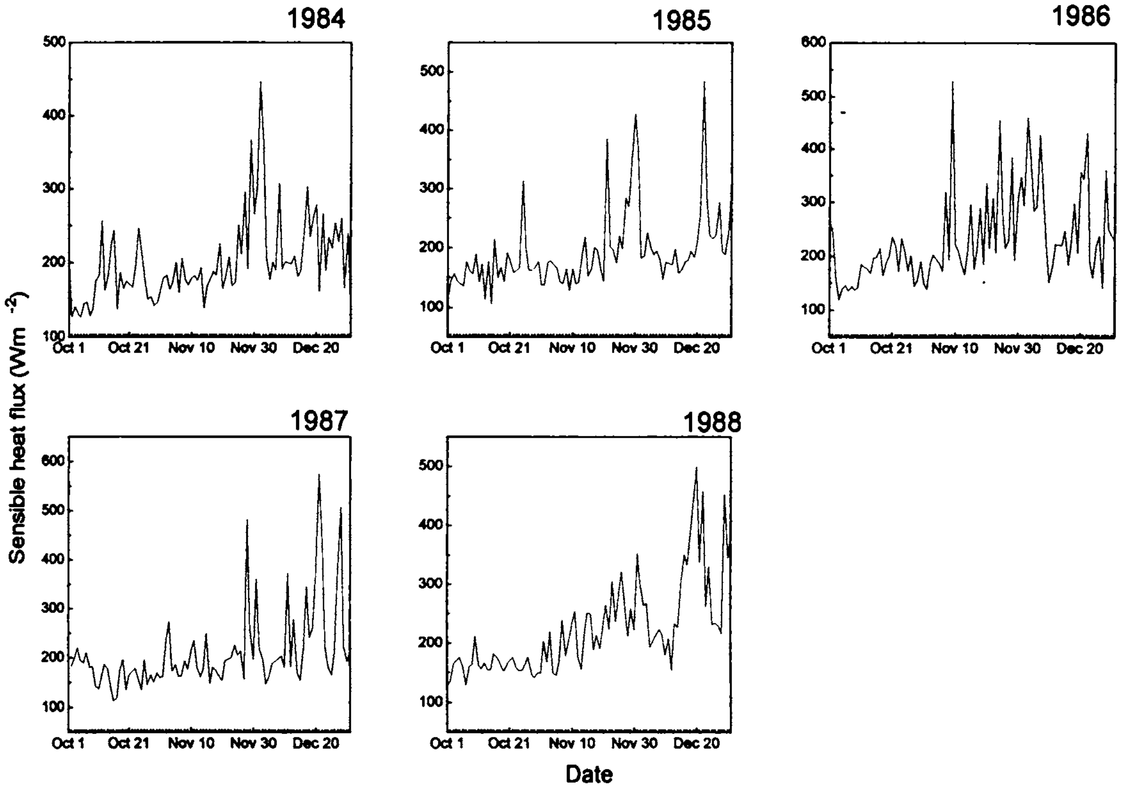

Fig. 2.4b Sensible heat flux during Oct.- Dec. 1984 -1988 at Mangalore

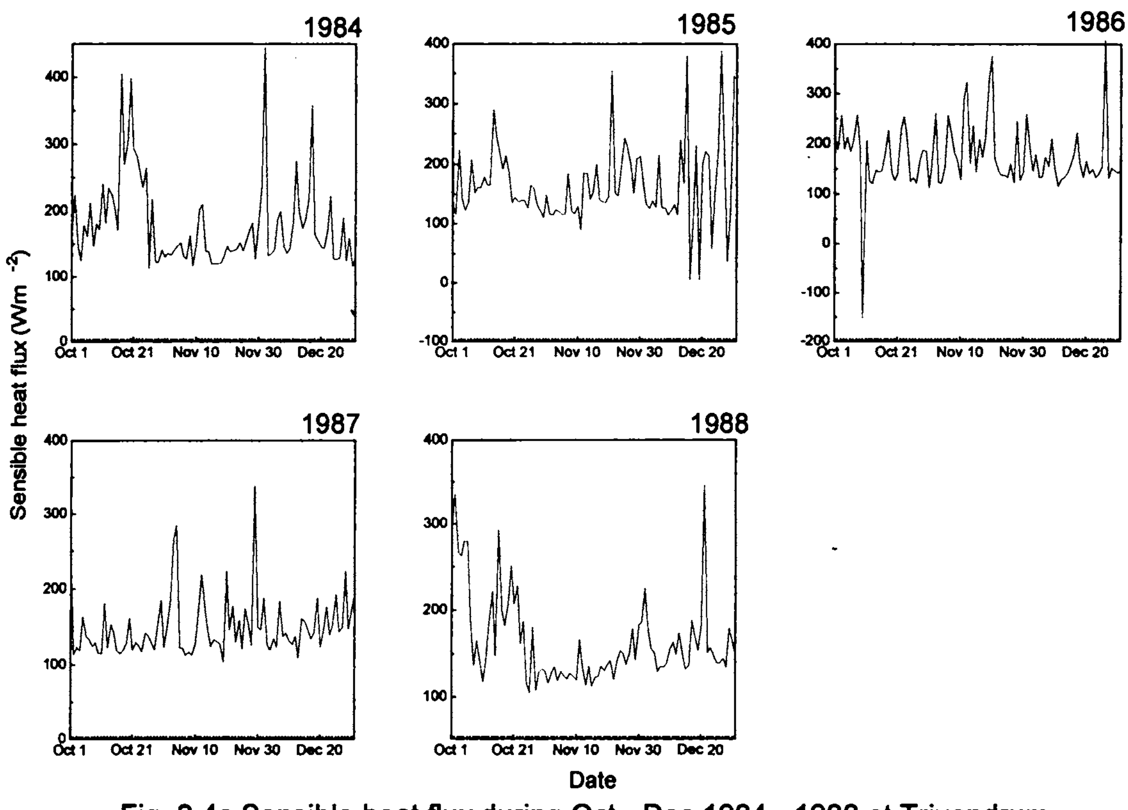

Fig. 2.4c Sensible heat flux during Oct.- Dec.1984 - 1988 at Trivandrum

c) Latent heat flux

The time series of latent heat fluxes at Bombay, Mangalore and Trivandrum during April to June, 1984-1988 is given in figures (2.5a to 2.5c) respectively. The latent heat flux during the period at Bombay was found to be between 250-500 Wm^{-2} with occasional rise. In certain days the flux values were less than 250Wm^{-2} and downward flux was noticed occasionally. Higher fluxes are noticed during late May and June, which is during the onset of southwest monsoon. At Mangalore the latent heat fluxes during April and May remain more or less constant and the range is found to be around 300-400Wm^{-2}. By monsoon onset high fluxes are noticed at the station and it may reach up to 650Wm^{-2}. At Trivandrum very high fluxes are noticed in certain days during the pre-monsoon months. By the monsoon onset the fluxes are very high. The flux may reach up to 750Wm^{-2}. The higher fluxes are due to increase in wind speed during the period in association with the onset. Higher fluxes are noticed at Trivandrum when compared to Bombay and Mangalore. This is because of the high-speed winds at the station. The surface latent heat flux is generally higher during the onset period.

Figures (2.6a to 2.6c) shows the time series of latent heat fluxes during October, November and December, from 1984-1988 at Bombay, Mangalore and Trivandrum respectively. At Bombay during the post-monsoon periods higher fluxes are noticed during October. After that the fluxes decreased and downward fluxes were noticed in certain days. Very high fluxes are noticed occasionally which may be due to the occurrence of disturbances. At Mangalore during October the flux remained almost constant, which was about 300-400Wm^{-2}. After that lesser flux values are noticed and downward flux transport occurs in certain days during November and December. At Trivandrum higher fluxes were noticed during the month of October. After that during November and December it remained almost constant and is found to be around 300Wm^{-2}. In certain days very high fluxes are noticed which is due to the disturbances during the period. Therefore in the post-monsoon periods the latent heat flux is found to decrease from October to December. Higher fluxes during October are due to the higher winds and rainfall activity in association with the withdrawal of southwest monsoon. After October the rainfall activity is very less therefore there will be less moisture

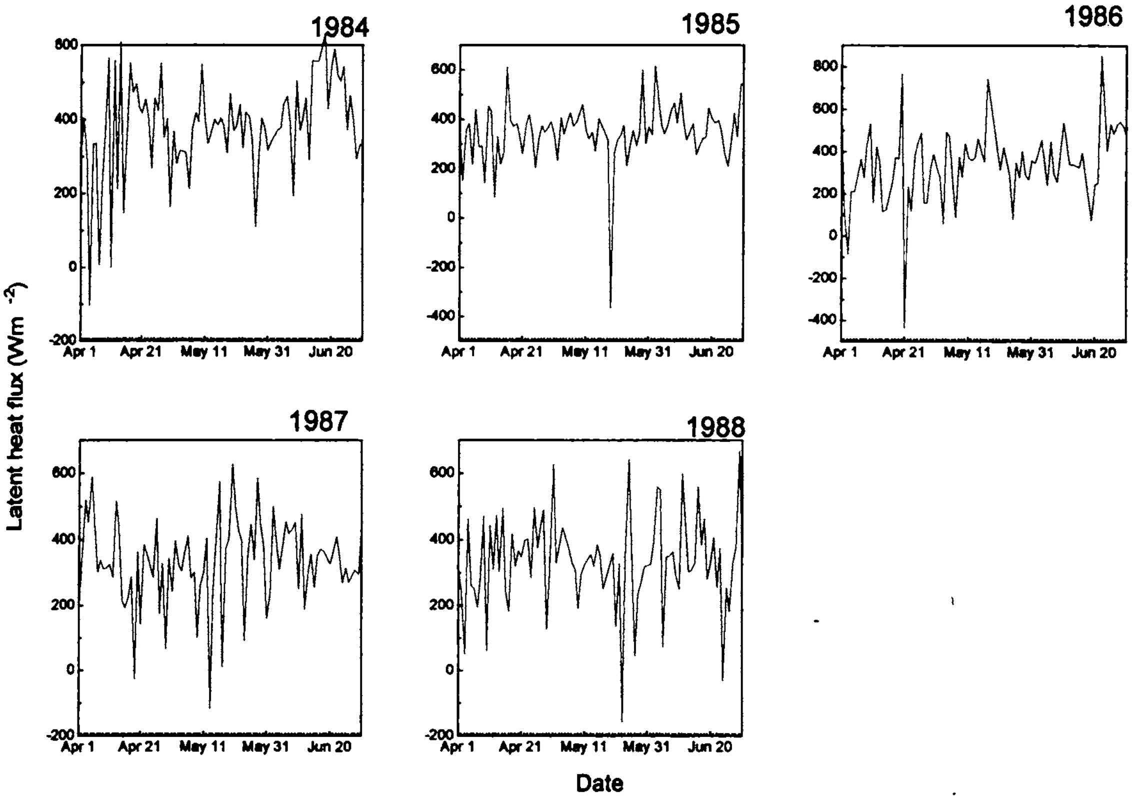

Fig. 2.5a Latent heat flux during April, May and June 1984 - 1988 at Bombay

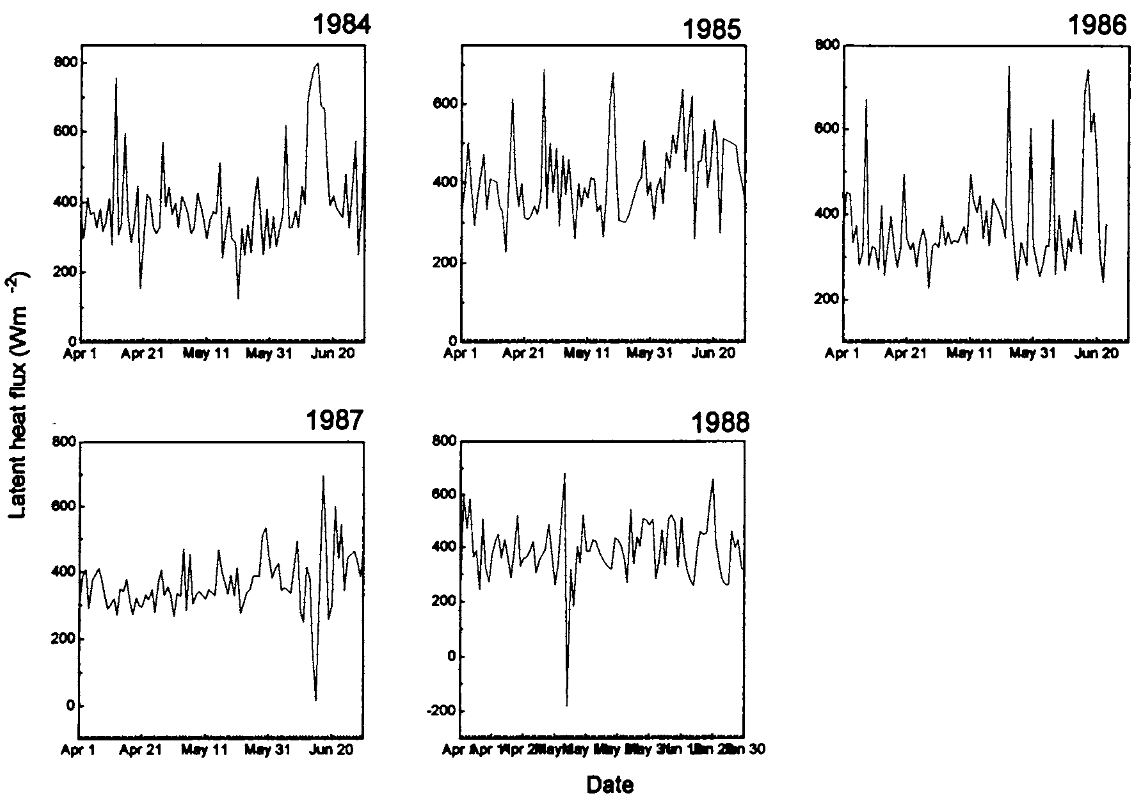

Fig. 2.5b Latent heat flux during April, May and June 1984 - 1988 at Mangalore

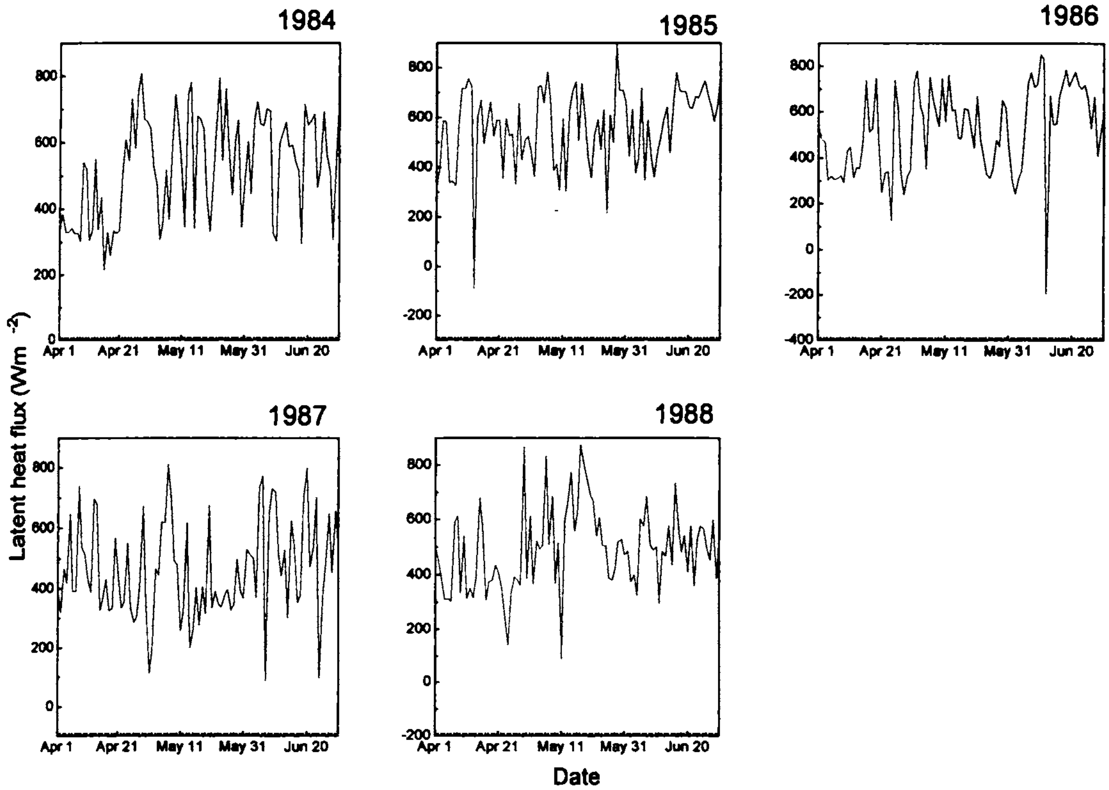

Fig. 2.5c Latent heat flux during April, May and June 1984 - 1988 at Trivandrum

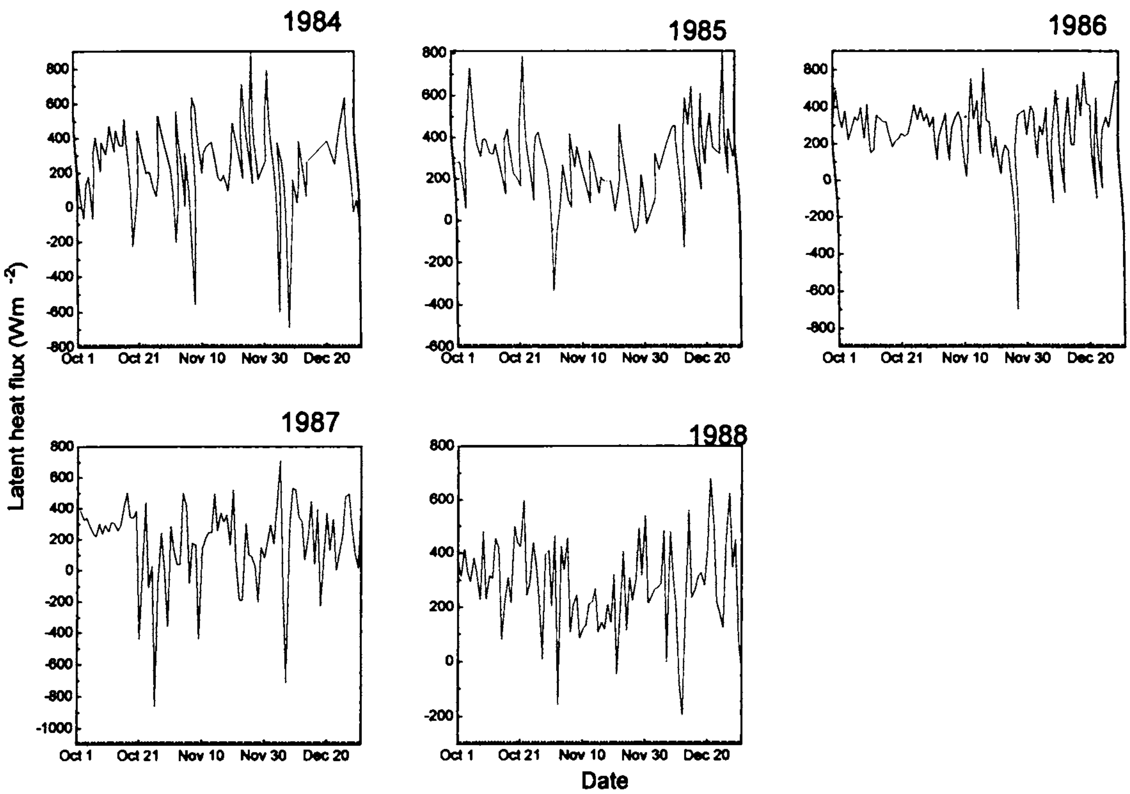

Fig. 2.6a Latent heat flux during Oct.-.Dec. 1984 - 1988 at Bombay

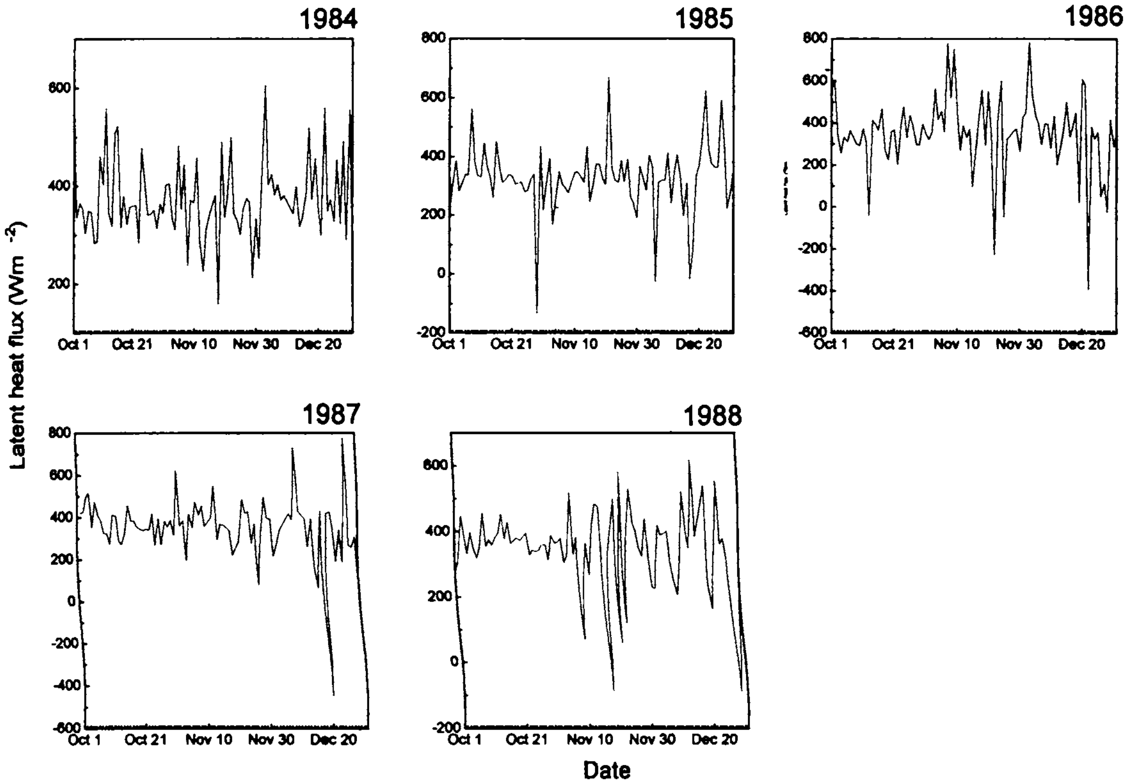

Fig. 2.6b Latent heat flux during Oct. - Dec. 1984 -1988 at Mangalore

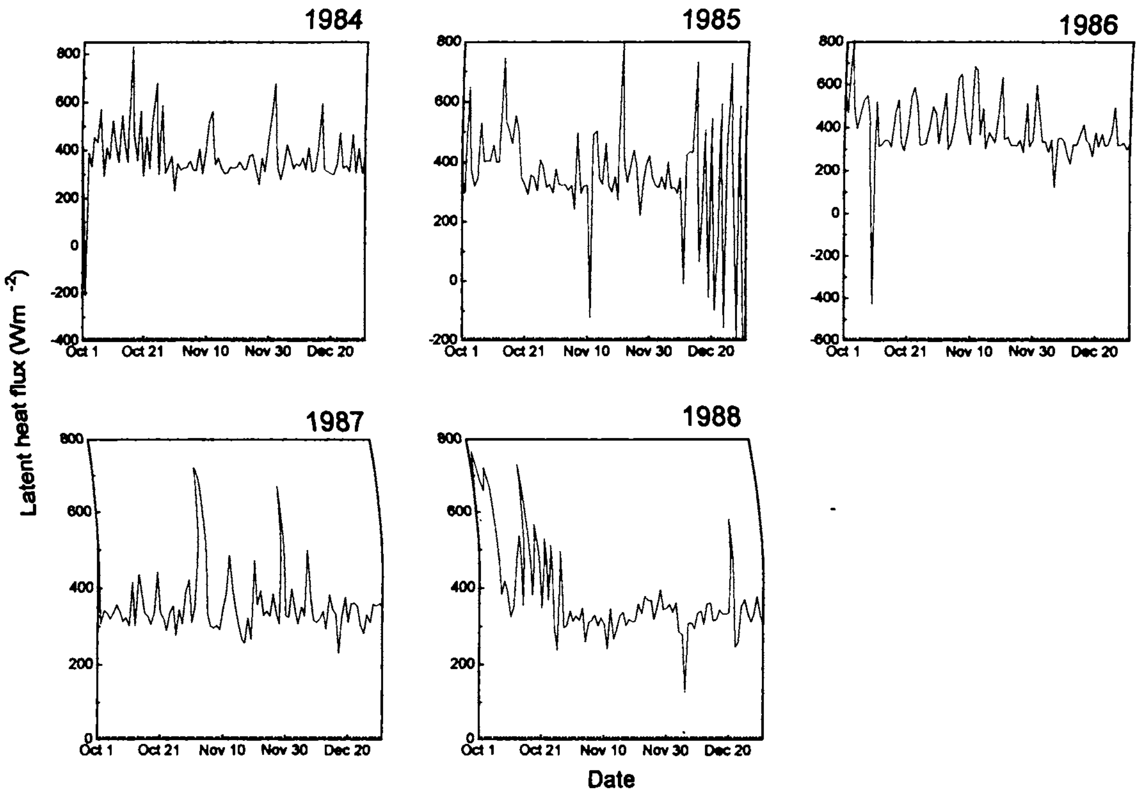

Fig. 2.6c Latent heat flux during Oct.- Dec. 1984 -1988 at Trivandrum

available at the surface for evaporation hence lesser values and downward transport is noticed. The sensible heat flux is found to increase at Bombay and Mangalore during the post-monsoon period whereas the latent heat flux is decreased. But during the onset period at all the three stations the sensible heat and latent heat fluxes increase. This shows that the monsoon activity influences the surface boundary layer characteristics by increasing the surface fluxes.

Surface stability parameter during pre-monsoon, onset and post-monsoon periods

Figures (2.7a to 2.7c) shows the variation of stability parameter during April, May and June from 1984-1988 at Bombay, Mangalore and Trivandrum. The stability parameter indicates that at Bombay the atmosphere tends to attain near neutral or less unstable condition during June by the monsoon onset. The surface layer which was highly unstable during April gradually tends to become less unstable in May and reaches near neutral condition in June. At Mangalore the stability parameter is highly fluctuating between unstable and neutral situations during the period. But by monsoon onset the atmosphere is in near neutral condition. At Trivandrum the atmosphere reaches near neutral condition most of the days in April and May. This may be due to the presence of moist winds during the period. In June the atmosphere is in near neutral condition through out the period.

The time series of stability parameter during October, November and December 1984-1988 at Bombay, Mangalore and Trivandrum are given in figures (2.8a to 2.8c) respectively. At Bombay the atmosphere which was less unstable during October becomes more unstable in most of the days during December. That is the atmosphere tends to become highly unstable in most of the days during December from the less unstable situation in October. As soon as it becomes highly unstable the atmosphere tries to attain neutrality or less unstable situation. At Mangalore the surface layer during the period fluctuates between highly unstable and less unstable condition. At Trivandrum in early October the atmosphere was in near neutral condition. After that the atmospheric condition is fluctuating between the highly unstable and less unstable conditions. That is during November and December the atmosphere is highly unstable and it always tends to less unstable situation from highly unstable by increasing the turbulence in the boundary layer. But in October which is the withdrawal period of monsoon the atmosphere is less

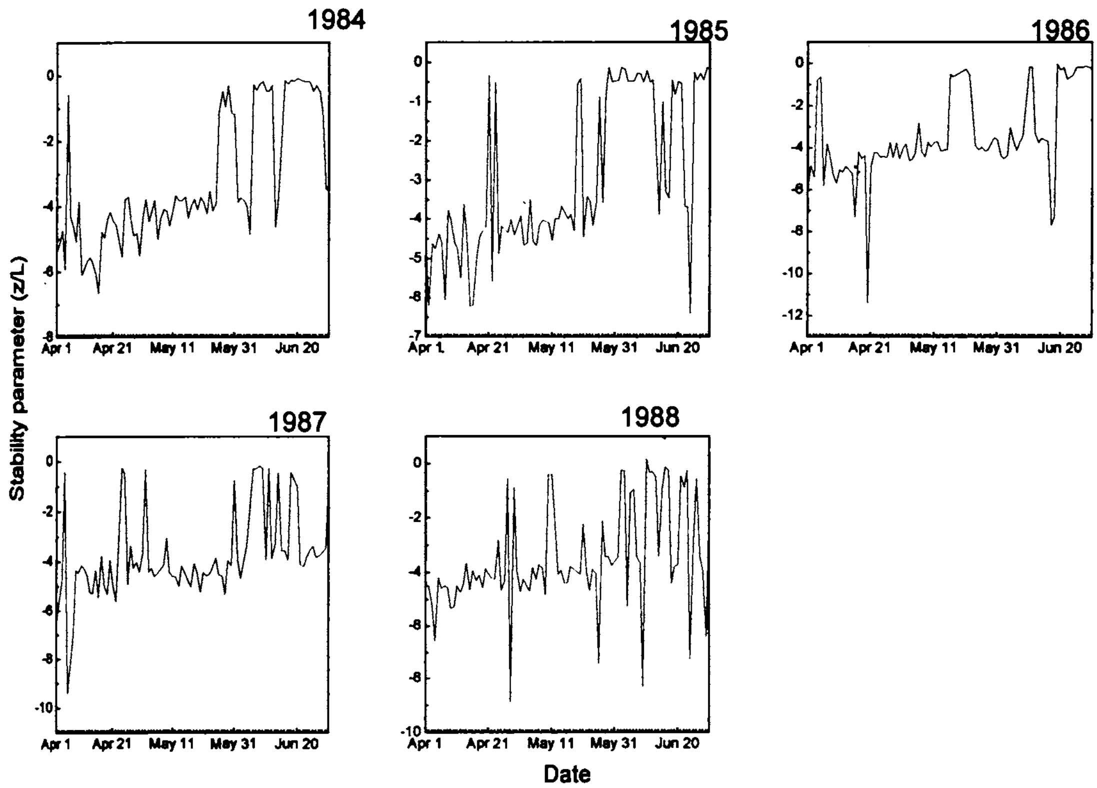

Fig. 2.7a Stability parameter during April-June 1984 - 1988 at Bombay

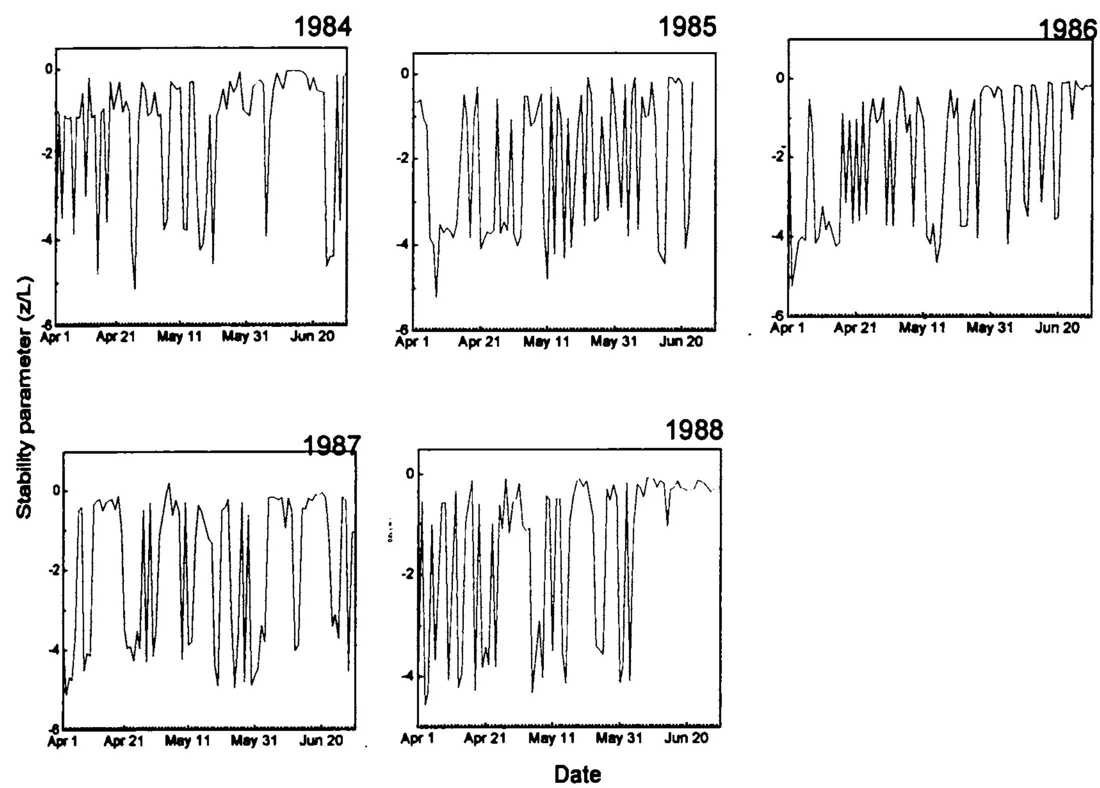

Fig. 2.7b Stability parameter during April-June1984 -1988 at Mangalore

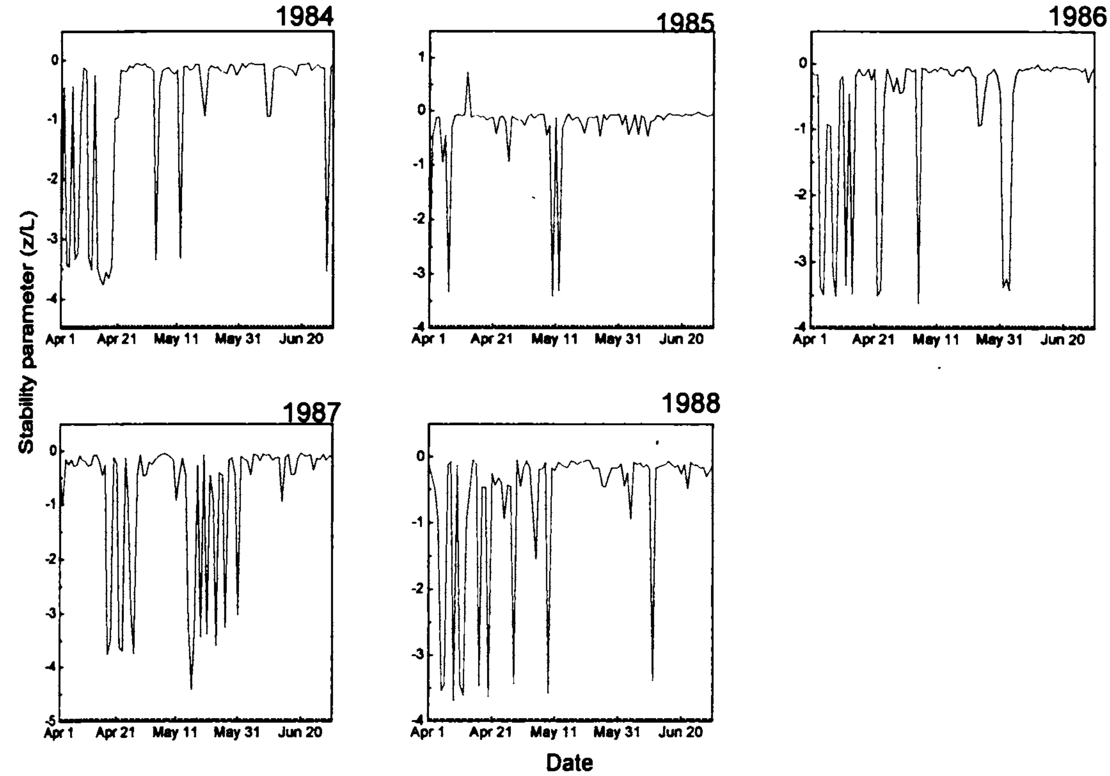

Fig. 2.7c Stability parameter during April-June 1984 - 1988 at Trivandrum

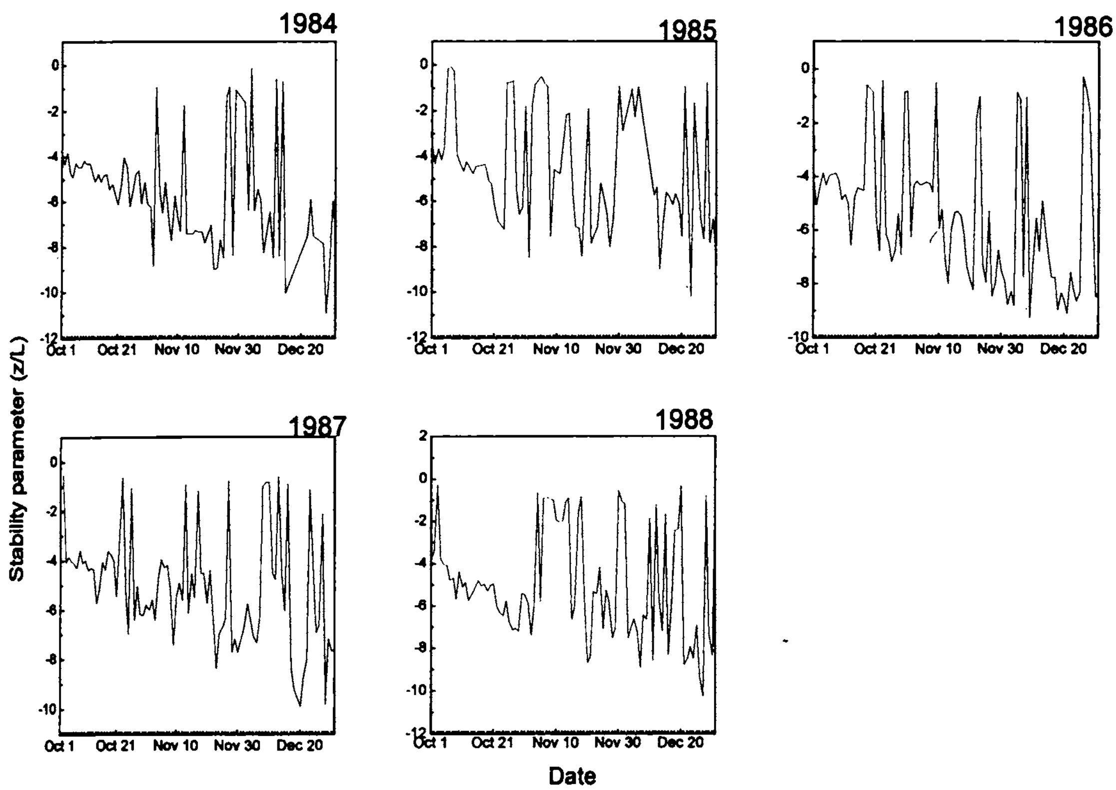

Fig. 2.8a Stability parameter during Oct.-Dec.1984 - 1988 at Bombay

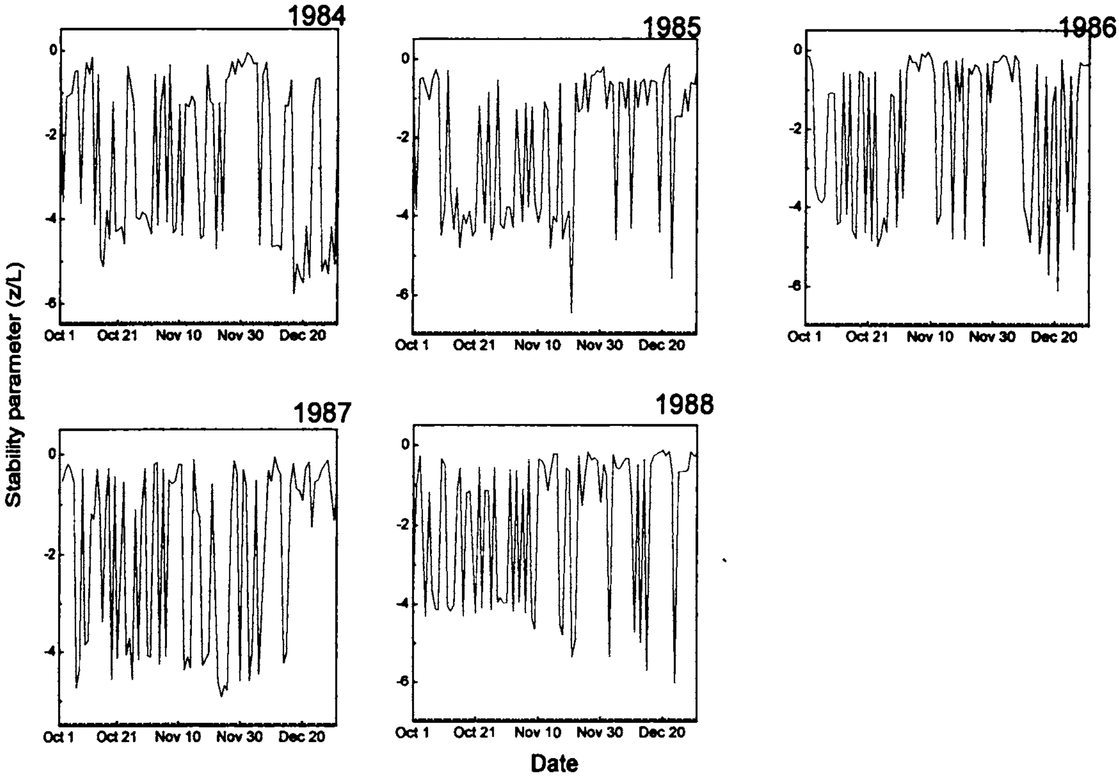

Fig. 2.8b Stability parameter during Oct.-Dec. 1984 -1988 at Mangalore

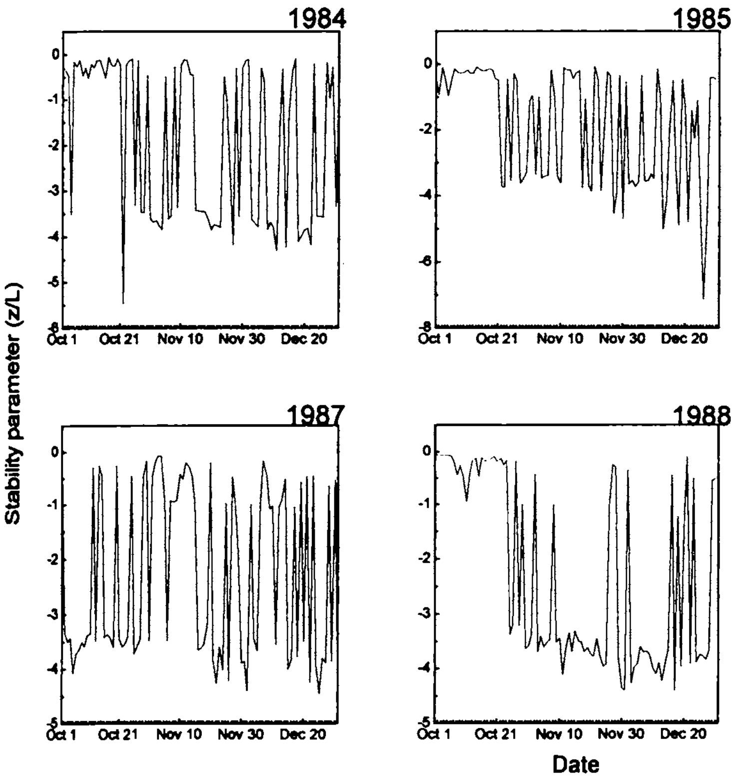

Fig. 2.8c Stability parameter during Oct.-Dec.1984 - 1988 at Trivandrum

unstable or in near neutral condition. Thus the activity of the monsoon has got profound influence on the surface layer stability. Therefore all the activities in the surface layer, particularly the exchange processes do get influenced by monsoon activity.

In general at the three stations considered the surface layer becomes less unstable or near neutral by the onset of monsoon. This is because of the presence of moist winds and radiative cooling of the surface layer due to overcast sky during onset period. The less unstable situation during October is also because of the moist winds and radiative cooling of the surface during the withdrawal period of monsoon. During the pre-monsoon and post-monsoon months when usually clear sky conditions prevail the atmosphere becomes highly unstable. Whenever the atmosphere becomes highly unstable it soon tries to attain less unstable situation. This is achieved by increasing the turbulence and therefore the turbulent fluxes. When the atmosphere becomes highly unstable the surface turbulence increases and the turbulent fluxes are increased so that the atmosphere attains less unstable condition. High fluxes are noticed soon after the atmosphere is highly unstable so that less unstable situation is attained.

Surface layer characteristics during the occurrence of thunderstorms

Variations in surface fluxes and surface stability during the occurrence of thunderstorms in the pre-monsoon and post-monsoon months at Trivandrum are shown in figures (2.9a to 2.9J). During the pre-monsoon period the frequencies of thunderstorm are more in the month of April whereas during the post-monsoon season the October and November months give more thunderstorms. The stability parameter shows highly unstable situation in the atmosphere before the occurrence of thunderstorm and it attains less unstable or near neutral condition soon after the occurrence of the thunderstorm. In April during the occurrence of thunderstorms the stability parameter is found to be fluctuating between highly unstable and less unstable situations. The atmosphere becomes less unstable or near neutral after the thunderstorm. This may be due to the radiative cooling of the ground and presence of moist winds after the rainfall from thunderstorms.

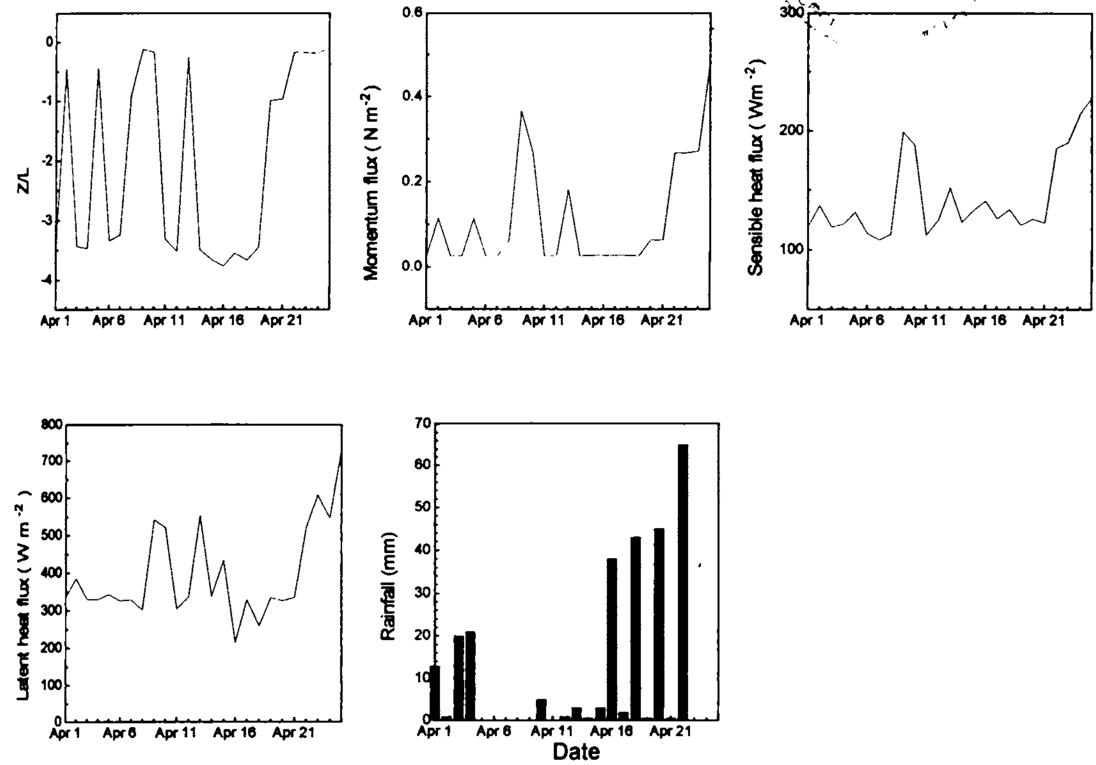

Fig. 2.9a Stability parameter, surface fluxes and rainfall associated with thunderstorms at Trivandrum during April 1984

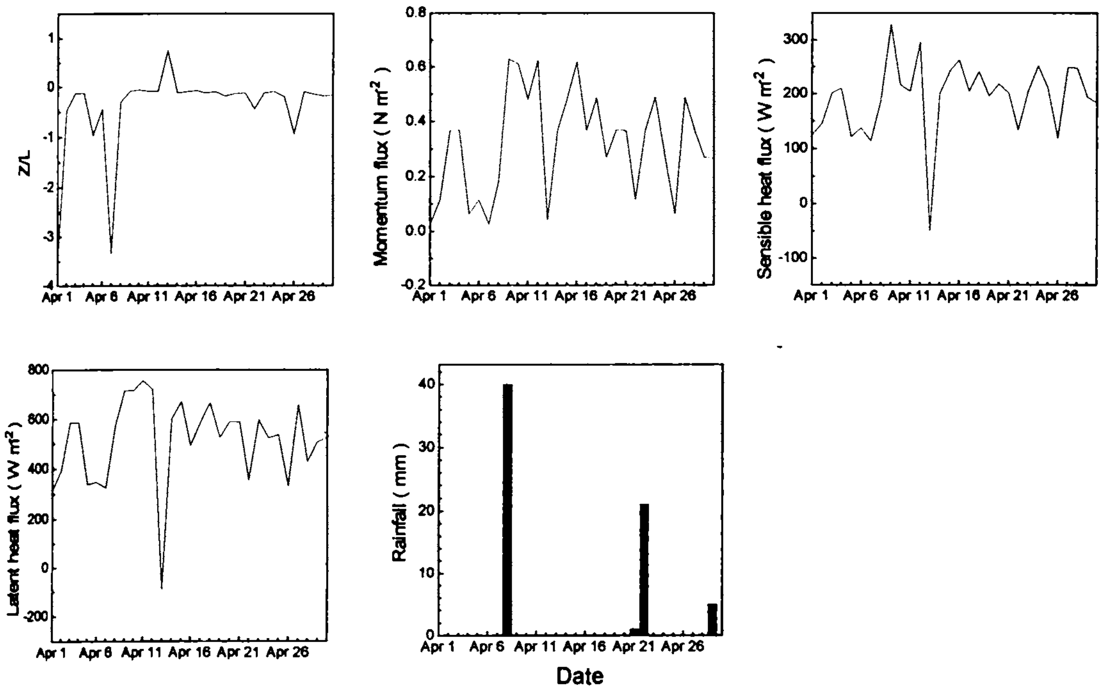

Fig. 2.9b Stability parameter, surface fluxes and rainfall associated with thunderstorms at Trivandrum during April 1985

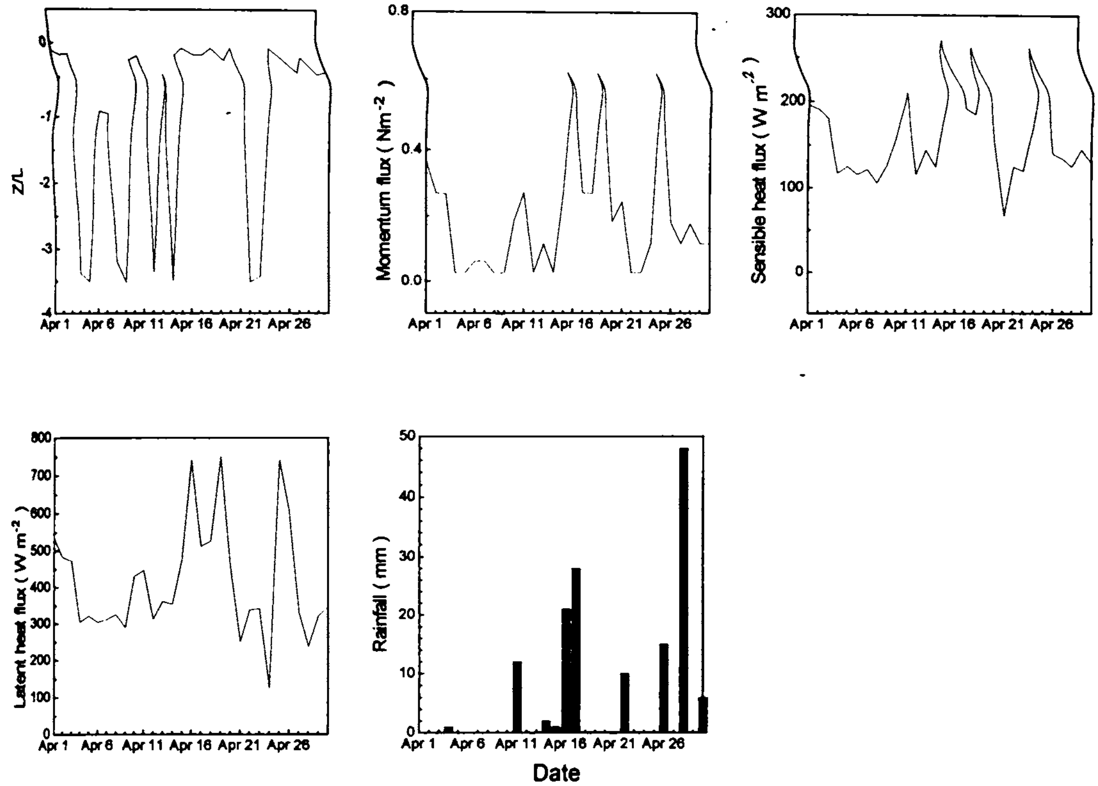

Fig. 2.9c Stability parameter,surface fluxes and rainfall associated with thunderstorms at Trivandrum during April 1986

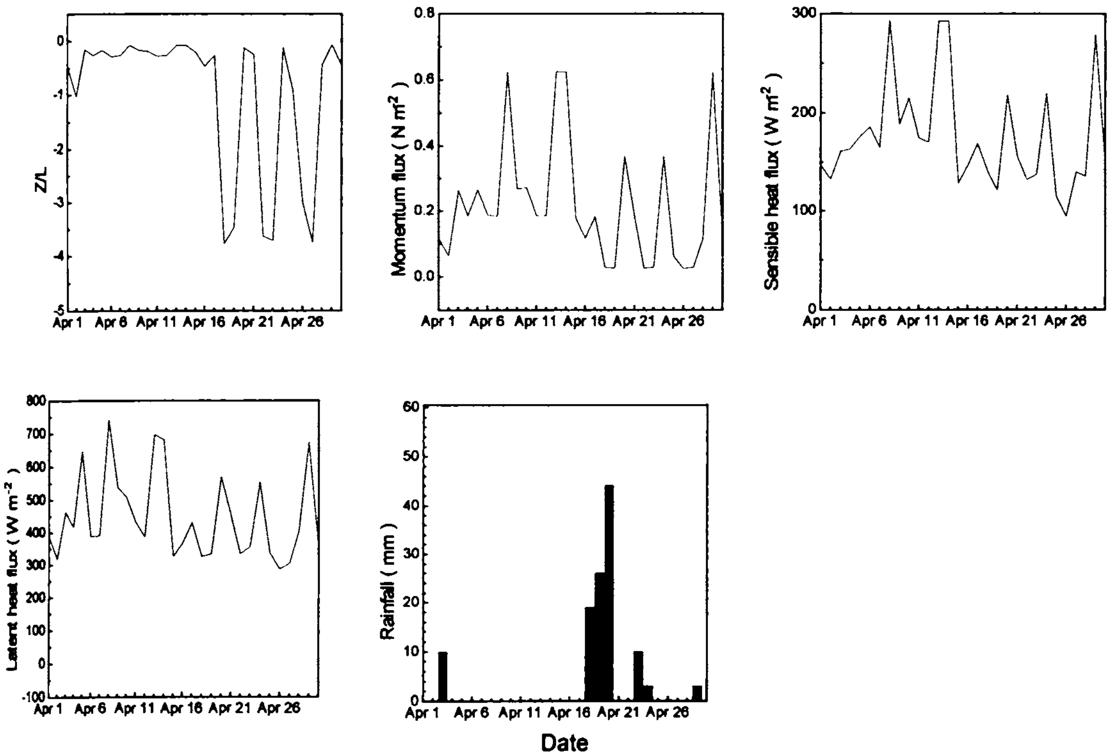

Fig. 2.9d Stability parameter, surface fluxes, rainfall associated with thunderstorms at Trivandrum during April 1987

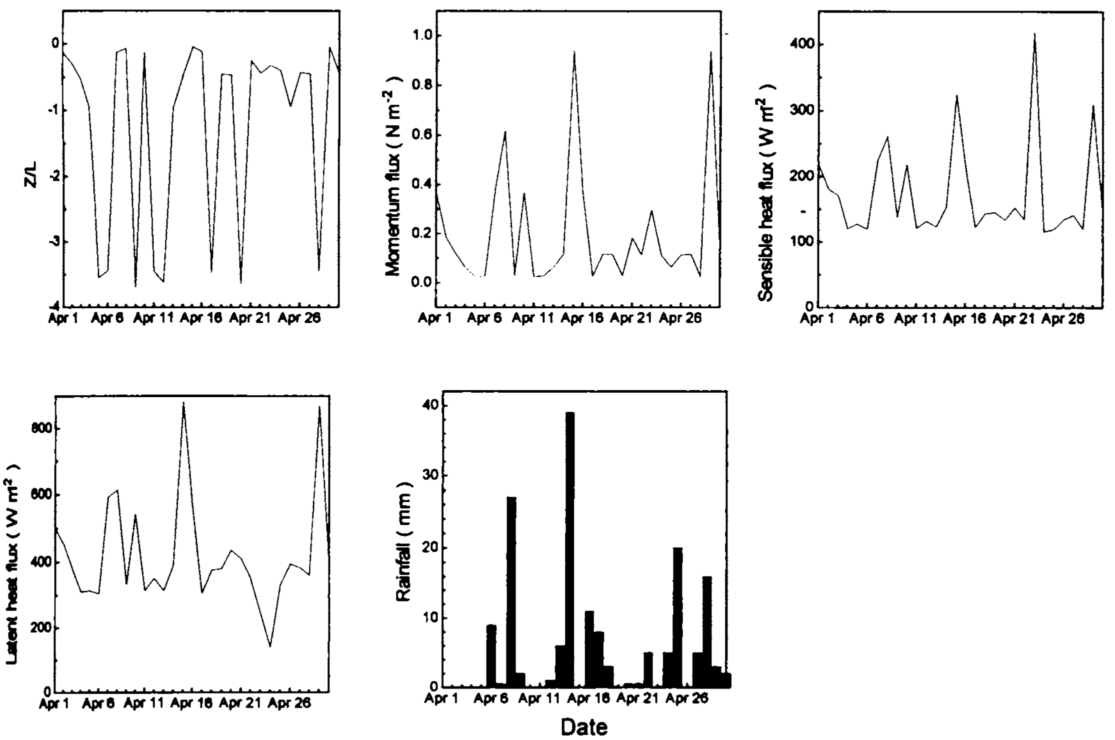

Fig. 2.9e Stability parameter, surface fluxes and rainfall associated with thunderstorms at Trivandrum during April 1988

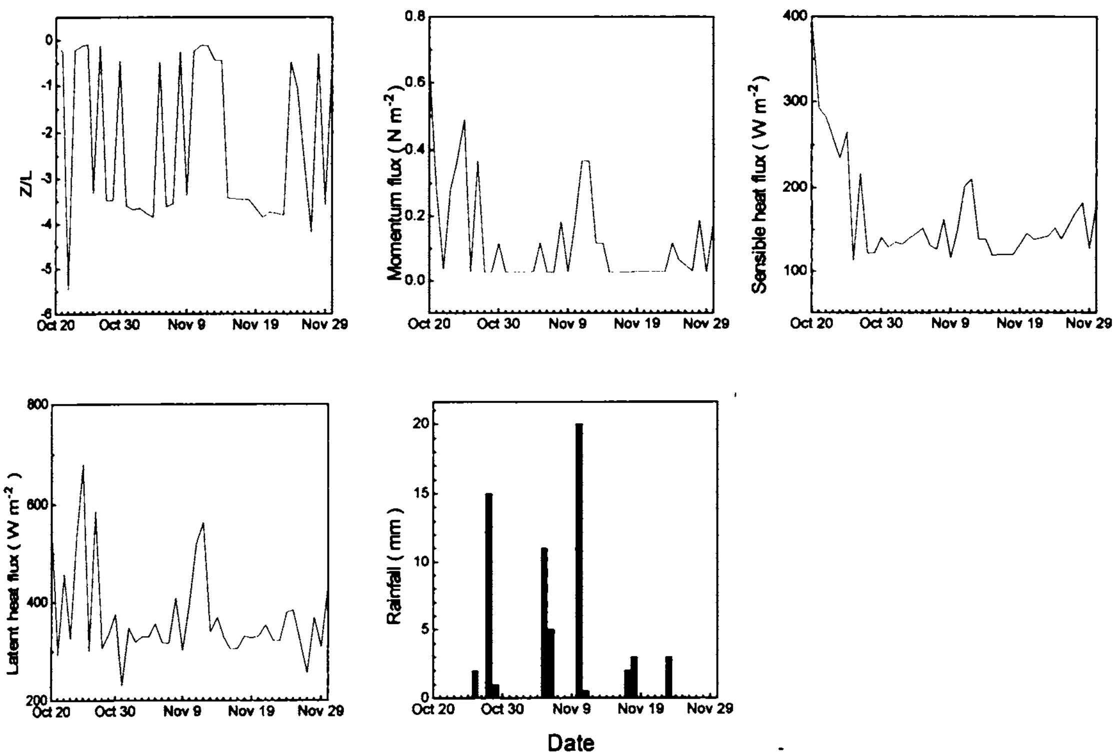

Fig. 2.9f Stability parameter, surface fluxes and rainfall associated with thunderstorms at Trivandrum during Oct. - Nov. 1984

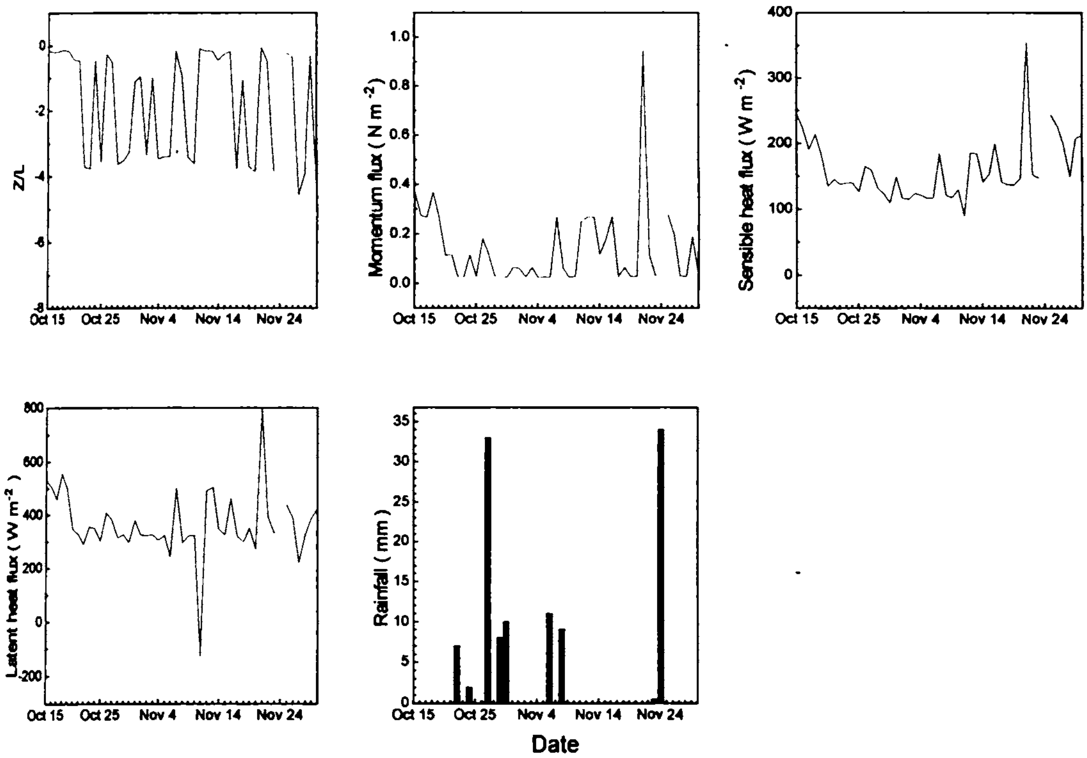

Fig. 2.9g Stability parameter,surface fluxes and rainfall asspciated with thunderstorms at Trivandrum during Oct.- Nov. 1985

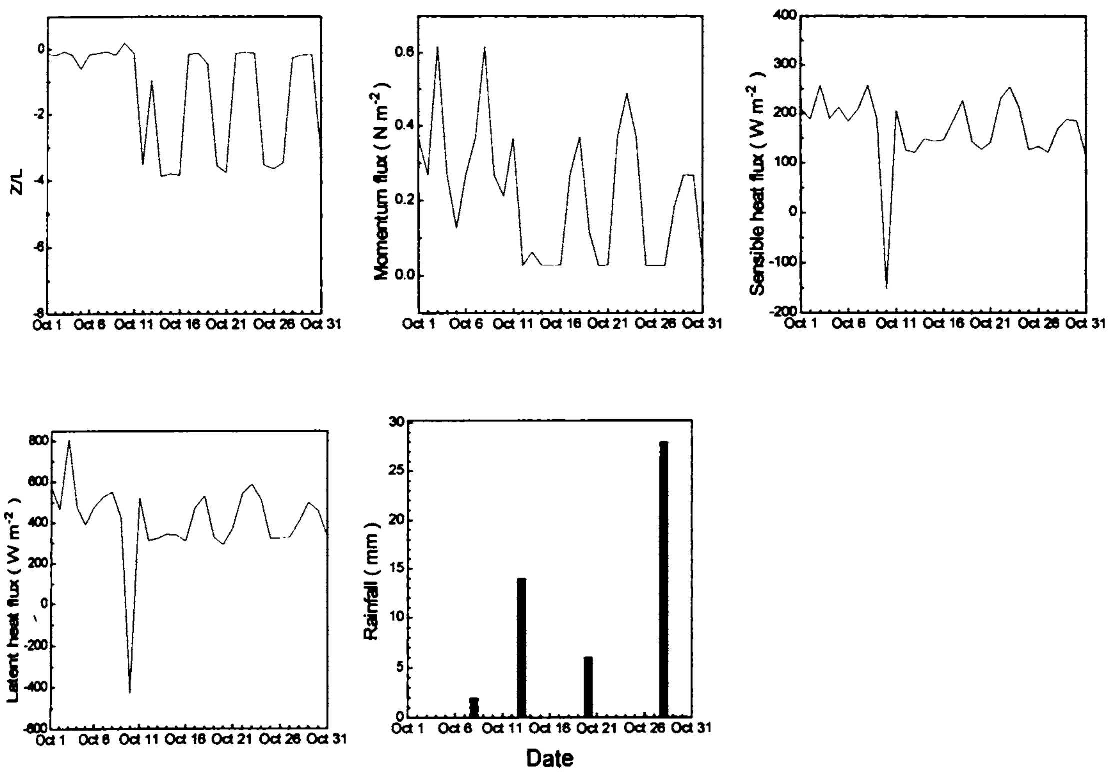

fig. 2.9h Stability parameter,surface fluxes and rainfall associated with thunderstorms at Trivandrum during Oct.-Nov. 1986

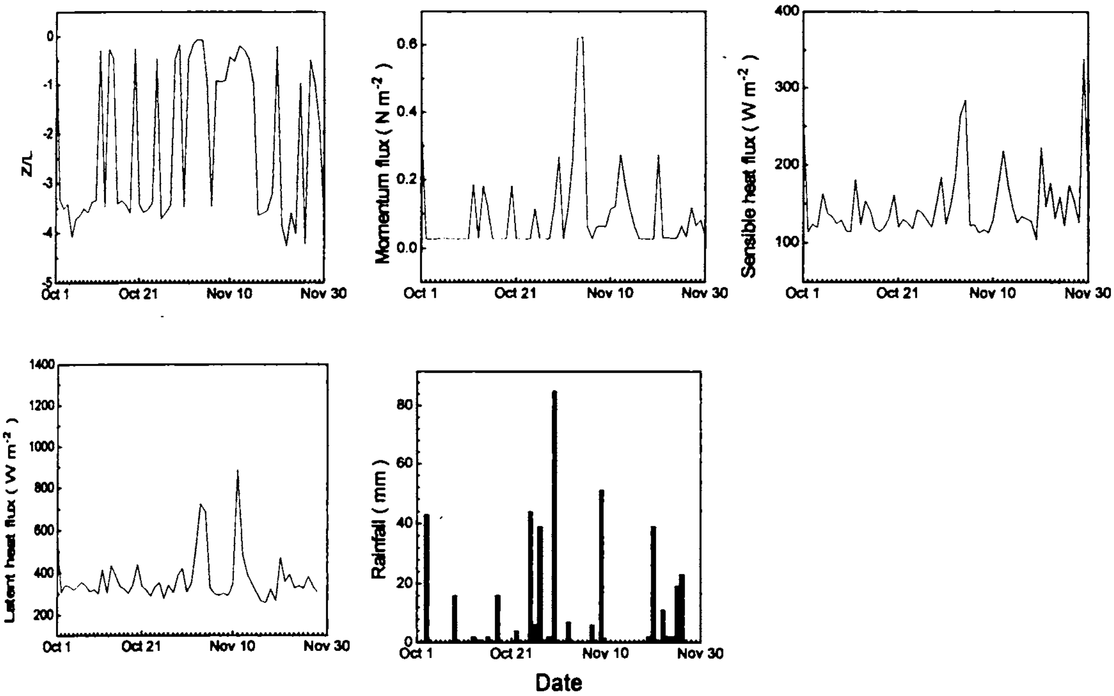

Fig. 2.9i Stability parameter, surface fluxes and rainfall associated with thunderstorms at Trivandrum during Oct.-Nov.1987

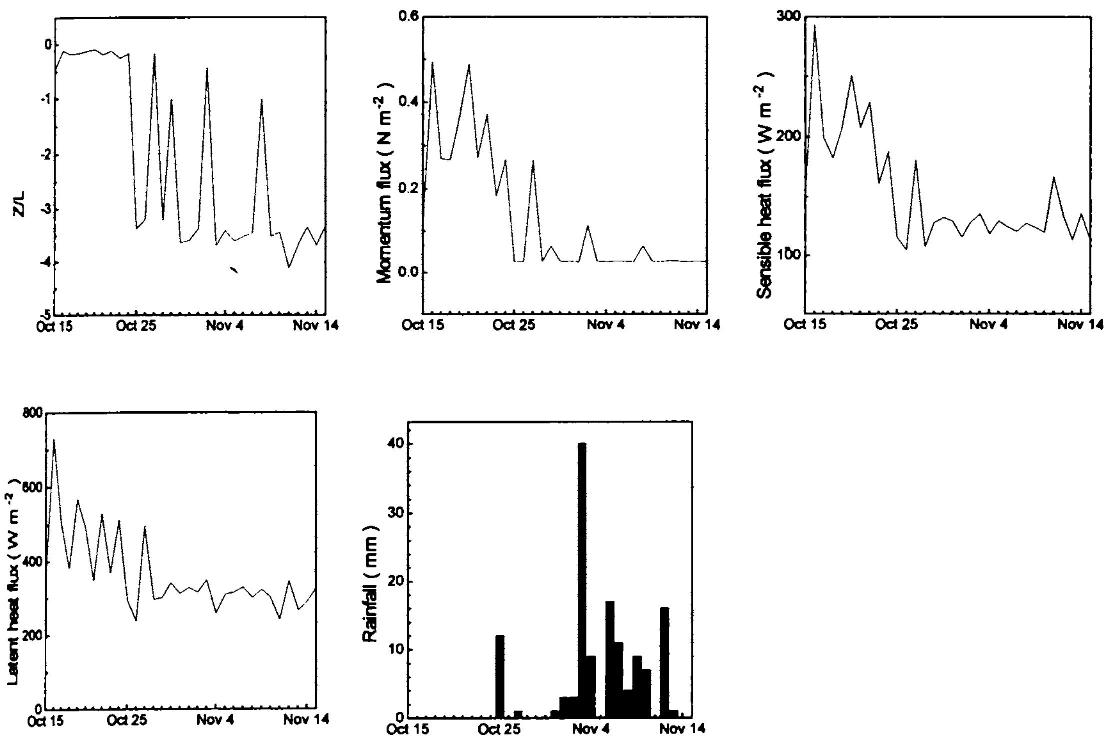

Fig. 2.9j Stability parameter, surface fluxes and rainfall associated with thunderstorms at Trivandrum during Oct.- Nov. 1988

In the post-monsoon period most of the thunderstorm activity is during October and November. During these months also the atmospheric stability condition fluctuates between highly unstable and less unstable after the thunderstorm activity. Before the thunderstorm the atmosphere becomes highly unstable and becomes less unstable soon after the thunderstorm activity. Therefore the stability parameter in the atmospheric surface layer gives a clear indication of occurrence of thunderstorms much before its activity occurs in the atmosphere. The thunderstorm activity very much influences the surface layer stability, hence the energy exchange processes in the surface layer also gets affected by it.

The surface layer momentum, sensible heat and latent heat flux shows an increase in association with the occurrence of thunderstorms. As soon as the atmosphere becomes highly unstable prior to the occurrence of thunderstorm the surface turbulence increases and the surface turbulent fluxes are increased so that the atmospheric flow becomes less unstable. During the pre-monsoon period the momentum flux is found to reach $0.6Nm^{-2}$ and occasionally more than that. The surface sensible heat flux reached a maximum of $300Wm^{-2}$ and in certain days higher values are noticed. The latent heat flux is found to be slightly higher in magnitude in certain days which was found to be about $800Wm^{-2}$. During the post-monsoon period the surface momentum flux was found to be $0.4Nm^{-2}$ and occasionally higher. The surface sensible heat flux was found to be less than $300Wm^{-2}$ and downward flux was noticed as the atmosphere becomes highly stable. The latent heat was found to reach a maximum of $600Wm^{-2}$ and sometimes higher. Therefore the thunderstorm activity influences the surface layer exchange properties. The surface fluxes were high during the occurrence of thunderstorms and because of the radiative cooling of the surface due to intense rainfall activity the atmosphere becomes neutral or stable. The convection in association with the thunderstorms increases the surface turbulence and hence the surface fluxes.

Surface layer characteristics during the occurrence of monsoon depression

Surface layer stability parameter and surface fluxes over various stations are computed during the life period of the depression (August2- August 6) formed over Bay of Bengal in 1988. The track of the depression is shown in figure (2.1). The stations selected are also given in the figure. The depression was formed over the head Bay on

August 2^{nd} and crossed West Bengal coast near Calcutta on the same day evening. The surface layer characteristics at Calcutta are given in figure (2.10a). The stability parameter shows unstable condition on August 1, which attained less unstable or near neutral condition by August2. Near neutral condition prevailed in the atmosphere till August 3^{rd} and became highly unstable at 1200 UTC on August 4 and then it fluctuated between highly unstable and less unstable condition. The atmosphere was found to be highly unstable prior to the formation of depression and it soon became less unstable as the system reached the station. The less unstable situation is reached as a result of cooling of the surface due to overcast skies and moist winds due intense rainfall activity during the period. As the effect of the system was decreased over the station the atmosphere returned to highly unstable condition and is found to transit between unstable and less unstable situation. The momentum flux was found to increase from August 1^{st} and reached maximum on August 2^{nd} and 3^{rd} and then decreased. The flux was found to increase as the system is near the station after that it decreased and increased at 12UTC on August 5^{th}. The surface layer is highly unstable on August 5^{th} at 00UTC and due to this surface turbulence increases and the turbulent exchange increases so that higher fluxes are noticed soon after. The surface latent heat and sensible heat fluxes also shows similar trend. A decrease in fluxes was noticed on August 2^{nd} and it increased as the system crosses land. The decrease was as a result of the radiative cooling of the ground due to overcast skies and presence of moisture as a result of the rainfall from the system. The atmospheric condition on the day was almost neutral. The latent heat flux decreased on August 4^{th} after the effect of the system over the station decreased. An increase was noticed further in association with the instability in the atmosphere. The fluxes were found to be slightly higher in magnitude, which might be due to the error occurred in the radiosonde observation during the period. The sensible heat flux decreases on August 3^{rd} due to radiative cooling of the surface layer as a result of the overcast skies. Then it increases on August 5^{th} due to increase in turbulence as a result of high instability in the atmosphere. At Masulipatnam (fig.2.10b) which is an east coast station the effect of the depression is noticed in the surface layer characteristics. The atmosphere, which was unstable on July 31^{st} and at 00UTC on August 1^{st} became less unstable and remained so on till August 2^{nd}. After that it fluctuated between highly unstable and less unstable situation. High instability was noticed on August 4^{th} at 12UTC. As soon as the atmosphere becomes highly unstable it tries to attain less unstable situation by increasing the surface turbulence and hence the surface fluxes. Increase in surface fluxes was

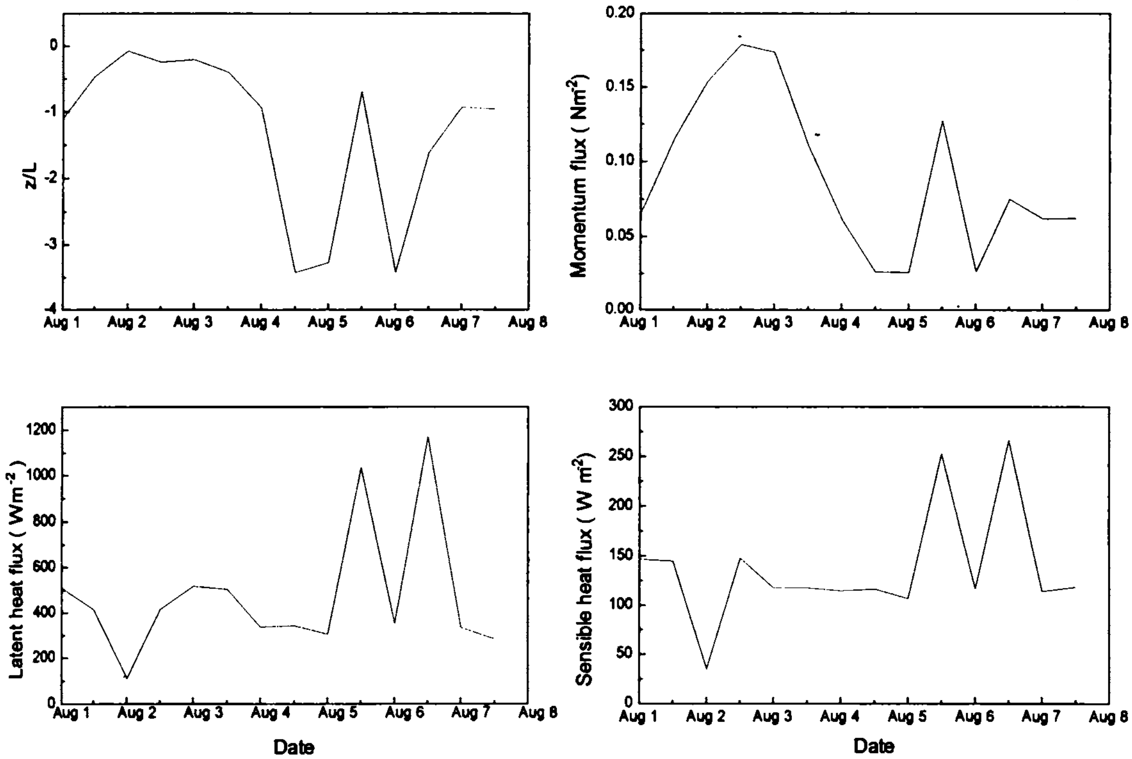

Fig. 2.10a Stability parameter and surface fluxes at Calcutta (Jul. 31- Aug. 7 1988)

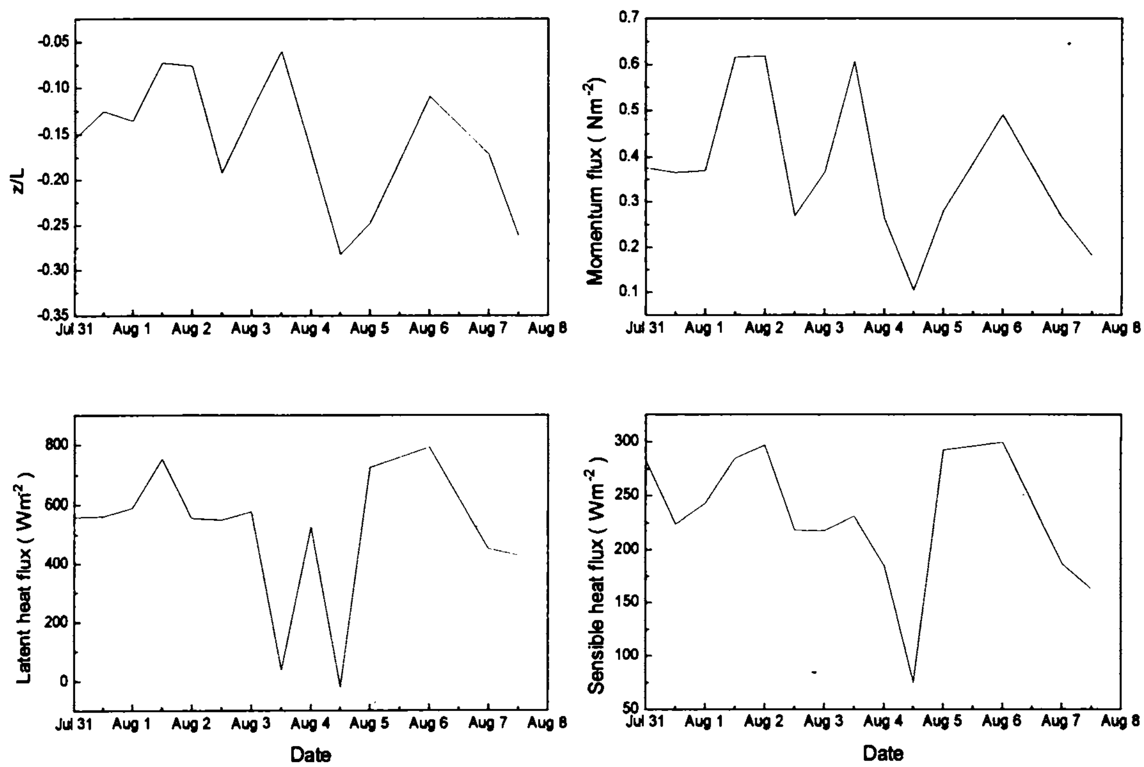

Fig. 2.10b Surface fluxes and stability parameter at Masulipatanam(Jul. 31-Aug7. 1988)

noticed soon after the atmosphere is highly unstable. The increase is due to strong winds during the disturbance. At Madras also high instability is noticed on July 31st before the formation of the depression (fig.2.10c). The atmosphere becomes less unstable during August 1-3 after which it again becomes highly unstable and transits between highly unstable and less unstable condition. The less unstable situation during the occurrence of depression is achieved as a result of the radiative cooling of the surface due to overcast skies and moist winds. The surface momentum, sensible heat and latent heat fluxes are higher during the system. Higher fluxes are noticed soon after the atmosphere becomes highly unstable. At Amini and Minicoy (figures 2.10d and 2.10e) the atmospheric condition fluctuates between highly unstable and less unstable situation during the disturbance and becomes highly unstable after the decay of the disturbance. Higher surface fluxes are noticed throughout the period at the two stations. At Mangalore (fig.2.10f) the atmosphere is in near neutral condition during the life period of the depression and becomes highly unstable after August 5th. Near neutral condition is due to overcast skies and moist winds due to the disturbance. After the effect of the depression decreases the atmosphere becomes highly unstable. The surface momentum, sensible and latent heat fluxes are very high during August 2-5 after which it decreases. The high-speed winds during the period are responsible for the increase in surface fluxes. At Goa the atmosphere is highly unstable prior to the formation of the depression and becomes near neutral or stable during the disturbance and becomes unstable again on August 5th (fig.2.10g). The surface fluxes are much higher during the period. Downward flux was noticed in latent heat flux on August 2nd and 6th. Downward sensible heat flux occurs over the station during the period as a result of the high stable condition reached in the atmosphere. At Bombay (fig.2.10h) the atmosphere was highly stable on July 31st which gradually became unstable on August 2nd. It was in near neutral condition during August 2 –5 due to the depression. After that the atmospheric condition fluctuates between highly unstable and less unstable condition. There was a decrease in surface flux during July 31st and downward sensible heat flux was noticed on the day. This was due to the highly stable condition of the atmosphere on July 31st. After that the surface fluxes increases and higher fluxes are noticed during August 1 - 4.

The atmosphere therefore becomes highly unstable much before the disturbance and it attains near neutral situation or less unstable situation during the life period of the disturbance due to radiative cooling of the ground because of highly overcast skies and

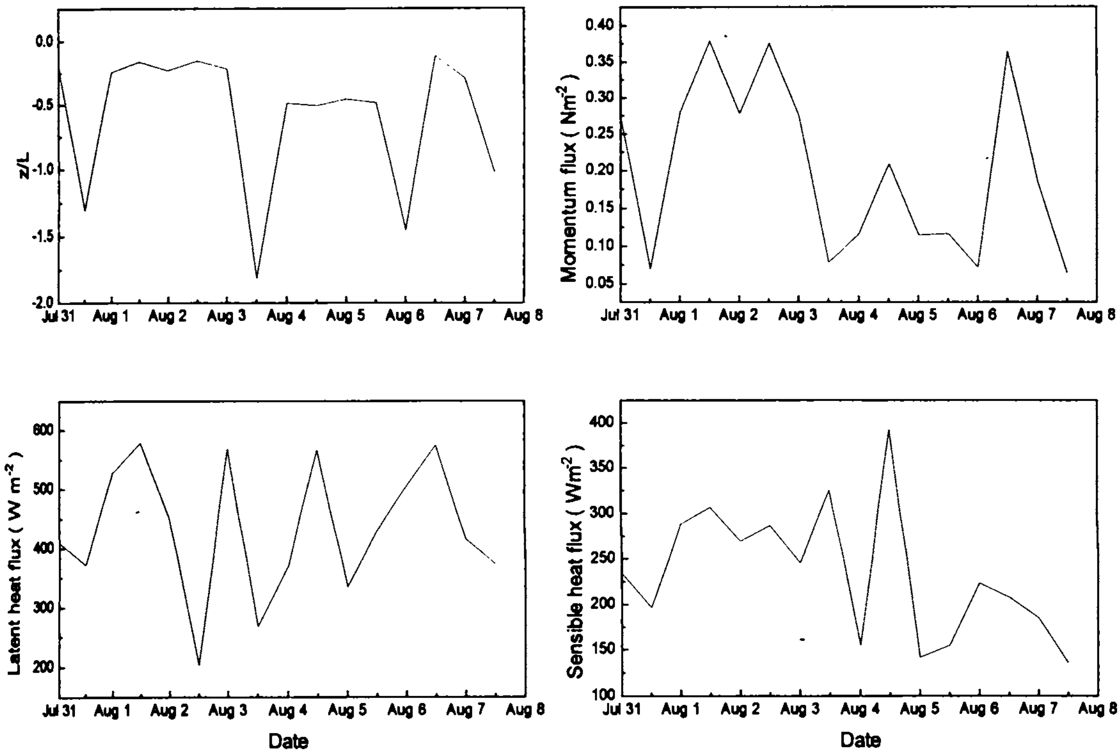

Fig. 2.10c Stability parameter and surface fluxes at Madras(Jul. 31-Aug.7 1988)

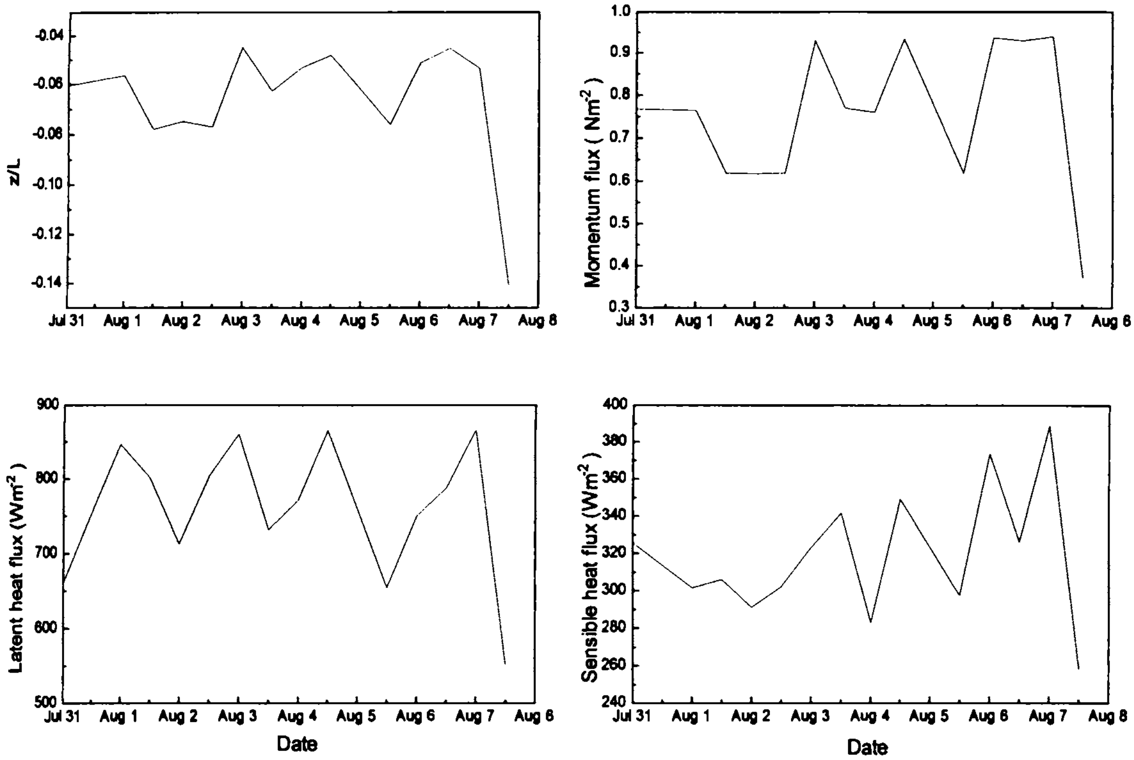

Fig. 2.10d Stability parameter and surface fluxes at Amini (Jul 31.-Aug7. 1988)

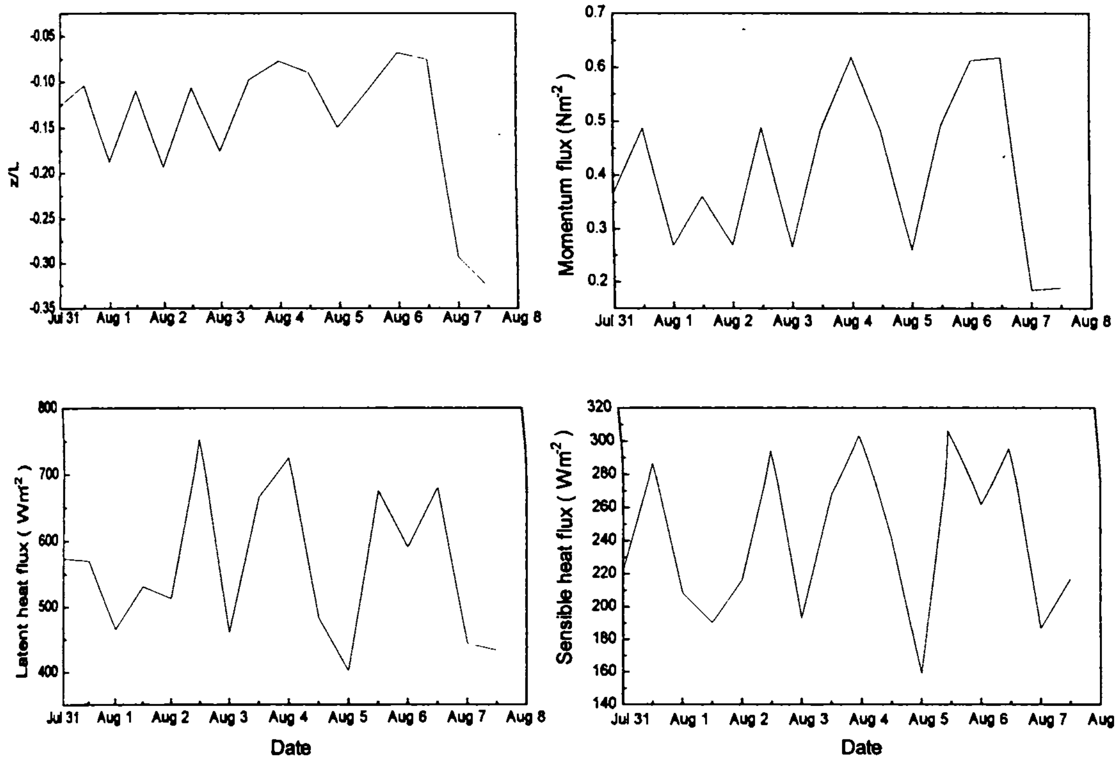

Fig. 2.10e Stability parameter and surface fluxes at Minicoy (Jul. 31 -Aug. 7 1988)

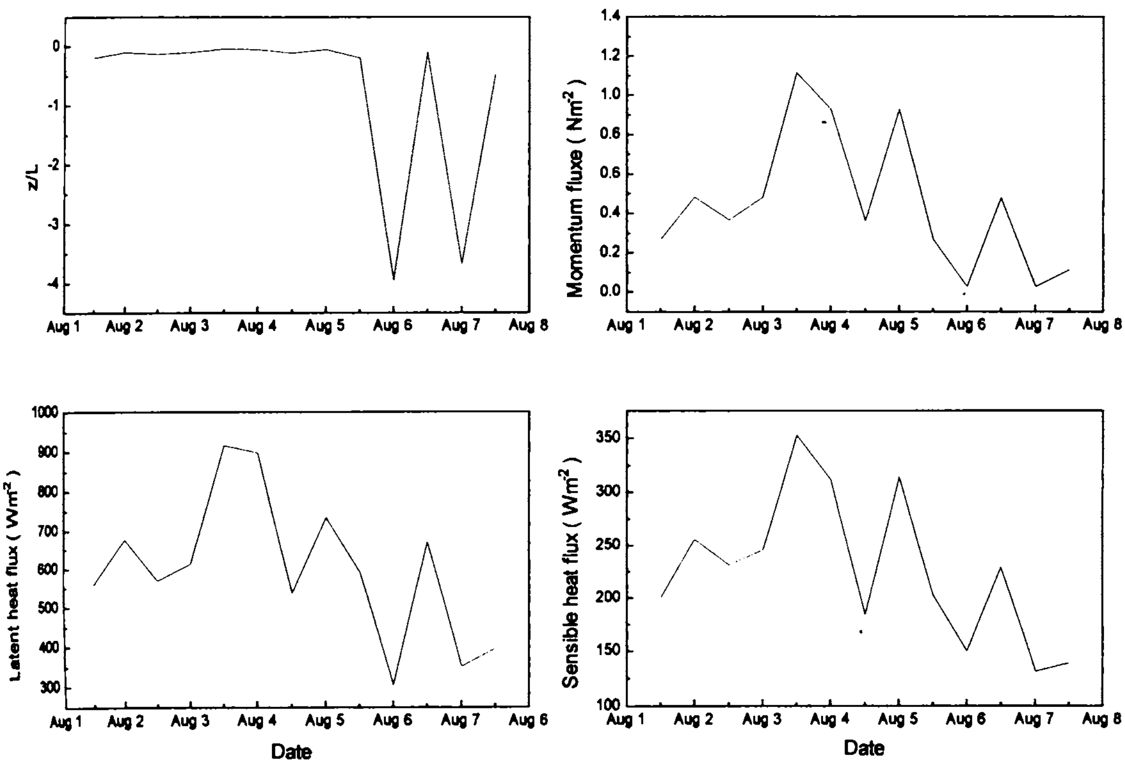

Fig. 2.10f Stability parameter and surface fluxes at Mangalore(Aug.1-7 1988)

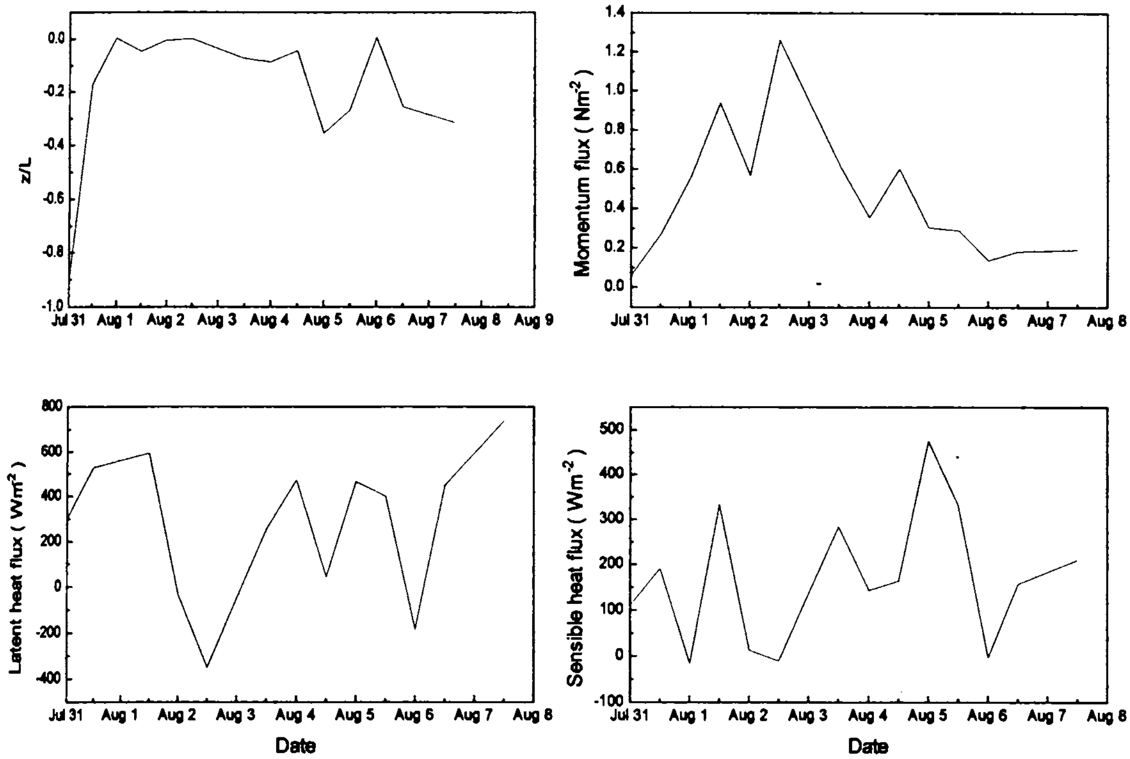

Fig. 2.10g Stability parameter and surface fluxes at Goa (Jul. 31-Aug. 7 1988)

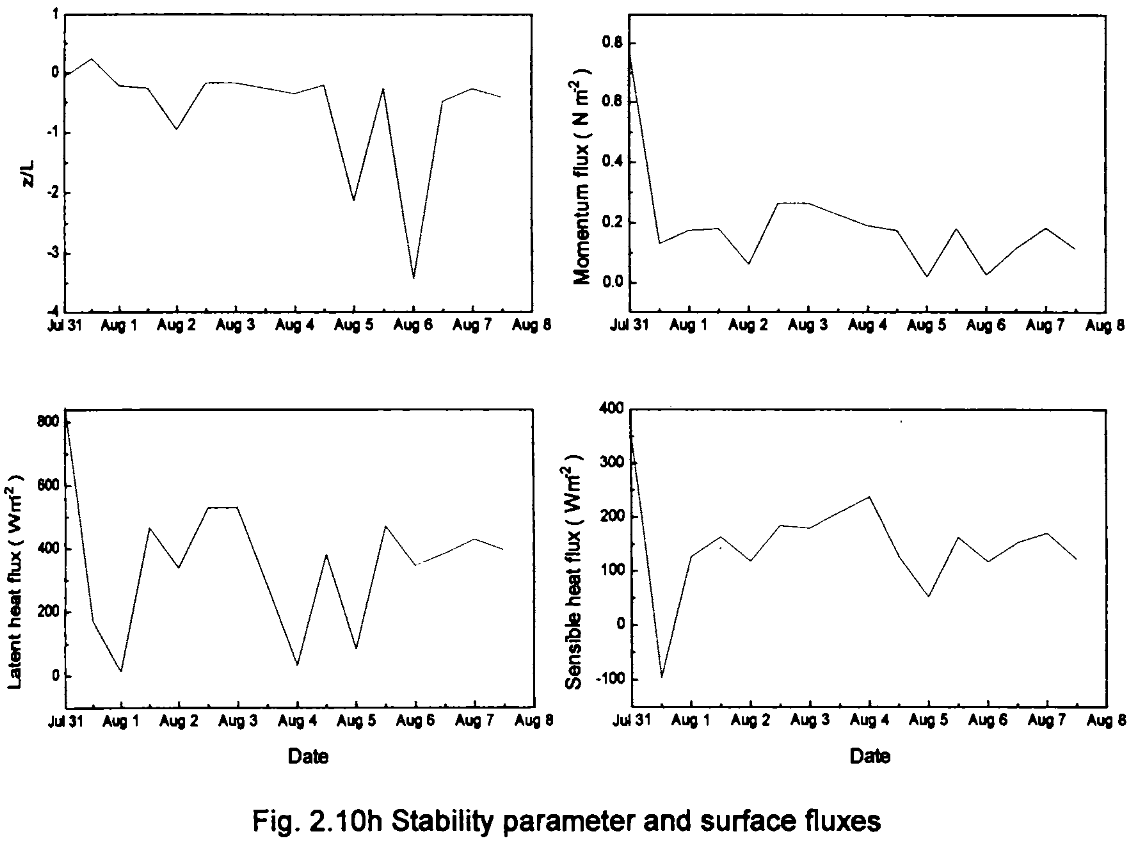

Fig. 2.10h Stability parameter and surface fluxes at Bombay (Jul. 31-Aug. 7 1998)

moist winds. This feature is noticed at all the stations considered. As the system decays or the effect gets reduced the atmosphere returns to highly unstable situation from the less unstable condition and transits between highly and less unstable conditions. Higher flux transport occurs during the depression period. Sivaramakrishnan et al (1996) and Murty et al (1996) noticed an increase in surface flux over the oceanic surface during the passage of a depression over the region. Therefore the surface layer features respond to the synoptic scale variations due to the depression formed over the head Bay and which moved over the subcontinent through a track along the monsoon trough. The disturbance is found to have an impact on the surface layer features not only near the region of its formation and along its path but also on the entire subcontinent.

Therefore the surface layer characteristic is found to vary during various seasons. Higher fluxes are noticed during the monsoon onset period due to increase in wind speed during the period. The stability condition shows near neutral or less unstable condition as the monsoon reaches a station. During the pre-monsoon and post-monsoon period the atmospheric condition is mostly unstable. The surface layer tends to become less unstable from the highly unstable condition by the onset and vice versa after the withdrawal of monsoon. The synoptic scale and mesoscale features and the surface layer characteristics are found to have close relation. An increase in surface fluxes is noticed during the occurrence of these disturbances. The surface layer is found to become highly unstable prior to the occurrence of these disturbances thus giving a prior indication of their formation. The occurrence of the synoptic scale disturbance affects the surface layer characteristics on the entire subcontinent as soon as it is formed.

CHAPTER 3

Drag coefficient during various seasons

The surface fluxes of momentum, heat and moisture are determined by several methods such as eddy correlation, profile, bulk aerodynamic method etc. The simplest and most widely used method is the bulk aerodynamic method. In this method the turbulent transport of momentum, heat and moisture are treated as proportional to wind speed, temperature and moisture between the underlying surface and a reference level. The proportionality constants known as bulk transfer coefficients are to be specified to obtain the turbulent transport. The bulk transfer coefficient for momentum transfer is called the drag coefficient C_D. No unique value of bulk transfer coefficients is given in the literature. These coefficients are estimated using the turbulence measurement made in field experiments. Many studies on the estimation of transfer coefficients are done using the experimental data over ocean and a few works over land region. Garratt (1977) has done a comprehensive review on the estimation of drag coefficients over oceans and continents and reported average values of neutral drag coefficient over tropical and extra tropical land surface on the basis of conservation of angular momentum. According to him the neutral value of drag coefficient over Asian landmass 20°N - 50°N is $3.9\times10^{-3}$and over south of 20°N is 27.7×10^{-3}. The dependence of drag coefficient on surface roughness, wind speed and stability has also been documented. The drag was found to increase when the surface roughness increases and when the atmospheric flow becomes unstable (Stull 1988).

Over the Indian region the pre-monsoon, onset and post-monsoon features has wide variability. The wind, temperature and humidity parameters show sharp variations during the various periods. Also there is a variation from region to region due to the varying topographical features of the region during these periods. These variations during the different seasons influence the boundary layer characteristics. The variations in surface winds and surface stability can cause changes in the surface drag. Also because of the wide variations in the surface roughness over different regions the surface drag also varies spatially. Mohanty et al (1995) obtained the drag coefficient at Jodhpur using a regression equation and found that the drag value is about 5.54×10^{-3} which is slightly greater than the neutral value given by Garratt (1977). The drag decreased with increase in wind speed and it increased with instability. Kusuma (1996) estimated the drag coefficient at Kharagpur and Jodhpur, which lies in the eastern and western end of the monsoon trough respectively. It was found that the drag coefficient at Kharagpur is smaller than that at Jodhpur for same range of wind speeds and the drag increased steeply when the wind speed was less than 2 ms^{-1}. The drag values obtained are higher than the neutral value given by Garratt (1977). In this study an attempt is made to obtain the drag coefficients over some selected regions during the pre-monsoon, onset and post-monsoon and their temporal and spatial variabilities are studied. The dependence of drag coefficient over the land surfaces on wind speed and stability is also investigated since no adequate information on its dependence is provided by the earlier studies.

Materials and Methods

The daily values of drag coefficient during April, May, June and October, November, December are obtained using the frictional velocity computed indirectly by the profile method. The computation is done for five years from 1984 to 1988 for three west coast stations Bombay, Mangalore and Trivandrum to study the variations in the drag coefficient during the pre-monsoon, onset and post-monsoon periods. The dependence of drag coefficient on the wind and stability was also studied for the three stations during the period. Computations are performed using the daily surface observations of wind and temperature taken at 00UTC provided by the India Meteorological Department.

The surface fluxes of momentum can be obtained using the profile method which is an indirect method based mainly on Monin- Obukhov similarity theory. The wind profile and temperature profile in the surface layer can be is described as (Dyer and Hicks 1970; Businger et al 1971)

$$\Delta \bar{u} = (u_* / k) (\ln z_2/z_1 - \psi_m (\zeta)) \qquad (3.1)$$

$$\Delta\ \bar{\theta} = (R\theta_{*} / k)\ (\ \ln z_2/z_1 - \psi_h\ (\zeta)\ ) \qquad (3.2)$$

where $\Delta u = u_2 - u_1$, u_1 and u_2 are winds at levels 1 and 2 respectively and $\Delta\theta = \theta_2 - \theta_1$, θ_1 and θ_2 are the temperatures at levels 1 and 2 respectively. u_* is the frictional velocity and θ_* is the temperature scale, ψ_m and ψ_h are the stability functions associated with wind and temperature respectively.

$\zeta = z/L$ where $z = z_2 - z_1$ is the difference in height of the two levels 1 and 2. z_2 is taken at 10m height which is the height at which surface winds are measured and z_1 is the roughness length z_o where u is zero. The roughness length is obtained following Delsol et al (1971) using the expression

$$z_o = 0.15 + 0.2(236.8 + 18.42h)^2 10^{-8} \qquad (3.3)$$

The temperature at the earth's surface is obtained by solving the energy balance equation and that at 10m is obtained by linearly interpolating the observed surface temperature to that level.

The surface energy balance can be expressed as

$$G_o = (1 - \alpha\ )\ Sw\downarrow + Lw\downarrow - \sigma T_s - F_s\uparrow - F_l\uparrow \qquad (3.4)$$

where $Sw\downarrow$ is incoming short wave radiation, $Lw\downarrow$ is longwave radiation flux into the ground, α the surface albedo which is taken as 0.15, σT_s is the outgoing longwave radiation where σ is the Stefan – Boltzman constant. F_s and F_l are the fluxes of sensible and latent heat respectively which are obtained using stability dependent bulk aerodynamic formula and G_o is the soil moisture flux. The surface temperature is obtained by solving the equation using the software by Krishnamurty(1986). The first guess for T_s is assumed to be the air temperature at the lowest observation level.

L is the Monin Obukhov length given by

$$L = \bar{T} u_*^2 / g k \theta_* \qquad (3.5)$$

R=0.74 is the ratio of eddy diffusivities in the neutral limit, k=0.4, the Von Karman constant.

The stability functions can be written as (Paulsen 1970, Barker and Baxter 1975)

In the unstable condition (ζ<0)

$$\psi_m(\zeta) = \ln((1+x^2)/2) + 2\ln((1+x)/2) - 2\arctan x + \pi/2 \qquad (3.6)$$

and

$$\psi_h(\zeta) = 2\ln((1+y)/2) \qquad (3.7)$$

where $x = (1-15\zeta)^{1/4}$ and $y = (1-9\zeta)^{1/2}$ (3.8)

In the stable condition ($\zeta > 0$)

$$\psi_m = -4.7\zeta \qquad (3.9)$$

and

$$\psi_h = -4.7\zeta / R \qquad (3.10)$$

The frictional velocity u_* and the temperature scale θ_* are computed iteratively. First a large value was assumed for L and u_* and θ_* are computed in neutral limits. L is then recomputed using the computed u_* and θ_* according to equation (3.5). Then u_* and θ_* are recomputed using equation (3.1) and (3.2) with the new L. Again L is recomputed with the recomputed u_* and θ_* and this process is repeated until L does not change in desired accuracy limits.

The surface momentum flux is computed as

$$\tau = \rho u_*^{\ 2} \tag{3.11}$$

Bulk aerodynamic method is the most conventional and widely used method for determining the surface fluxes. In this method it is assumed that the surface wind stress is in the direction of surface wind. The surface fluxes are given in terms of the differences in mean meteorological parameters at two lower most levels in the surface layer. The surface momentum flux is expressed as

$$\tau = \rho C_D u^{\ 2} \tag{3.12}$$

where C_D is the momentum transfer or drag coefficient and u is the wind speed at a reference height.

The drag coefficient can be determined from equation (3.11) and (3.12) as

$$C_D = (\ u_* / u)^2 \tag{3.13}$$

where u_* the frictional velocity is computed iteratively using equation (3.1). u is the wind speed at 10m height.

Results and discussion

The computed drag coefficient at Bombay during April, May and June and October, November and December for five years from 1984 to 1988 are given in fig. (3.1) and fig. (3.2) respectively. The drag values are found to be higher during pre-monsoon period which is between 2.4×10^{-2} to 2.8×10^{-2}. It gradually decreases by the monsoon onset and the values are found to be about $1\text{x}10^{-2}$. The drag coefficient therefore shows a decreasing trend by the onset of monsoon. After the effect of monsoon activity the drag coefficient shows an increasing trend during the post-monsoon period. The value which was between 2.2×10^{-2} to $2.8\text{x}10^{-2}$ during October gradually increased to more than 3×10^{-2} during most of the days and reached even upto $4\text{x}10^{-2}$ in certain days in December. The drag coefficient is higher during the post-monsoon period than

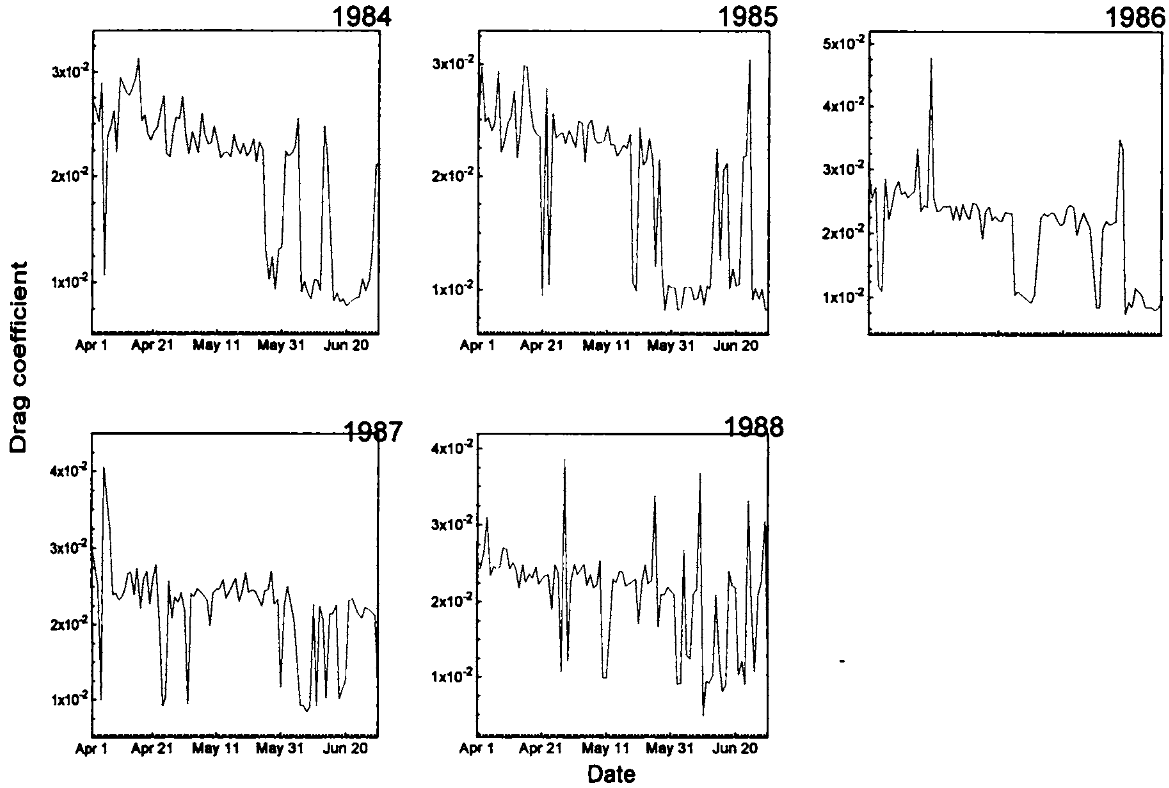

Fig. 3.1 Drag coefficient during April - June 1984 - 1988 at Bombay

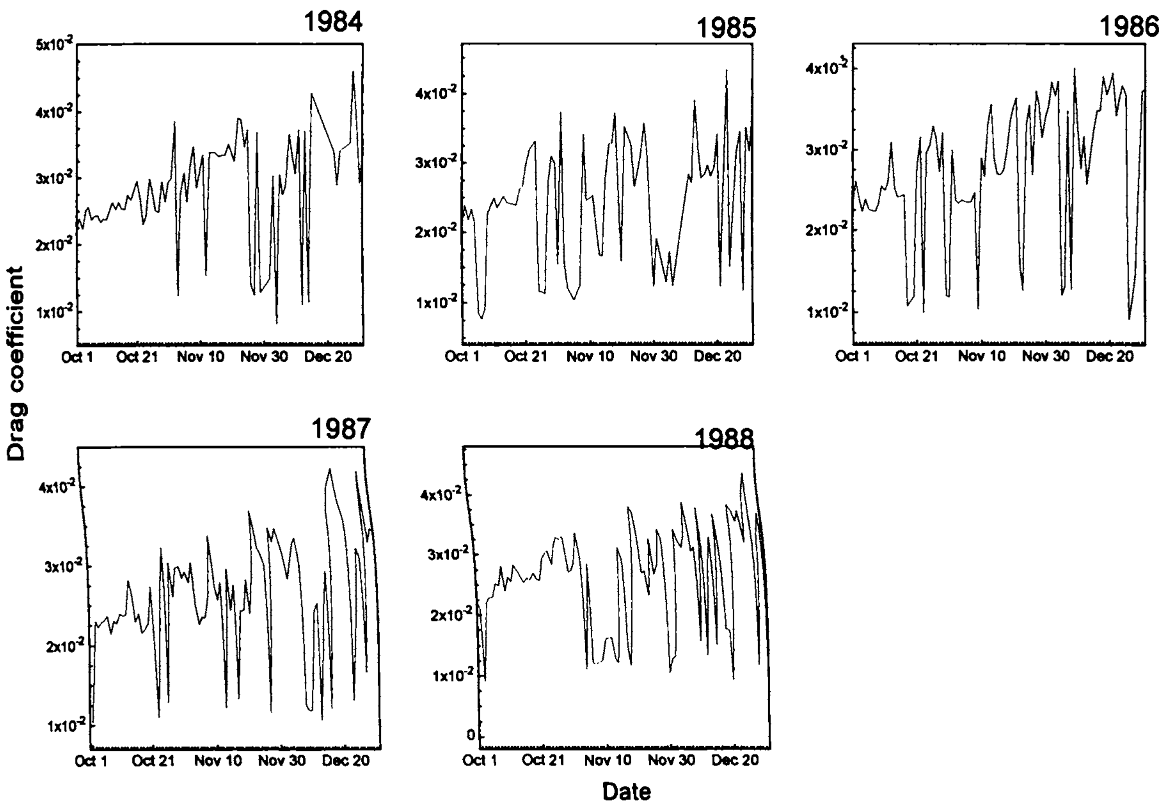

Fig. 3.2 Drag coefficient during Oct.- Dec.1984 - 1988 at Bombay

that during the pre-monsoon. The drag coefficient during the pre-monsoon and onset period at Mangalore for five years from 1984 to 1988 are given in fig. (3.3). The drag coefficient is found to be highly fluctuating at the station and it ranges between 1×10^{-2} to 2.5×10^{-2}. By the onset the drag coefficient is found to decrease and is found to be less than 1×10^{-2} during most of the days in June. Figure (3.4) shows the drag coefficient at Mangalore during October, November and December. The drag coefficient values are highly fluctuating during the post-monsoon period and are found to range between 0.7×10^{-2} to 2.5×10^{-2}. The drag coefficient during pre-monsoon and onset period at Trivandrum is given in fig. (3.5). During April the drag values are found to be high and are found to be about 2.2×10^{-2}. It decreases by the onset of monsoon and its value is found to be less than 0.8×10^{-2}. After the monsoon activity during the post-monsoon (fig. (3.6)), the drag coefficient is found to increase after October and is having maximum value greater than 2.2×10^{-2} in most of the days. Higher drag values are found during the post-monsoon period. The drag coefficient is higher at Bombay and lesser at Trivandrum during pre-monsoon and post-monsoon periods. But by onset the three stations are having almost same values. The monsoon activity is found to reduce the surface drag and along the west coast the drag is found to increase northwards.

Dependence of drag coefficient on wind speed

The wind speed at 10m level observed during April, May and June and during October, November and December from 1984 to 1988 at Bombay is shown in fig. (3.7) and fig. (3.8) respectively. The drag coefficient is found to decrease with increase in wind speed. The wind speed increases by the onset of monsoon and is very high during the monsoon. Therefore by the onset the drag coefficient decreases and reaches its minimum during the monsoon. In the northeast monsoon period also the drag decreases when the wind speed is higher. In figures ((3.9) and (3.10)) the wind speed during pre-monsoon and onset phase and northeast monsoon period at Mangalore are given. The wind speeds shows an increasing trend by the onset of monsoon. Following the increasing trend in the wind speed the drag coefficient shows a decreasing trend. After the monsoon the wind speed decreases. During the northeast monsoon high wind speed was noticed in association with depressions and low-pressure systems. The wind speed during pre-monsoon, onset and post-monsoon are higher at Mangalore than Bombay in most of the days. Hence the drag coefficient at Mangalore is less compared to Bombay.

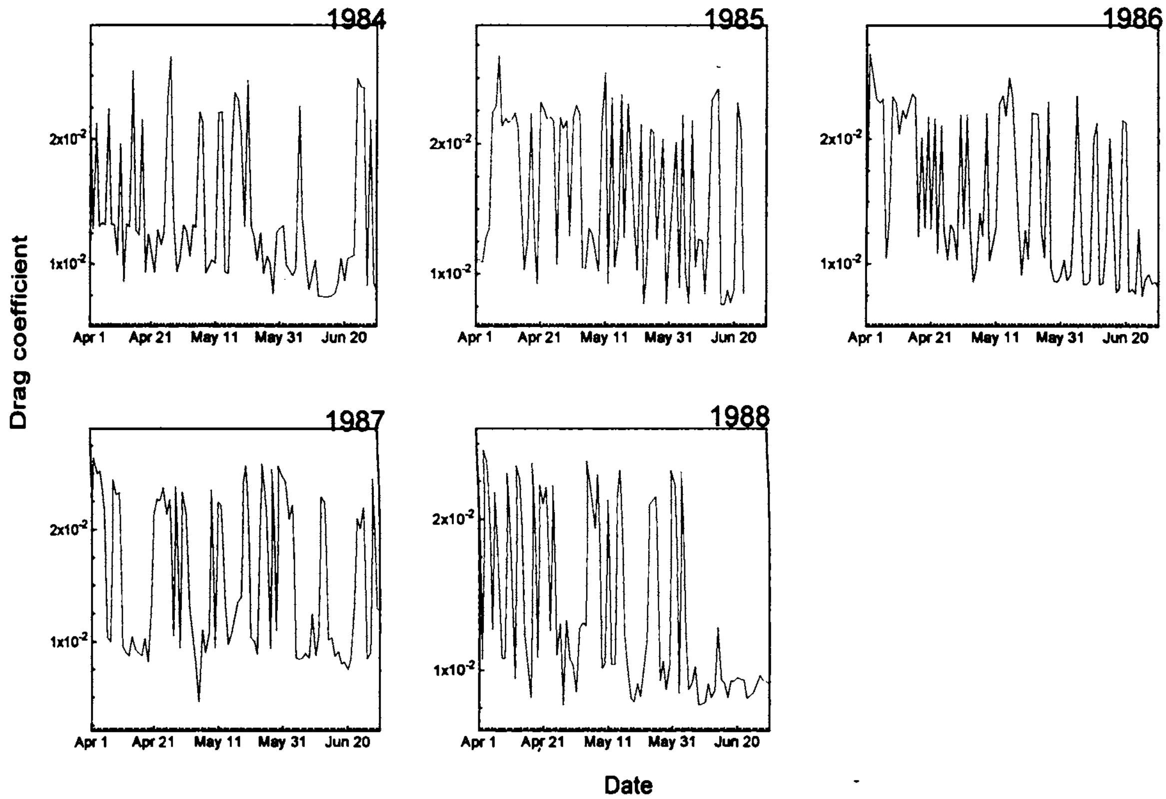

Fig. 3.3 Drag coefficient during April-June, 1984-1988 at Mangalore

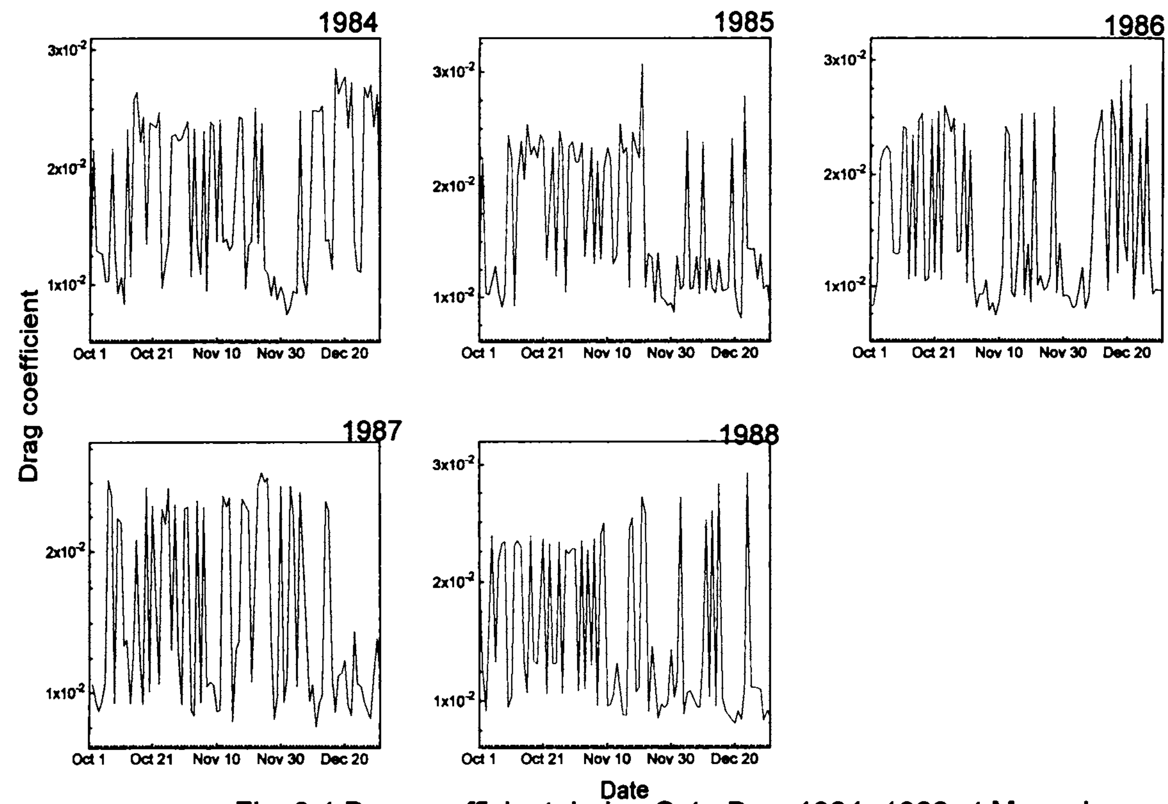

Fig. 3.4 Drag coefficient during Oct.- Dec.,1984 -1988 at Mangalore

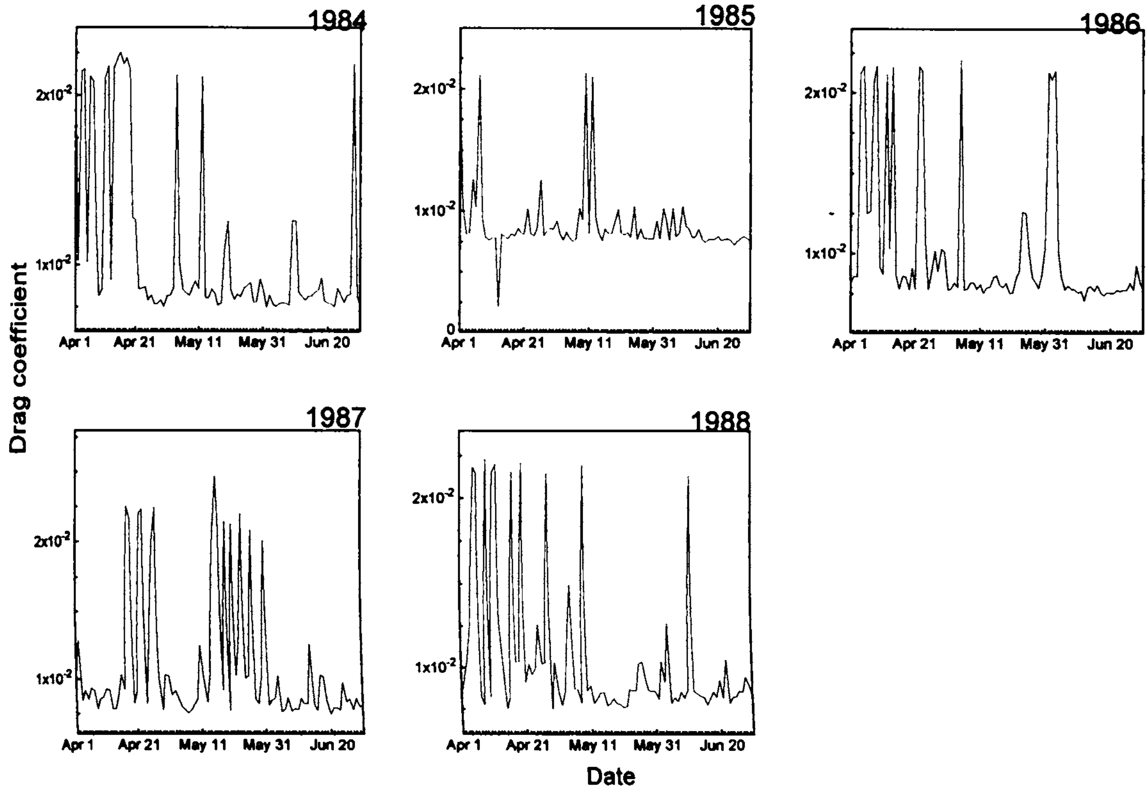

Fig. 3.5 Drag coefficient during April - June, 1984-1988 at Trivandrum

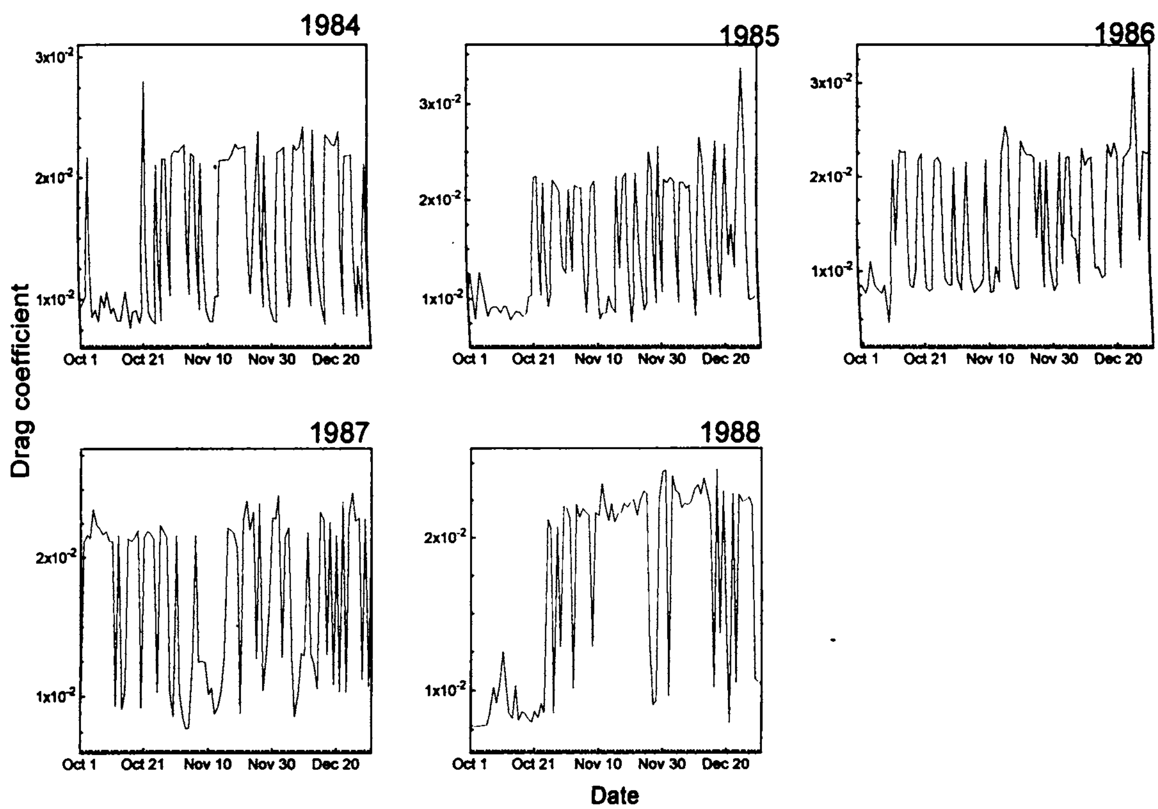

Fig. 3.6 Drag coefficient during Oct.-Dec.,1984 - 1988 at Trivandrum

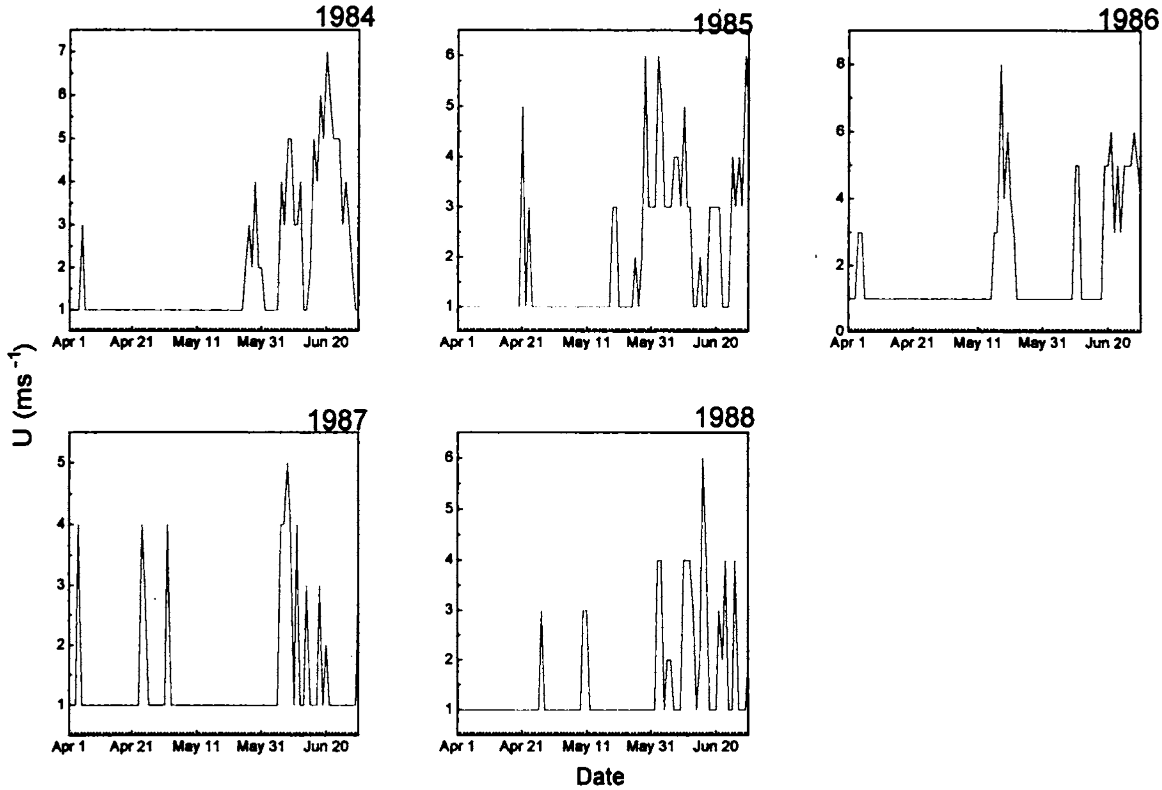

Fig. 3.7 Wind speed during April-June,1984 - 1988 at Bombay

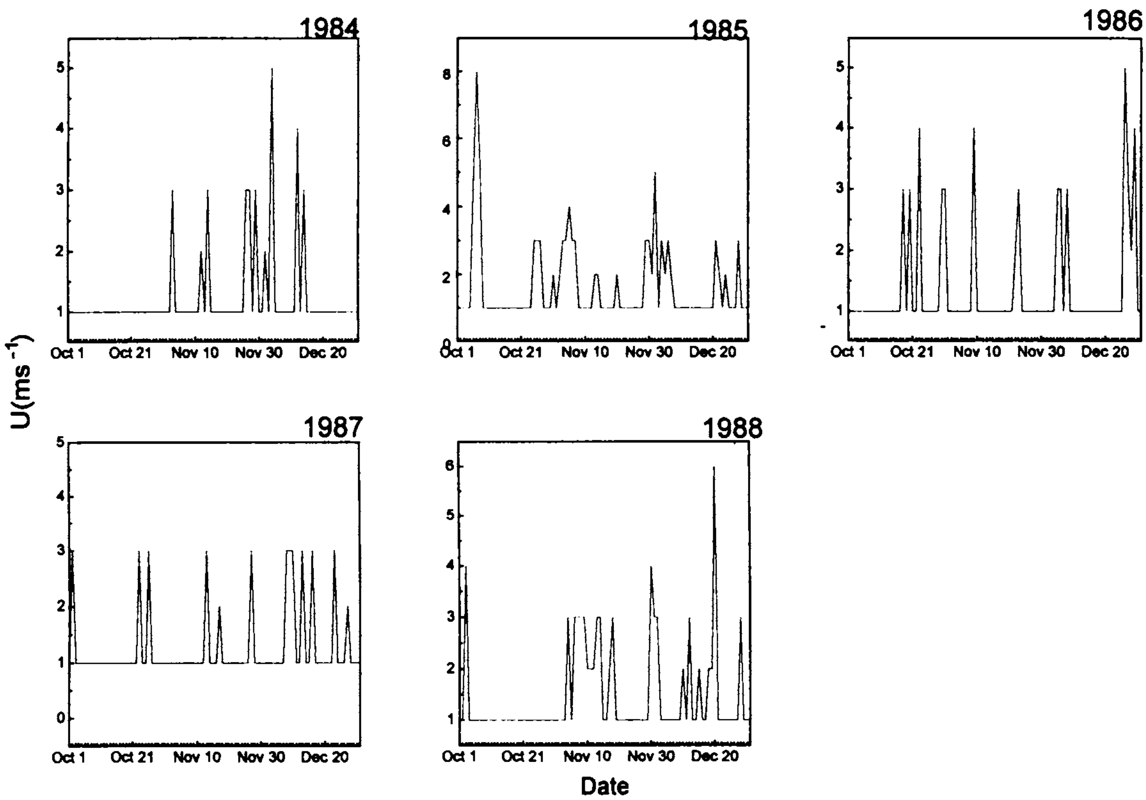

Fig. 3.8 Wind speed during Oct.-Dec., 1984 - 1988 at Bombay

At Trivandrum (figures (3.11 and 3.12)), the wind speed increased by May and it further increased by the onset of monsoon. With the increase of wind speed the drag coefficients gradually decreased and become much less by the onset of monsoon. By the withdrawal of monsoon wind speed decreased and following the decrease the drag coefficient is found to increase. Wind speed observed at Trivandrum was higher during pre-monsoon and monsoon than that observed at Mangalore. Hence the drag coefficient during the period was much lesser than that at Mangalore. The wind speed at 10m observed during the pre-monsoon, onset and northeast monsoon for five years was plotted against the drag coefficient computed for Bombay, Mangalore and Trivandrum and is shown in fig. (3.13). From the figure the drag coefficient C_D shows a steep variation when the wind speed becomes less than 4 ms^{-1}. When the wind speed is higher than $4ms^{-1}$ the drag coefficient is found to vary gradually and the variation is very small. At Bombay when the wind speed is less than 2 ms^{-1} C_D is found to have a wider range than the other two stations. By using the power law the variation of drag coefficient with wind at Bombay is given by $C_D = 0.0237\ U_{10}^{-0.686}$ and at Mangalore the variation is given by $C_D = 0.0213\ U_{10}^{-0.577}$ and at Trivandrum $C_D = 0.0189\ U_{10}^{-0.453}$ where U_{10} is the wind speed observed at 10m level.

Dependence of drag coefficient on the surface layer stability

The drag coefficient is found to show an increasing trend with increasing instability. The effect of stability is obtained using the stability parameter $\zeta = z/L$. The surface layer which was highly unstable during the pre-monsoon period tends to become less unstable or attains neutrality by the onset of monsoon and it returns to more unstable situation after the withdrawal of monsoon at Bombay (figures (3.14 and 3.15)). Comparing the drag coefficient and stability parameter during April, May and June it can be seen that the drag coefficient was higher during the highly unstable pre-monsoon period and it decreases by the onset when the atmosphere is tending to become less unstable. During the northeast monsoon period the surface layer of the atmosphere returns to highly unstable condition from the less unstable condition during the monsoon season and following the atmospheric condition the drag coefficient is found to increase. At Mangalore also the drag coefficient follows the stability condition of the atmosphere (figures (3.16 and 3.17)). The drag coefficient decreases by the monsoon onset during

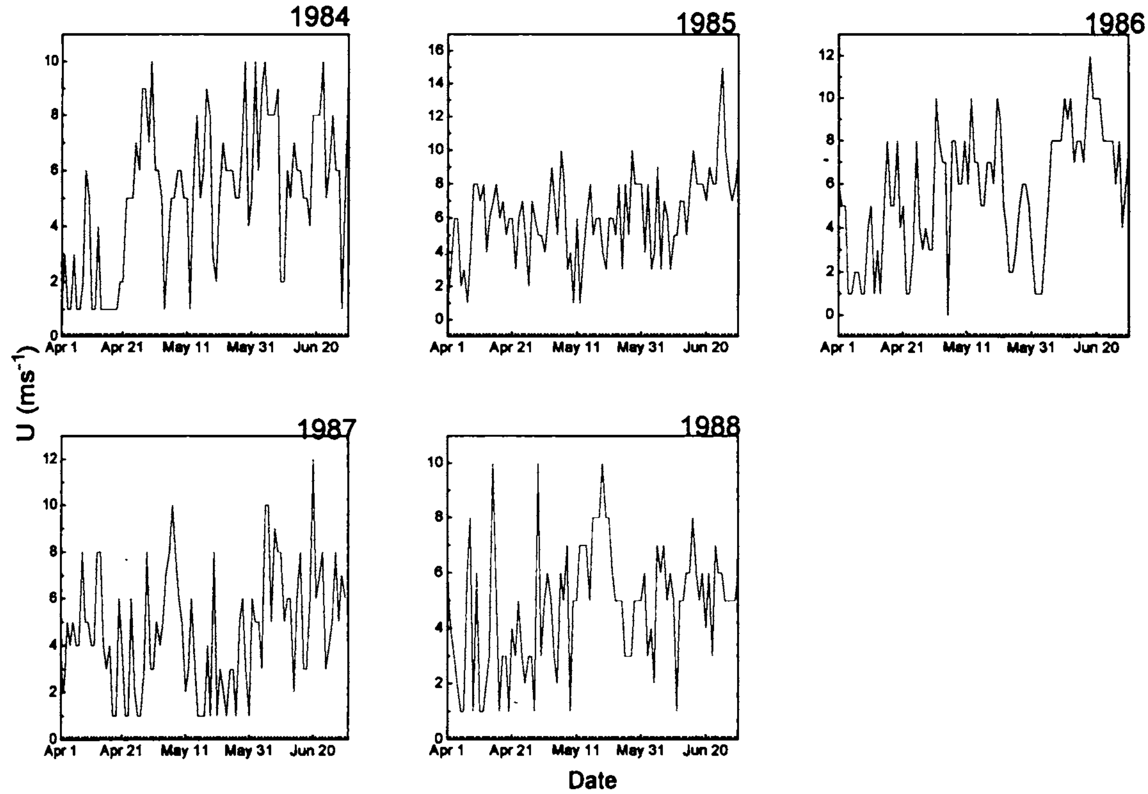

Fig. 3.11 Wind speed during April-June,1984 - 1988 at Trivandrum

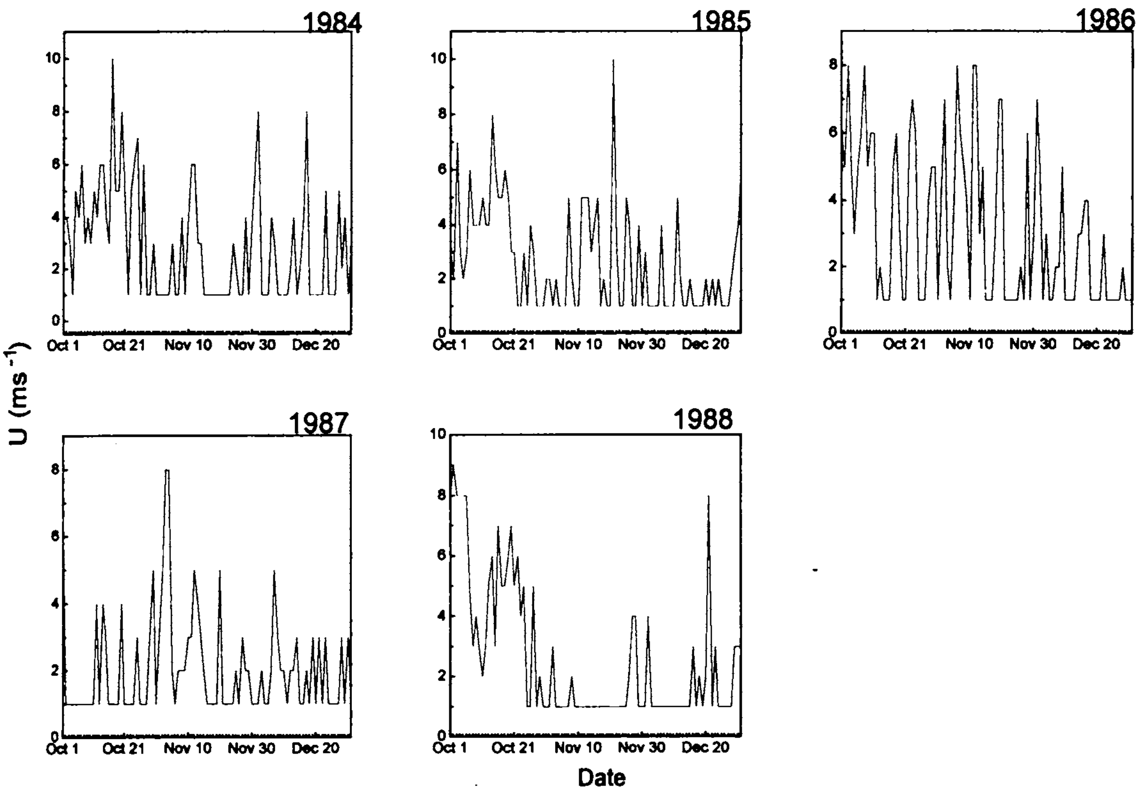

Fig. 3.12 Wind speed during Oct.- Dec., 1984-1988 at Trivandrum

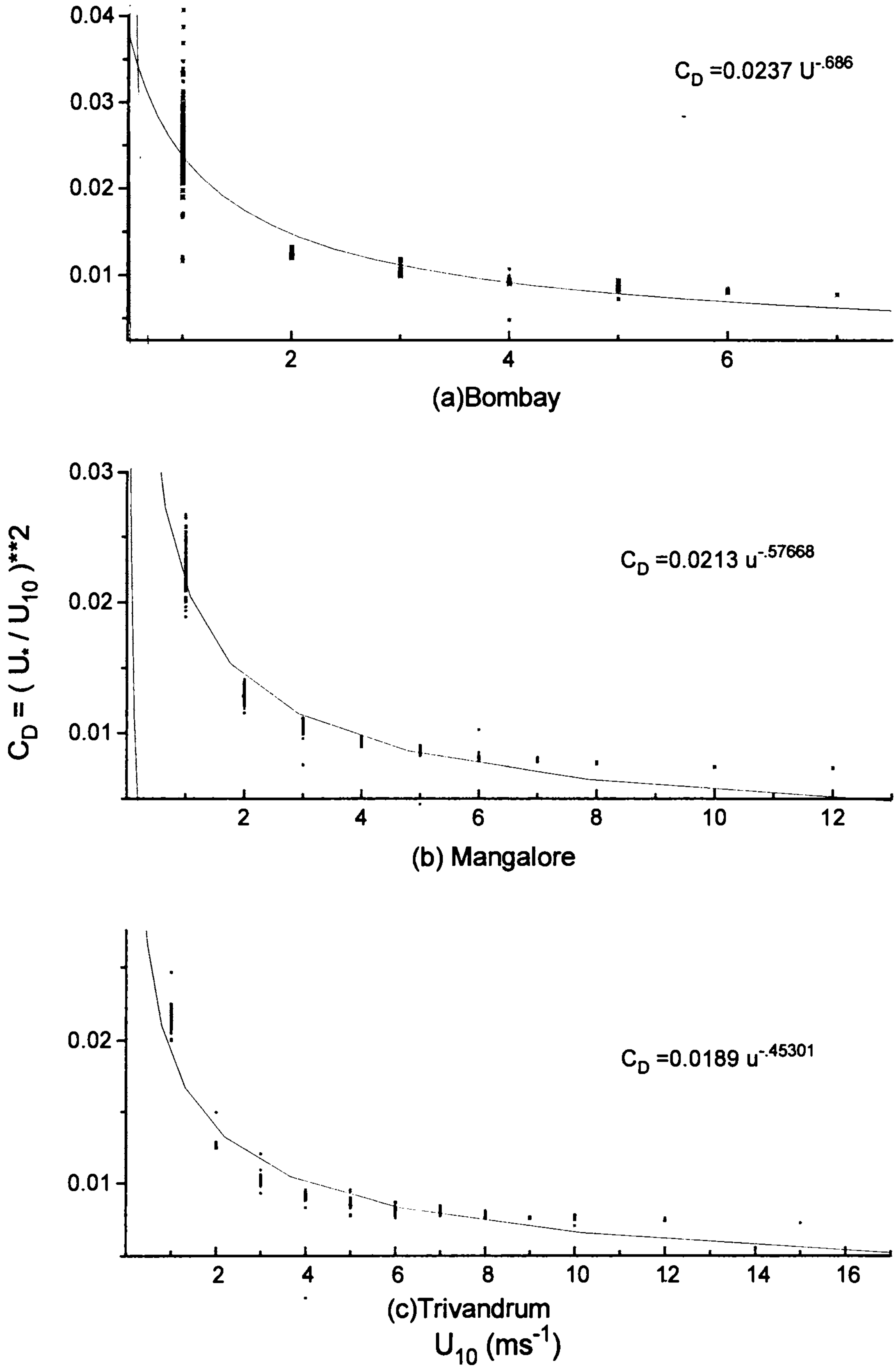

Fig. 3.13 Variation of C_D with respect to wind for
a) Bombay, b) Mangalore and c) Trivandrum

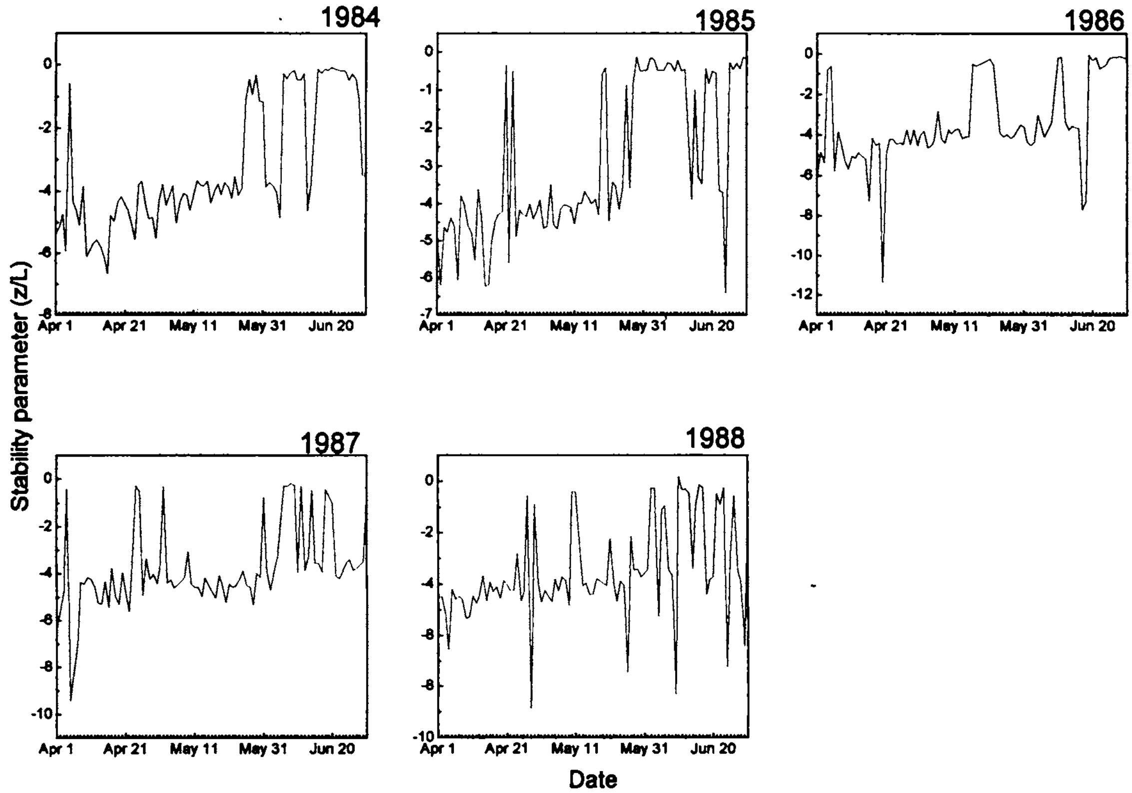

Fig. 3.14 Stability parameter during April-June, 1984-1988 at Bombay

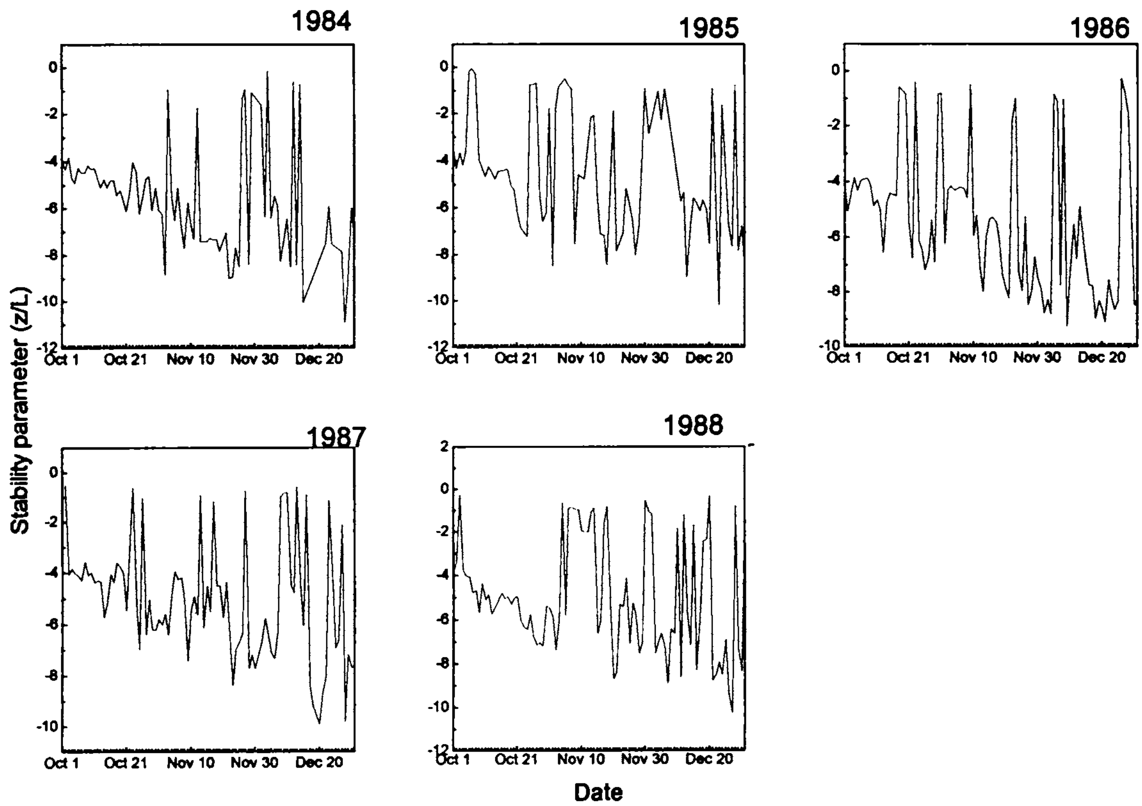

Fig. 3.15 Stability parameter during Oct.-Dec.1984-1988 at Bombay

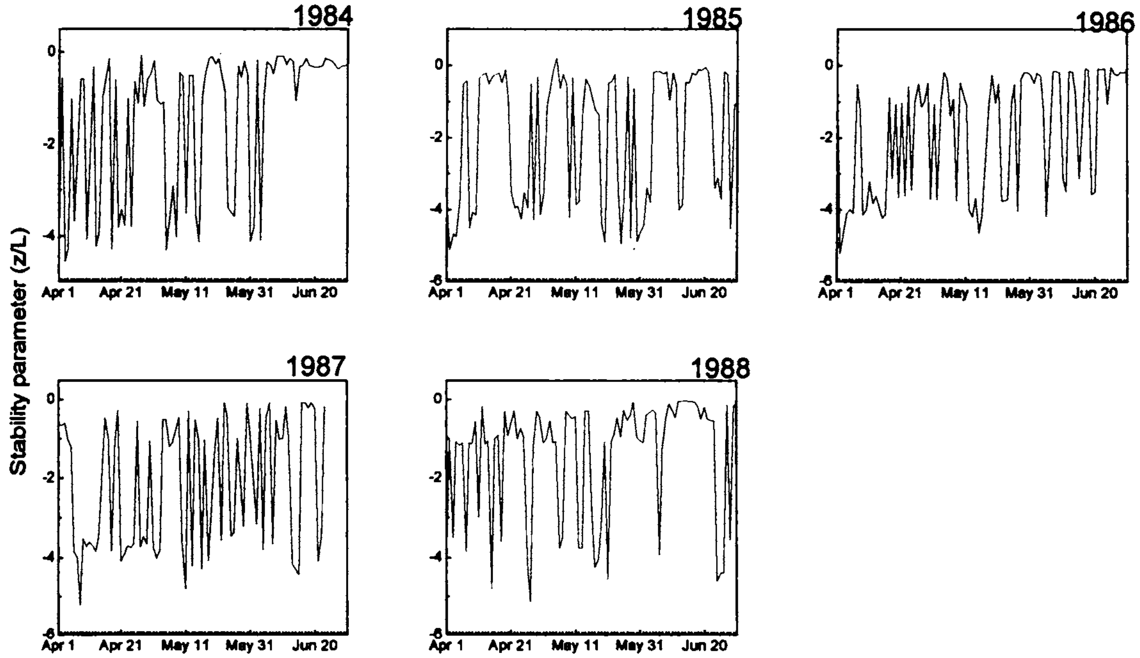

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Fig. 3.16 Stability parameter during April-June, 1984 -1988 at Mangalore

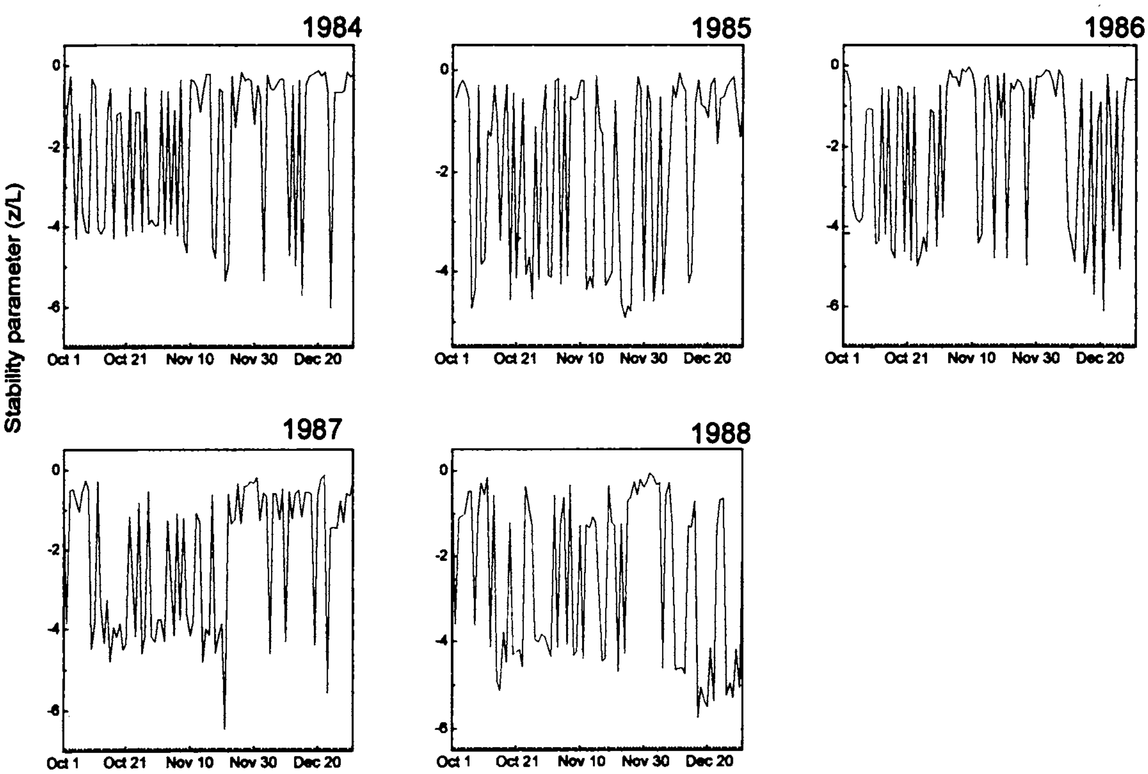

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Fig. 3.17 Stability parameter during October-December, 1984 -1988 at Mangalore

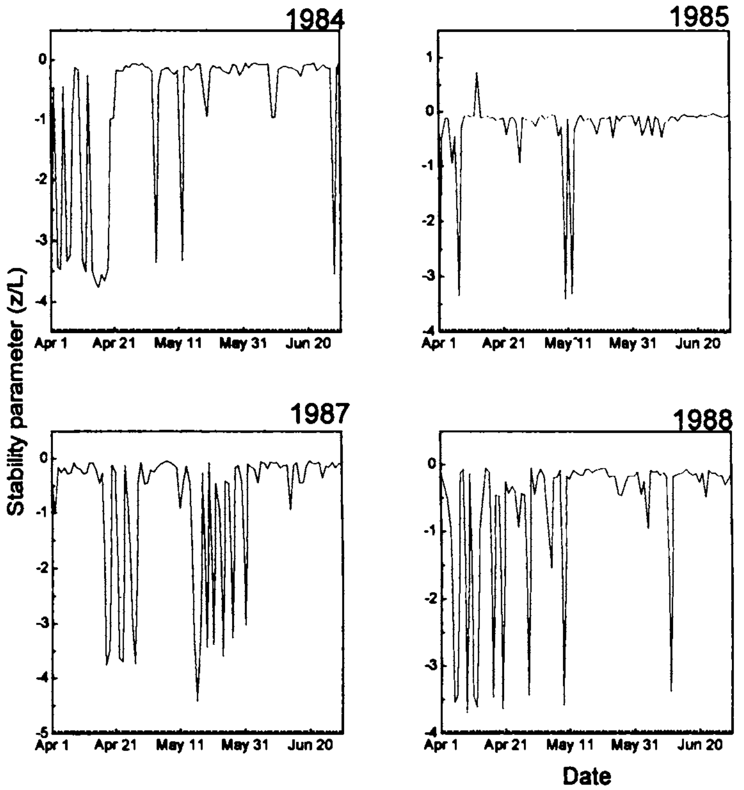


Fig. 3.18 Stability parameter during April-June,1984 - 1988 at Trivandrum

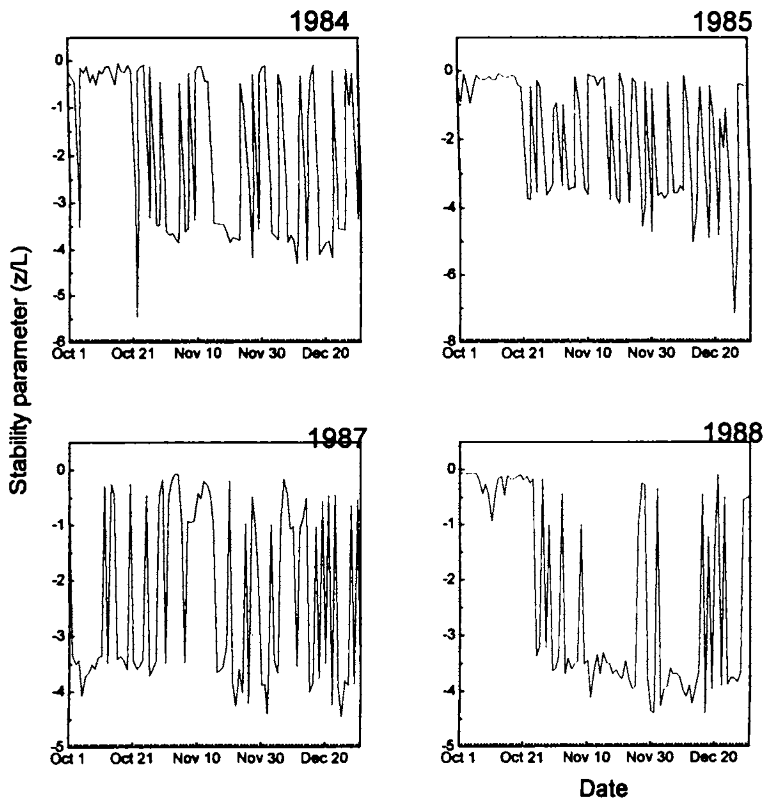


Fig. 3.19 Stability parameter during Oct.-Dec., 1984 - 1988 at Trivandrum

which the atmospheric surface layer is less unstable when compared to that during the pre-monsoon period. The surface layer becomes highly unstable most of the days during the northeast monsoon period and the drag coefficient is found to be higher during the highly unstable state of the atmosphere. The surface layer at Trivandrum is found to be less unstable or neutral much before the onset of monsoon (fig. (3.18)). The drag coefficient is also found to be smaller during May, which is much before the onset. The drag coefficient gradually increased following the increasing instability at the station after the withdrawal of monsoon. The surface layer returns to highly unstable situation during the northeast monsoon period (fig. (3.19)) and the drag coefficient is found to increase during the period. Therefore there is a positive correlation between the drag coefficient and instability of the atmosphere.

Thus the drag coefficient is found to decrease with increasing wind speed and decreasing instability. The surface drag is higher during the pre-monsoon and post-monsoon periods when compared to that during the onset and monsoon period. Steep increase in drag occurs when the wind speed becomes less than $4ms^{-1}$. According to Garratt (1977), to the south of 20^oN in Asian continent the neutral drag coefficient is $27.7x10^{-3}$. The three stations considered lies to the south of 20^oN. The obtained drag coefficients for the three stations during less unstable or neutral condition are lesser than the neutral value given by Garratt. They may reach near the neutral drag value during the highly unstable condition of the atmosphere. There is a slight decreasing trend in the drag value from the north to the south along the west coast.

CHAPTER 4

Marine Boundary Layer characteristics

The surface boundary layer over oceans play an important role in the exchange of energy, mass and momentum across the water surface, which influences to a large extent the atmospheric and oceanic circulations. The energy of the major rain producing systems in the tropics like monsoon and tropical cyclones are derived from these exchange of properties across the air-sea interface. Transfer of properties across the air-sea interface is by both molecular and eddy exchanges. Molecular transfer dominates near the sea surface whereas exchange is by eddies above the molecular sub layer.

Most important characteristic of the sea surface is its temperature. Due to large heat capacity and efficient mixing in the upper oceanic mixed layer the oceanic surfaces are characterised by temporal and spatial homogeneity of temperature. The meteorological factors affecting sea surface temperature are the net radiation to or from the sea surface, evaporation and precipitation and sensible heat exchange with the atmosphere. The radiation balance near the sea surface is much complicated. The solar radiation received at the surface can penetrate into the water to considerable depth and the albedo of the surface is dependent on solar altitude. The long wave radiation from the atmosphere is also absorbed by the thin surface film of water, which also gives out radiation to the atmosphere. The net long wave radiation varies according to presence of clouds and absorption and emission by water vapour.

Latent heat flux is the major component of energy balance at the sea surface. It is the major source of energy that drives the atmosphere. The latent heat flux is usually one order larger than sensible heat flux. The sensible heat flux across the air-sea interface is much smaller than latent heat flux and radiative fluxes. It becomes significant when the air-sea temperature difference is higher. It is difficult to obtain the heat exchange *through water, which is by turbulent mixing, currents and upwelling and downwelling* motions. Therefore the energy balance near sea surface can be considered as

$$R_N = H + H_L + \Delta H_S \qquad (4.1)$$

where R_N is net radiation, H the sensible heat flux, H_L the latent heat flux, and ΔH_S the heat storage which can be obtained from temperature measurements at several depths in the mixed layer.

The transfer of momentum from atmosphere to the ocean at their interface is the driving force for ocean currents and turbulence in the upper layers of ocean and is responsible for generation of surface waves. The total momentum exchange can be measured from the atmospheric surface layer well above the height of highest wave crests. In the fully developed surface layer the logarithmic wind profile is valid under neutral conditions and the Monin-Obukhov similarity relation are applicable under general stratified conditions. However the roughness parameter z_o determined using logarithmic profile law is found to depend on wind and wave and can vary over a very wide range (10^{-6} to 10^{-2}m). The surface drag coefficient C_D even though related to the much varying roughness parameter z_o is found to have smaller range of variation (0.001-0.003) over oceans and the surface stress is directly related to it irrespective of wind profile. The drag coefficient can be directly determined from momentum flux measurements and mean wind speed at one level while to estimate z_o requires accurate wind profile measurements. Therefore C_D is preferred over z_o by meteorologists and oceanographers.

Characteristics of the surface marine atmospheric boundary layer are covered by Roll (1965), Gill (1982) and Arya (1988). Most field experiments for investigating turbulent processes in the marine atmospheric boundary layer in the midlatitudes are over the tropical Pacific and Atlantic oceans and extensive literature is available on the marine boundary layer as a result of these experiments. The first field experiment over the Indian seas was the International Indian Ocean Experiment (IIOE) in 1964. The observations from the experiment established the existence of low level atmospheric inversion over the west and central Arabian Sea that disappeared as the monsoon current approached the west coast of India (Colon 1964; Sikka and Mathur 1965). The boundary layer over Arabian Sea was further investigated by Pant (1978,1982) and Ramanathan (1978) using the Indo-Soviet Monsoon Experiment ISMEX-78. Monsoon-77, Monsoon-88 and MONEX-79 gave further insight into the Arabian Sea marine boundary layer. Mohanty and Mohankumar (1990) studied the link between energy fluxes over Arabian

Sea and monsoon activity over Indian subcontinent using MONEX-79 data and found that surface fluxes increases during onset causing an oceanic heat loss which produces a positive feed back for maintenance of deep cumulus convection in the atmosphere. Monsoon Trough Boundary Layer Experiment-90 (MONTBLEX-90) was conducted to understand the variability of the atmospheric as well as marine boundary layer in relation to monsoon trough over land and northern Bay of Bengal. Murty et al (1996) found that active air-sea interaction with positive feed back from sea to air prevailed during development of deep depression and the heat loss is utilised to enhance the latent heat flux. Warm core eddies when moved away causes a net heat loss. The surface fluxes are found to increase during depressions over Bay of Bengal and decrease as it passes away (Singh 1992, Sivaramakrishnan et al 1996). These studies were done using the MONTBLEX-90 data over the Bay of Bengal region.

The observations over Bay of Bengal during MONTBLEX-90 are done mainly to study the variabilities in the marine atmospheric boundary layer in relation to the monsoon trough. Recently a field experiment called the Bay of Bengal and Monsoon Experiment (BOBMEX) was carried out in the Bay of Bengal region under the Indian Climate Research Program (ICRP). The experiment was done to study the air-sea interactions and intraseasonal oscillations in the Bay of Bengal during the summer monsoon using new types of measurements from ships and met-ocean buoys. Prior to the main BOBMEX in 1999 a Pilot experiment was carried out from October 23rd to November 8th 1998. The main experiment was conducted in two phases. The first phase was from July 17 to August 7 1999 and the second phase during August 12-30,1999. In the Pilot and main experiment direct air-sea fluxes using fast response sensors, surface meteorological parameters including radiation and rain, upper atmospheric parameters using radiosondes/minisondes, upper ocean temperature, salinity, current profiles and turbidity are measured. In this study the surface flux characteristics of the marine atmospheric boundary layer over Bay of Bengal during northeast monsoon are examined using the observations taken during the Pilot experiment period.

Synoptic situation

The Pilot experiment (October 23-November 8) was done during the northeast monsoon season. During the experimental period a low pressure was formed on October

26^{th} over southeast Bay. It intensified into a depression by evening. The depression moved in a northwesterly direction and intensified into a deep depression on 27^{th}. It weakened into a low pressure on 29^{th} evening over southwest Bay off north Tamilnadu coast and became less marked by 31^{st}. The depression track was shown in fig. (4.1). Another low pressure was formed over northwest Bay of north Tamilnadu on Nov.2. It persisted there and weakened on 7^{th}.

Materials and methods

The surface fluxes and stability parameter are computed for the Pilot experiment period using the surface ship observations taken during the periods. The momentum, sensible and latent heat fluxes are computed following the Monin-Obukhov similarity relations and Bulk aerodynamic method. Following Dyer and Hicks (1970) and Businger et al (1971) the wind, temperature and specific humidity profiles in the surface layer of the atmosphere can be described as

$$\Delta u = (u_* / k) (\ln z_2/z_1 - \psi_m (\zeta)) \qquad (4.2)$$

$$\Delta \theta = (R\theta_* / k) (\ln z_2/z_1 - \psi_h (\zeta)) \qquad (4.3)$$

$$\Delta q = (Rq_* / k) (\ln z_2/z_1 - \psi_h (\zeta)) \qquad (4.4)$$

where $\Delta u = u_2 - u_1$, u_1 and u_2 are wind speed at levels 1 and 2 respectively, $\Delta\theta = \theta_2 - \theta_1$, is the difference in temperature at levels 1 and 2 respectively and $\Delta q = q_2 - q_1$, is the difference in specific humidity at levels 1 and 2 respectively. u_* is the frictional velocity, θ_* is the temperature scale and q_* is the humidity scale. ψ_m and ψ_h are the stability functions associated with wind and temperature respectively.

$\zeta = z/L$ is the stability parameter, where $z = z_2 - z_1$ is the difference in height of the two levels 1 and 2. z_2 is the height at which surface winds are measured and z_1 is the roughness length z_o where u is zero. The roughness length over the oceanic region is obtained following Charnock (1955) as

$$z_0 = \alpha\, u_*^2/g \qquad (4.5)$$

where α is an empirical constant. The Charnock's formula was verified for large values of z_0 and u_*^2/g using experimental set of wind profile and drag data by Wu (1980) and α is obtained as 0.0185. In this study also α is taken as 0.0185.

L is the Monin Obukhov length given by

$$L = \bar{T}\, u_*^2 / g\, k\, \theta_* \qquad (4.6)$$

R=0.74 is the ratio of eddy diffusivities in the neutral limit, k=0.4, the Von Karman constant.

The stability functions can be written as (Paulsen 1970, Barker and Baxter 1975)

In the unstable condition (ζ<0)

$$\psi_m(\zeta) = \ln\left((1+x^2)/2\right) + 2\ln\left((1+x)/2\right) - 2\arctan x + \pi/2 \qquad (4.7)$$

and

$$\psi_h(\zeta) = 2\ln\left((1+y)/2\right) \qquad (4.8)$$

$$\text{where } x = (1-15\zeta)^{1/4} \text{ and } y = (1-9\zeta)^{1/2} \qquad (4.9)$$

In stable conditions ($\zeta > 0$)

$$\psi_m = -4.7\zeta \qquad (4.10)$$

and

$$\psi_h = -4.7\zeta / R \qquad (4.11)$$

Wind speed u_1 at z_1 is taken as zero and θ_1 as sea surface temperature and q_1 is the saturation specific humidity at sea surface. u_2, θ_2, q_2 are wind speed, temperature and specific humidity respectively measured at 11m level from the sea surface.

The frictional velocity u_*, temperature scale θ_* and humidity scale q_* are computed iteratively. First a large value was assumed for L and u_* θ_* and q_* are computed in neutral limits. L is then recomputed using the computed u_*, and θ_* according to equation (4.6). Then u_*, θ_* and q_* are recomputed using equation (4.2), (4.3) and (4.4) with the new L. Again L is recomputed with the recomputed u_* and θ_* and this process is repeated until L does not change in desired accuracy limits. The convergence limit is kept at the fifth decimal.

The surface momentum flux is computed as

$$\tau = \rho u_*^{\,2} \qquad (4.12)$$

Surface sensible heat flux is given by

$$H = -\rho C_p u_* \theta_* \qquad (4.13)$$

And surface latent heat flux can be expressed as

$$E = -\rho L_e u_* q_* \qquad (4.14)$$

where ρ is the density of air which does not have much variation in the layer considered and is taken as 1.25×10^3 kgm^{-3}, C_P is the coefficient of specific heat capacity at constant pressure and L_e is the latent heat of evaporation/condensation.

Bulk aerodynamic method: - This is the simplest and most widely used method for determining the surface fluxes. In this method it is assumed that the surface wind stress is in the direction of surface wind. The surface fluxes are given as the difference in mean meteorological parameters at two lower most levels in the surface layer.

Surface momentum flux is given by

$$\tau = \rho C_D U^2 \qquad (4.15)$$

Surface sensible heat flux is

$$H = -\rho C_H U(\theta_2 - \theta_1) \qquad (4.16)$$

and surface latent heat flux is

$$E = -\rho C_E U(q_2 - q_1) \qquad (4.17)$$

where ρ is density of air which is taken as a constant as in profile method. C_D is the drag coefficient. C_H is the transfer coefficient for heat or Stanton number and C_E is the transfer coefficient for moisture or Dalton number. Since these exchange coefficients does not have much variations over oceanic surfaces they are treated as constants. C_D is taken as 1.1×10^{-3} and $C_H = C_E = 1.5 \text{ X } 10^{-3}$. U, θ_2 and q_2 are wind speed, temperature and specific humidity at 11m level and θ_1 and q_1 are sea surface temperature and saturation specific humidity at sea surface respectively.

Drag coefficient can be obtained from the computed frictional velocity u_* using profile method and observed surface wind using the following equation

$$C_D = (u_*/u)^2 \qquad (4.18)$$

The fast response data collected at a sampling rate of 10 Hz acquired for 65 minutes once every three hours (00, 03, 06, 09,12, 15, 18 and 21 hours IST), which comprises of wind velocity at two levels, air temperature, humidity and ship acceleration obtained from the fast response sensors for November 1 and 3 during the pilot experiment period was used for a comparison of surface flux computations using the profile method and bulk aerodynamic method with the eddy correlation method. Also the variation in surface layer characteristics between the two days are looked into. November 1 was a rainy day, the rainfall as a result of deep convection from noon. The total rainfall was 19mm and winds were generally high. On November 3 there was an early morning shower and cloud free morning. Thick clouds were noticed in late afternoon. Total rain was about 2mm in two spells, one around 4 am and another late in the afternoon and both were from individual cumulonimbus clouds. Winds were low in the morning and the speed picked up in the evening.

Using the eddy correlation method the turbulent exchanges of momentum, heat and moisture are obtained directly with fast response sensors. The surface momentum, sensible heat and latent heat fluxes are computed using the equation (4.12), (4.13) and (4.14) respectively and the Monin-Obukhov length is computed using equation (4.6).

$$\text{where} \quad u_* = (\overline{u'w'}^2 + \overline{v'w'}^2)^{1/4} \tag{4.19}$$

$$\theta_* = -\overline{w'\theta'}/u_* \tag{4.20}$$

$$\text{and} \quad q_* = -\overline{w'q'}/u_* \tag{4.21}$$

The computations are carried out using fast response wind speed, temperature and specific humidity data at levels 2 (11 m) and 3(12.5 m). Anemometer data is obtained in the three principal axes. The data set includes horizontal velocity components due to ship motion. A correction is applied on the observed horizontal wind components by considering the horizontal velocity of the ship, computed at every second with the knowledge of ship position data obtained from the Global Positioning System (GPS). The acceleration due to the pitching and rolling motions of the ship is obtained using the accelerometers mounted along three directions (normal to ship track, along the ship track and in the vertical) on the sonic anemometer and appropriate corrections on the velocity components are made. The data acquired for 65 minutes once in every three hours is averaged for 5 minutes and the averaged data is used in the computations.

Results and discussion

Surface fluxes, stability parameter and surface drag over Bay of Bengal during northeast monsoon

The surface momentum, sensible heat and latent heat fluxes and stability parameter during the Pilot experiment of BOBMEX are given in fig. (4.2). The variations of all the surface fluxes are in phase. The fluxes are found to be high most of the days during the period. Increase in fluxes is noticed in association with the depression during October 26-31 and the low-pressure area during November 2-7. The stability parameter z/L shows high instability most of the days. The instabilty is due to the depression and the low-pressure area and the convective activity during the period. The surface fluxes increases as soon as the atmosphere is unstable and the atmosphere attains less unstable or near neutral condition. The sensible heat flux is directed downwards when the atmospheric surface layer is highly stable. The latent heat flux is

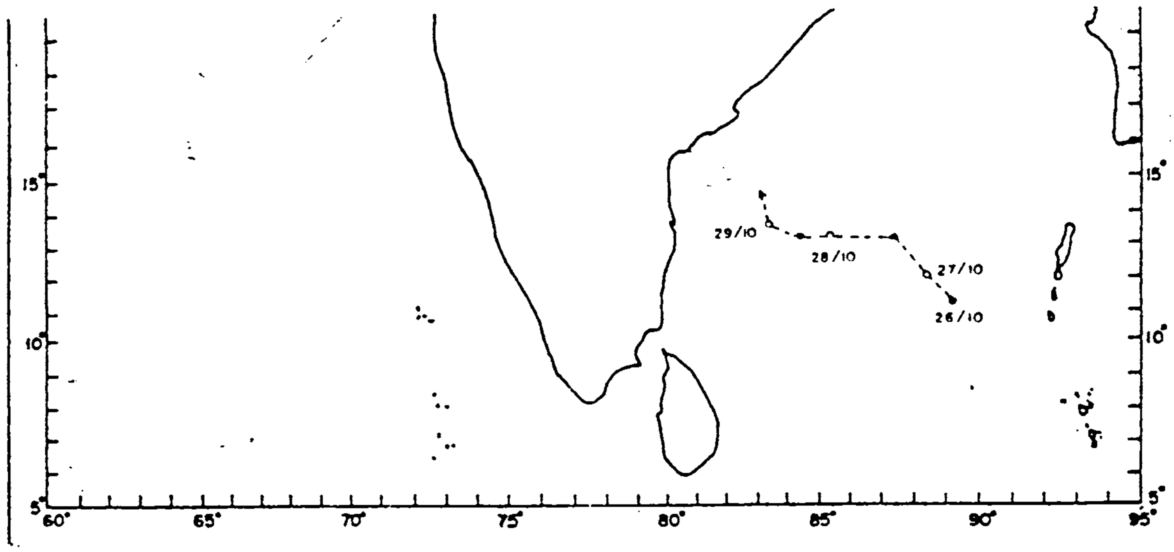


Fig. 4.1 Track of Depression

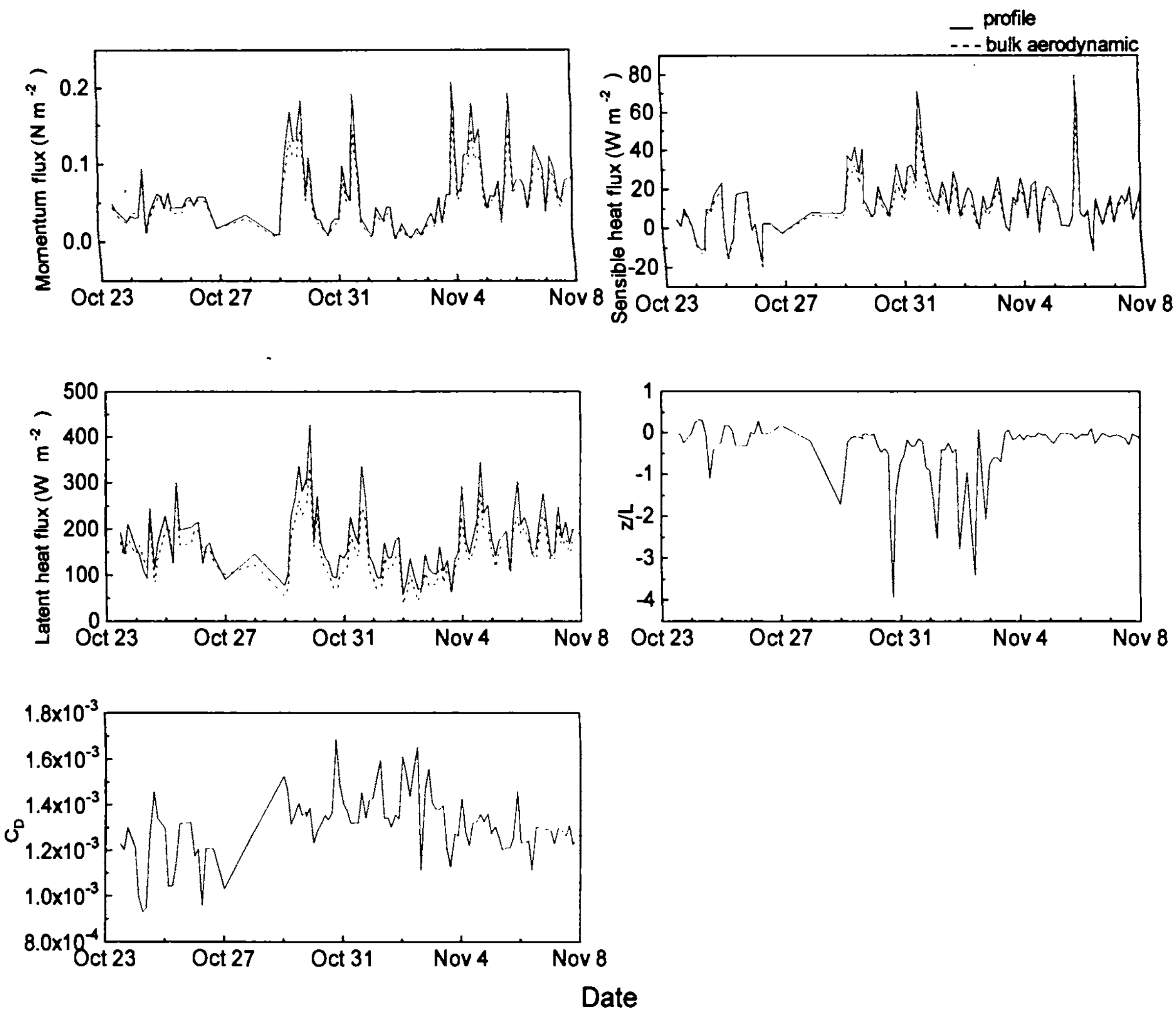


Fig. 4.2 Surface fluxes, stability parameter and drag coefficient during BOBMEX Pilot experiment

found to be one order higher than the sensible heat flux. The surface flux computation using bulk aerodynamic method gives slightly lesser values than that by profile method. The variation is slightly more in the case of latent heat flux. The fluxes obtained by the bulk aerodynamic method are in phase with that computed using the profile method.

The drag coefficient computed during the Pilot experiment period of BOBMEX is shown in fig. (4.2). Higher surface drag occurs when the atmosphere is highly unstable. Higher drag values are noticed during Nov. 1 – 3 of the Pilot experiment. During these days there was immense deep convection over the region. In association with the instability due to the depression during Oct. 26 – 31 the surface drag is found to be about 1.5×10^{-3}. The drag coefficient ranges between 1×10^{-3} and 1.6×10^{-3} during the Pilot experiment. Higher drag values occur during highly unstable situation and it decreases when the atmosphere is less unstable or near neutral. The computed drag coefficients are higher than that considered for the computation using bulk aerodynamic method, which is 1.1×10^{-3}. This may be the reason for the slightly lesser values of surface fluxes than that by profile method.

Comparison of the surface fluxes obtained using different methods

Surface fluxes computed using profile, bulk aerodynamic and eddy correlation methods for November 1 and 3 are shown in fig. (4.3) and fig. (4.4) respectively. The surface fluxes using profile and bulk aerodynamic shows an increase from 03 hours and a maximum is reached during 06 hours and then it decreases on November1. The eddy correlation method also shows an increase from 06 IST and then decreases and increases slightly after 12 IST. The increase in flux is as a result of the increase in turbulence due to deep convection during the period. The atmosphere will be highly unstable during the convection and the turbulence is increased during the period, which resulted in the increase in surface fluxes. After the rainfall in the noon the atmosphere becomes stable and hence the surface fluxes are less. The variations in momentum flux at levels 2 and 3 are in phase and higher flux values occur at level 2. The surface fluxes computed using profile and bulk aerodynamic methods are much higher than that by eddy correlation method and the profile method gives higher values than bulk aerodynamic method.

On November 3 which has an early morning shower and cloud free morning and clouds late in the afternoon, the surface fluxes by profile and bulk aerodynamic method

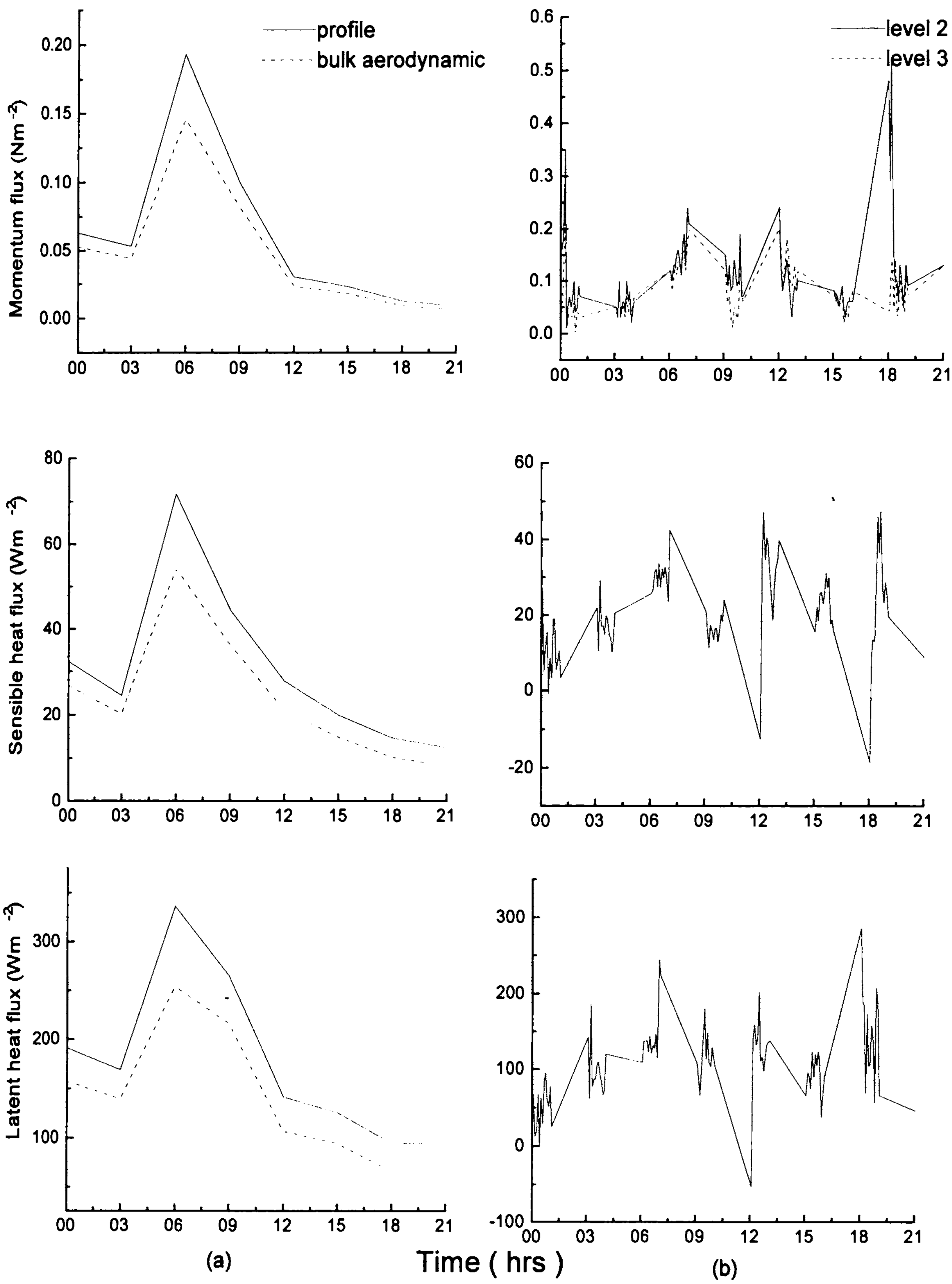


Fig. 4.3 Momentum, Sensible and Latent heat fluxes by (a) profile and bulk aerodynamic and (b) eddy correlation methods for Nov.1 1998.

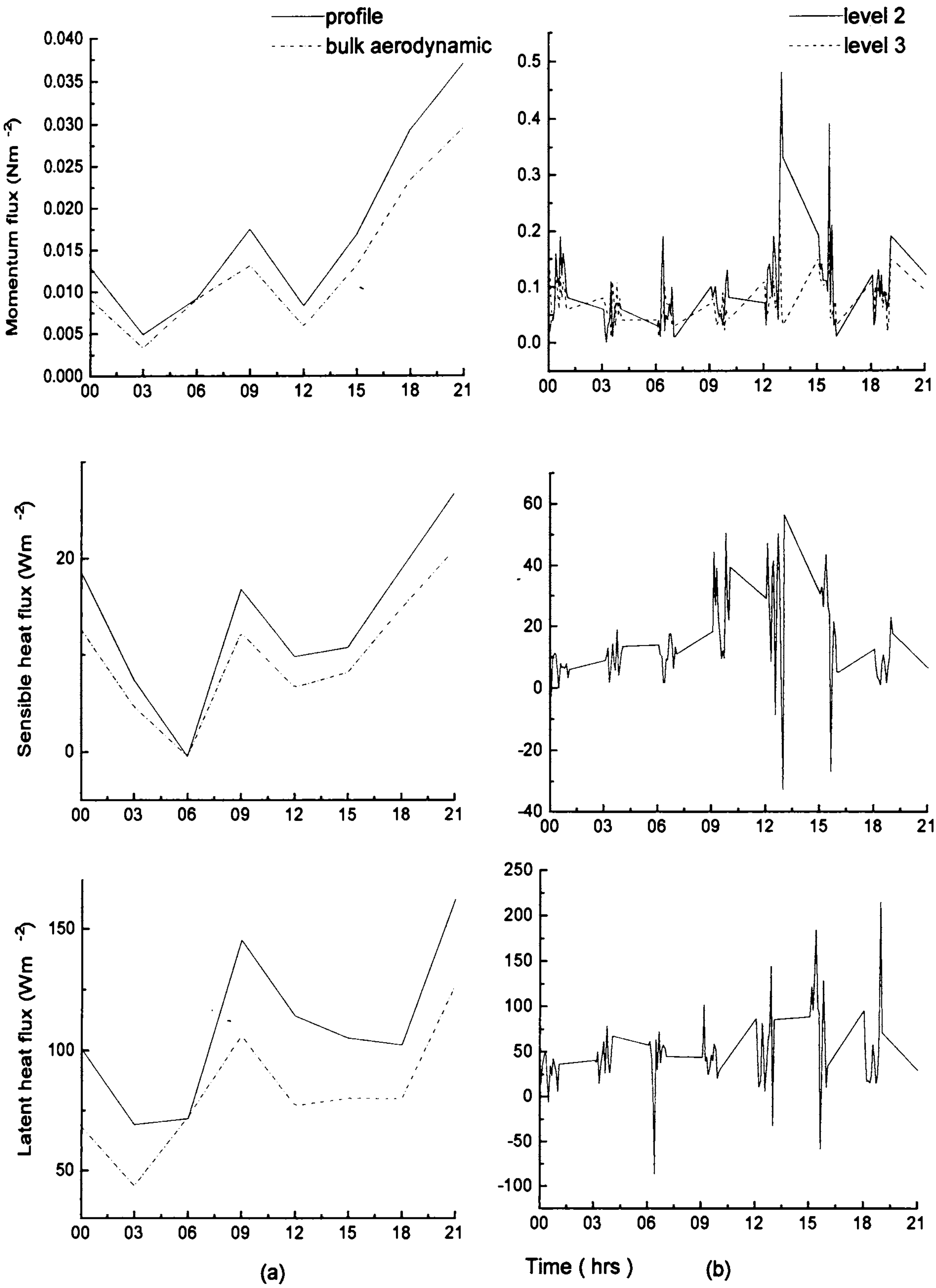


Fig. 4.4 Momentum, Sensible and Latent heat fluxes by (a) profile and bulk aerodynamic and (b) eddy correlation methods for Nov. 3 1998.

shows a decrease in the early morning hours and then gradually increase by 09IST. The fluxes then decrease at 12 IST and then increase in the afternoon. The eddy correlation method shows a gradual increase in fluxes from 06 to 12 IST and gradually decrease after that. Higher fluxes are noticed between 09-15 hours IST. After the early morning shower the atmosphere will be less unstable hence the surface fluxes are also less. In the cloud free morning hours the instability may increase which resulted in the gradual increase in fluxes. The wind speed increased by late afternoon after the rainfall. This increase in wind speed increased the surface fluxes in the afternoon hours. This increase is not much significant in the surface fluxes by eddy correlation method. Momentum and sensible heat fluxes computed using eddy correlation method are found to be higher than that by the other methods whereas profile and bulk aerodynamic methods give higher latent heat fluxes. Also higher difference is noticed between the latent heat flux obtained using profile and bulk aerodynamic methods. Latent heat flux is one order higher than the sensible heat flux.

The surface fluxes are therefore higher during most of the days during the experiment period. An increase in fluxes occurs in association with the formation of depression and deep convective activity. Though the fluxes computed by various methods shows more or less similar trend the magnitudes are not same. The eddy correlation method using fast response data from sensors which gives data at 0.1s interval will be more accurate than the other two methods. The latent heat flux is one order higher than the sensible heat flux. This is because of the high evaporation over the ocean surface. The computed drag coefficient gives a higher value than that used in the bulk aerodynamic method. The drag is higher during the occurrence of disturbances and deep convection.

CHAPTER 5

Thermodynamic Structure during various seasons

The cloud and rain features over the Indian region vary during the pre-monsoon, onset and post-monsoon periods. During the pre-monsoon and post-monsoon seasons the rainfall is mainly from thunderstorms or micro scale systems. Sometimes rainfall may result from mesoscale scale systems also. The presence of clouds can alter the amount and distribution of short and long wave radiative flux in the atmosphere boundary layer. This radiative flux along with the latent heating can change the boundary layer dynamics, turbulence generation and evolution. Therefore it is essential to study the thermodynamic structure of the atmosphere to have a better understanding of the variations in the cloud patterns and associated rainfall during the various periods of monsoon. There are several studies done on the thermodynamic structure and associated rainfall over the Indian region during the pre-monsoon and southwest monsoon seasons. Srinivasan and Sadasivan (1975) studied the difference in thermodynamic structure of the atmosphere between active and weak monsoon. No significant change in the dry bulb temperature was noticed. Desai (1986) noticed significant variation in the dew point temperature and moist static energy between the active and weak monsoon period and no difference in dry bulb temperature was obtained. Ananthakrishnan et al (1965) investigated the seasonal variation of precipitable water vapour in the atmosphere over India and noticed that maximum precipitable water vapour is in the monsoon months and minimum in the winter months. Mukerji (1962) found that the maximum moisture is not reached on the date of monsoon onset and the effect of monsoon is noticed in all layers up to 600hPa. No significant change on moisture content is noticed in the southern peninsular stations at the time of onset over Kerala.

Several studies have been done to study the convective activity and associated rainfall. The initiation of convection requires not only moist instability and a supply of energy from the large-scale environment as indicated by the Convective Available Potential Energy (CAPE) but also dynamic conditions such as rising motion and suitable wind shear, to efficiently release their environmental energy (Garstang et al 1994 and Cohen et al 1995). Large-scale thermodynamic conditions determine the organised convection, which depends on the energy from the environment. Hence the changes in thermodynamic structure can modulate the frequency and the strength of convection. Eltahir and Pal (1996) noticed a positive correlation between the occurrence of convection and CAPE in the Amazon Basin. Williams and Renno (1996) and Fu et al (1994) suggested that convection does not necessarily occur when CAPE exists but other factors such as the negative buoyancy of the atmosphere below the Level of Free Convection (LFC) referred to as Convective INhibition Energy (CINE) and proper dynamic conditions are also likely to control the occurrence of convection. Fu et al (1999) found that the convection responds to the changes in thermodynamic structure and CINE. Weakening of CINE was noticed during convection peaks. The study suggests that changes in large-scale circulation are needed to establish suitable thermodynamic conditions for convective activity and the seasonal changes of convection are more controlled by CINE. Wilde et al (1985) studied the relation of Lifting Condensation Level (LCL) to cumulus onset and found the effect of horizontal non-uniformities in LCL on the time of cumulus onset and amount of cloud cover. Zawadzki et al (1981) reported that the structure of convection and rainfall rate is determined by the thermodynamic variables. Ackerman (1982) and Watson and Blanchard (1984) investigated the influence of low level convergence on convective precipitation and found that the convective development depends mainly on the boundary layer wind and thermodynamic properties of the atmosphere. Lipps and Hemler (1986) did a numerical simulation of deep tropical convection associated with large-scale convergence. The effect of large-scale convergence on the generation and maintenance of deep moist convection was examined by Crook and Moncrieff (1988). Smith and Noonan (1998) found that the low-level convergence lines over northeastern Australia are responsible for the initiation and maintenance of cloud lines observed over the region. These studies indicates that the convective activity is very much dependent on the thermodynamic structure of the atmosphere and low level large–scale convergence. Hence the varying thermodynamic structure and large scale circulations during the various periods such as pre-monsoon, onset and post-monsoon has got a strong influence on the rainfall activity during these periods. Also the thermodynamic structure varies during the occurrence of mesoscale systems. In this work CAPE and CINE over some selected stations are obtained during the pre-monsoon, onset and post-monsoon periods to have a better understanding of the varying thermodynamic conditions during various periods. Also the variations in LCL during the various periods are obtained for the stations. It was reported by Rao (1976) and Mukerji (1962) that the moisture content does not show any increase by the advance of monsoon. This feature was reexamined for the selected stations. According to Ananthakrishnan et al (1968) and Rao (1976) the lower tropospheric westerly speed and depth increased over the south peninsula at the time of onset of monsoon over the subcontinent. This feature was investigated for various stations over the Indian region during the onset of monsoon over each station.

Materials and Methods

The CAPE, CINE and LCL are computed using the 00UTC upper air radiosonde data at Bombay, Mangalore and Trivandrum from India Meterological Department for the months April, May, June, October, November and December. The period of study is for five years from 1984 to 1988. The data is for every 50 hPa difference from the surface up to 250hPa. The CAPE, CINE and LCL computations are also carried out using upper air data at 00 and 12UTC during May, June 1988 for Mangalore, Bangalore, Calcutta, Guwahti and Jodhpur to have a better understanding of the spatial variations in the thermodynamic structure.

For an unsaturated air parcel at pressure p the saturation level is reached by dry adiabatic ascent to the pressure level where the parcel is just saturated with no cloud liquid water. This level is known as the Lifting Condensation Level (LCL) and is often taken as cloud base. All the clouds are not positively buoyant. Although condensation starts at LCL, the air parcel will be often negatively buoyant. Latent heat is released as condensation occurs and if the parcel has sufficient inertia to overshoot high enough then its potential temperature may rise and at a point it becomes warmer than the environment. This point where the air parcel becomes buoyant first is called the Level of Free Convection (LFC). Then the air parcel becomes active and continues to rise due to its own buoyancy. Eventually while rising it reaches a point where it becomes cooler than the environment. This point is called the Limit Of Convection (LOC). The cloud parcel may overshoot beyond LOC because if its inertia but stops rising at the cloud top (Stull 1988).

Convective Available Potential Energy (CAPE) is a measure of maximum possible kinetic energy that a statically unstable parcel can acquire (neglecting effects of water vapour and condensed water on the buoyancy) assuming that the parcel ascends without mixing with environment and instantaneously adjusts to the local environmental pressure. In other words, CAPE is the total energy that can be utilised by the air parcels for rising from LFC to LOC. Accordingly, the average vertical velocity of rising parcels in clouds can be derived from CAPE. In fact, the typical upward vertical velocity of a parcel in a thunderstorm is usually between 10 to 20ms^{-1} though the vertical velocity derived from CAPE is one order higher. This is due to the entertainment or detrainment processes, negative buoyancy of liquid water in the cloud, heat loss from the systems associated with dropping out of liquid water and frictional loss. The role of CAPE arises only when the surface air parcel rises to LFC by external forcing. Orographic lifting, updraft by system, frictional convergence due to strong winds can act as the mechanism for lifting the air parcels from surface to LFC. When the lowest layer of the atmosphere is superadiabatic, convection can set in at the surface. Usually the thickness of this unstable layer is small and may not be sufficient to raise the air parcels to LFC. Once the surface air parcels reaches LFC it can rise by its own buoyancy until it becomes cooler than the environment at LOC. The strength of the upward buoyancy is proportional to the magnitude of the difference in temperatures of the parcel and environment. CAPE is the total energy per unit mass supplied by the buoyancy force throughout the vertical layer in which the parcel is warmer than the environment. The negative buoyancy of the atmosphere which is referred as the Convective INhibition Energy (CINE) is the energy to be supplied to the surface air parcel of unit mass to lift it to cross the stable layer, that is from surface to LFC (Holton 1992). Usually the lowest layer of the atmosphere is stable.

Daily values of CAPE, CINE and LCL are computed to study their variations during pre-monsoon, onset and post-monsoon periods. CAPE and CINE are computed as follows. The upward buoyancy acceleration of saturated air parcel where it is warmer than the environment is given by parcel method (Hess 1959) as

$$\frac{dw}{dt} = \frac{g(T_p - T_e)}{T_e} \qquad (5.1)$$

where g is the acceleration due to gravity, T_p and T_e is parcel temperature and environment temperature respectively.

CAPE is the total energy used up by the surface air parcel when it rises from LFC to LOC.

Therefore CAPE = (dw/dt) dz (5.2)

But from hypsometric equation we can write

$$dz = RT \ln(P_1/P_2)/g \qquad (5.3)$$

where T is the temperature of the environment i.e. T_e

Therefore CAPE = R $(T_p - T_e) \ln (P_1/P_2)$ (5.4)

where P_1 is the pressure at LFC and P_2 is that at LOC and R is the specific gas constant for air.

Since $(T_p\text{-}T_e)$ is not the same throughout the layer from LFC to LOC, the CAPE values are obtained by integrating the above equation, So the equation becomes

$$CAPE = {}_{LFC}\int^{LOC} R(T_p - T_e)\, dp/P \qquad (5.5)$$

CAPE value is computed by considering thin layers of the atmosphere of 1hPa thickness from LFC to LOC. T_p values are obtained from saturated adiabat profile and T_e from interpolating the environmental profile at 1hPa interval.

CINE is computed using the following equation

$$CINE = {}_{surface}\int^{LFC} R(T_p - T_e)\, dp/P \qquad (5.6)$$

Computational procedure is same as that as CAPE by taking dp = 1hPa from surface to LFC.

LFC is computed on the principle that at that level the equivalent potential temperature of the surface air and environmental air is the same. The level at which the equivalent potential temperature of the surface air and environmental air are again same above the LFC is taken as LOC. The equivalent potential temperatureθ_e for θ is obtained as (Holton 1992)

$$\theta_e = \theta \exp((L\, q/(Cp\, T)) \qquad (5.7)$$

However, Bolton (1980) found an error in the above approximation, which causes an error of more than 3°C and suggested a formula for evaluating θ_e to an accuracy of 0.0018°C.

$$\theta_e = \theta_{LCL} \exp((3.036/T_{LCL} - 0.00178)(1 + 0.448 \times 10^{-3} w)w) \quad (5.8)$$

where T_{LCL} is the dry bulb temperature at LCL, θ_{LCL} is the potential temperature at LCL.

$$\theta_{LCL} = T\,(1000/(p-e))^{0.2854}\,(T/T_{LCL})^{0.00028w} \qquad (5.9)$$

T, P, e, and w are dry bulb temperature, pressure, vapour pressure and mixing ratio in g/kg of the surface air respectively. The equation (5.8) is solved iteratively using the software by Babu (1996).

The specific humidity of the surface air parcel gradually decreases beyond LCL due to condensation. The level at which the actual specific humidity of the parcel begins to decrease is the LCL. LCL is determined iteratively solving the specific humidity equation for a specific humidity of the atmosphere same as the surface specific humidity.

The specific humidity of the surface air is

$$q = 0.622e/(P - 0.378e) \qquad (5.10)$$

where e is the actual vapour pressure for water in hPa and P the surface pressure. The vapour pressure e in the program corresponding to the temperature, T is evaluated using the Teten's formula (1930) as

$$e = 6.11 \exp(A(T - 273.16)/(T-B)) \qquad (5.11)$$

where T is the dew point temperature of the parcel at the surface in Kelvin and the constants

$$A = 21.87\,;\ B = 7.66 \text{ when } T < 263$$

$$A = 17.26\,;\ B = 35.86 \text{ when } T \geq 263$$

LCL is computed as per the detailed method available by Babu (1996).

The profile of zonal wind was obtained from the upper air wind data at Bombay, Mangalore and Trivandrum during May and June for a five year period from 1984 to 1988 to study the variations in wind structure during the onset phase of monsoon. The zonal wind structure at Trivandrum, Minicoy, Amini, Mangalore, Bangalore, Goa, Bombay, Jodhpur, Lucknow, Calcutta and Guwahati during three days before the onset at each station, onset day and three days after the onset during 1988 was also investigated. The data is from the radiosonde observations from the India Meteorological Department.

Zonal component of wind u is computed as

$$u = - ff \, \mathrm{Sin}\,(dd) \qquad (5.12)$$

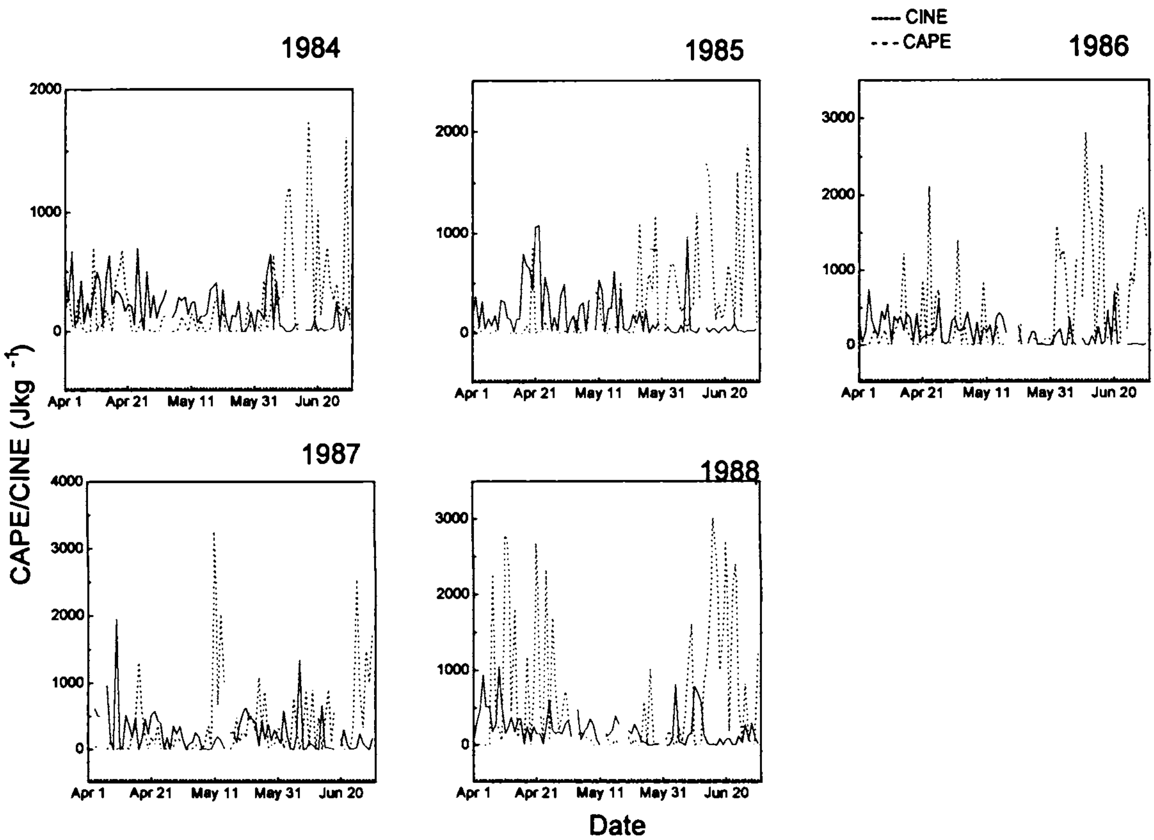
where ff is the wind speed and dd is the wind direction .

The variation in precipitable water content in the atmosphere during the monsoon onset over the subcontinent was investigated using NCEP reanalysis data during 1987 to 1991. The details of the data are given by Kalnay et al (1996). The total preciptable water content from surface to the top of the atmosphere over the region 40°E to 110°E and 50°S to 40°N during three days before onset, onset day and two days after the onset are taken for the study.

Results and discussion

Variation of CAPE and CINE during various periods

CAPE, CINE and rainfall associated with thunderstorms over Bombay, Mangalore and Trivandrum during April, May and June are given in figures (5.1a to 5.1c) respectively. CAPE values are higher than CINE at Bombay by late May and June. Higher CAPE values are noticed on most of the days at Mangalore and Trivandrum during the period. This indicates that during the pre-monsoon and onset period favourable condition exist in the atmosphere for supply of energy from the large-scale environment to the air parcels. This is one of the required conditions for the occurrence of convection. The thunderstorm activity during the period shows that most of the higher CAPE values are associated with thunderstorms. It was also noticed that not all the high CAPE is related to convection. This shows that CAPE is not the only factor that controls



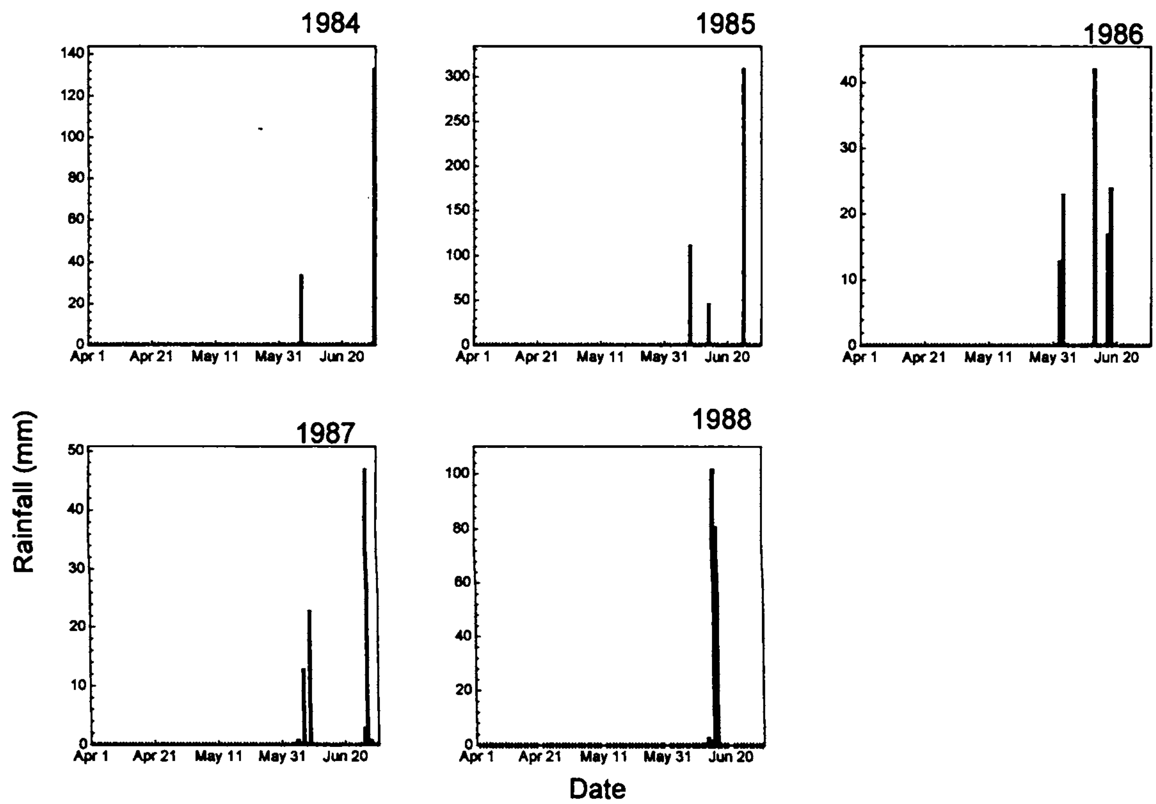


Fig. 5.1a CAPE/CINE and rainfall associated with thunderstorms during April- June 1984 - 1988 at Bombay

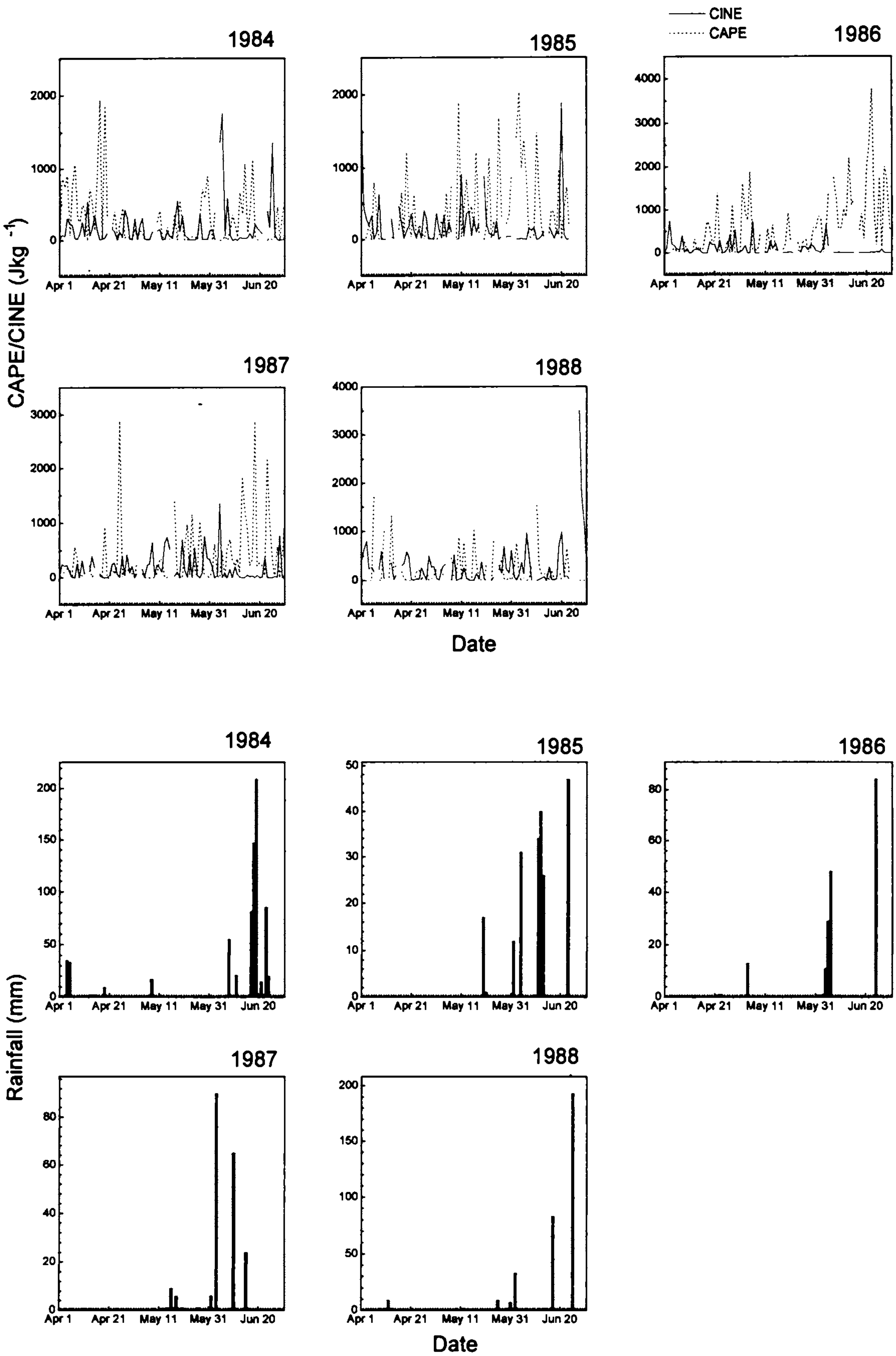


Fig. 5.1b CAPE/CINE and rainfall associated with thunderstorms during April - June, 1984 - 1988 at Mangalore

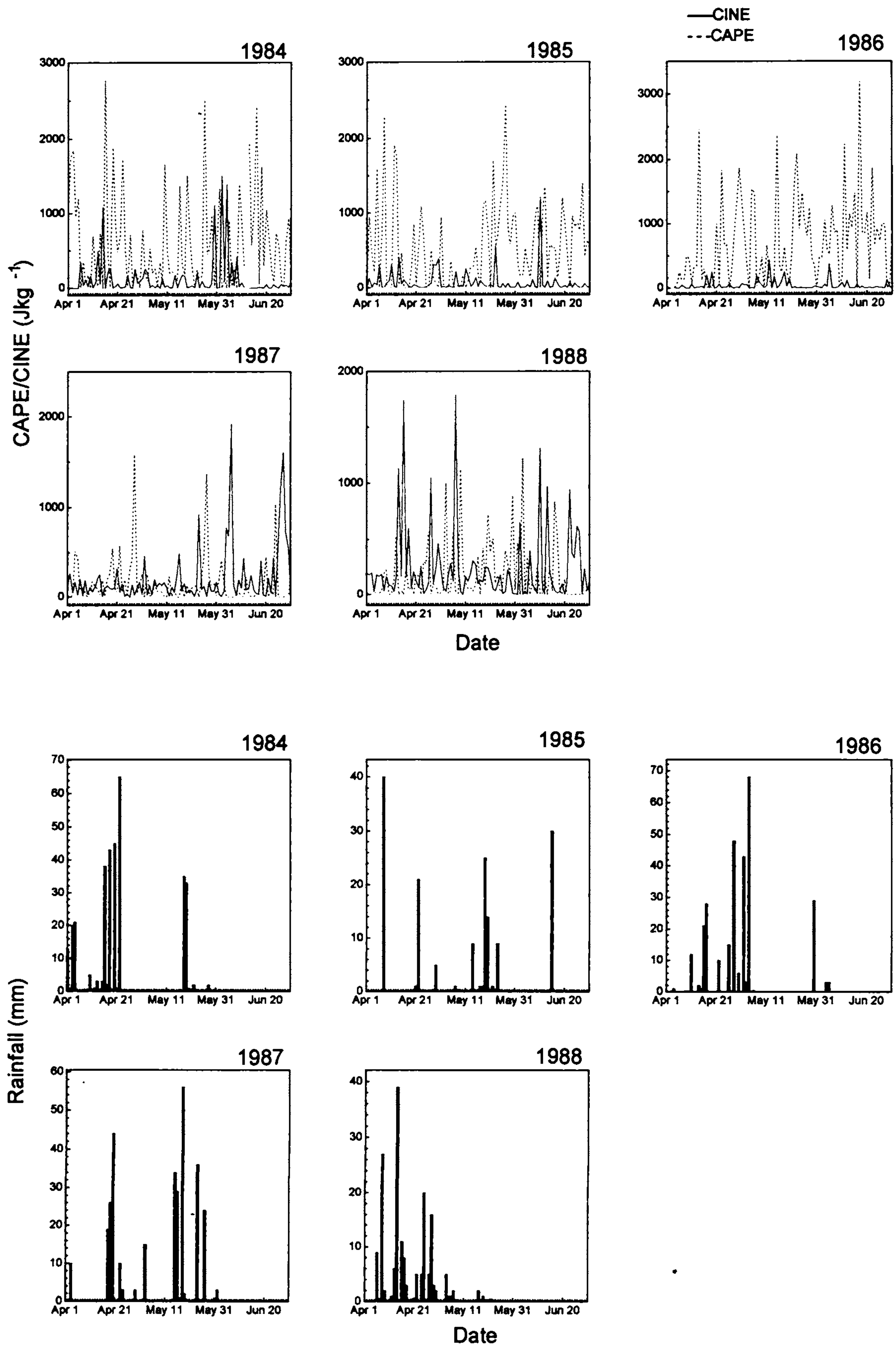


Fig. 5.1c CAPE/CINE and rainfall associated with thunderstorms during April-June, 1984 -1988 at Trivandrum

the convection. Fu et al (1994) and Williams and Renno (1993) also noticed this. CINE values are found to the much lesser during the thunderstorms and it increases soon after the thunderstorm activity. This indicates that the atmosphere becomes favourable for the convection to set in by weakening the CINE and soon after the rainfall activity the lowest layer of the atmosphere becomes highly stable. High CINE values shows that high energy is required to lift an air parcel from the surface to LFC.

CAPE, CINE and rainfall during the post monsoon period over Bombay, Mangalore and Trivandrum are given in figures (5.2a to 5.2c). Higher CINE values exist throughout the period except early October at Bombay. At Mangalore and Trivandrum also CINE values are higher during most of the days throughout the period. Even if the surface air parcel receive enough energy from external agency to reach LFC the CAPE values are much smaller. This indicates that the atmospheric condition is not favourable to supply energy to the air parcel so that it can rise by buoyancy. Most of the thunderstorms during the period are associated with high CAPE and low CINE. The frequency of thunderstorm activity and higher CAPE are more at Mangalore and Trivandrum than Bombay during the post-monsoon period. The thunderstorms are higher during October and November months during the post-monsoon season. The high CINE during the period indicates highly stable lower atmosphere. The thunderstorm activity is more frequent during pre-monsoon period than the post-monsoon season. Therefore we can say that a positive correlation exists between the convective activity and CAPE, and CINE is having a control over the convective activity and their seasonal variations.

CAPE and CINE during May-June 1988 at Mangalore, Bangalore, Calcutta, Guwahati and Jodhpur are given in fig (5.3). CAPE is higher than CINE in most of the days during the period at all the stations except at Jodhpur and Bangalore. At Jodhpur CINE is higher than CAPE in almost all the days during the pre-monsoon and onset period. At Bangalore high CAPE occurs only on certain days and CINE is higher than CAPE in most of the days. This indicates that the lower layer of the atmosphere is highly stable at Jodhpur throughout the period and in most of the days at Bangalore. Even if the air parcels are lifted upto LFC by some other external mechanism the atmospheric condition is not favourable for the parcel to rise by buoyancy force because of lesser CAPE. Therefore at these stations unfavourable atmospheric condition prevails for convective activity during the period. At Mangalore, Calcutta and Guwahati the CINE

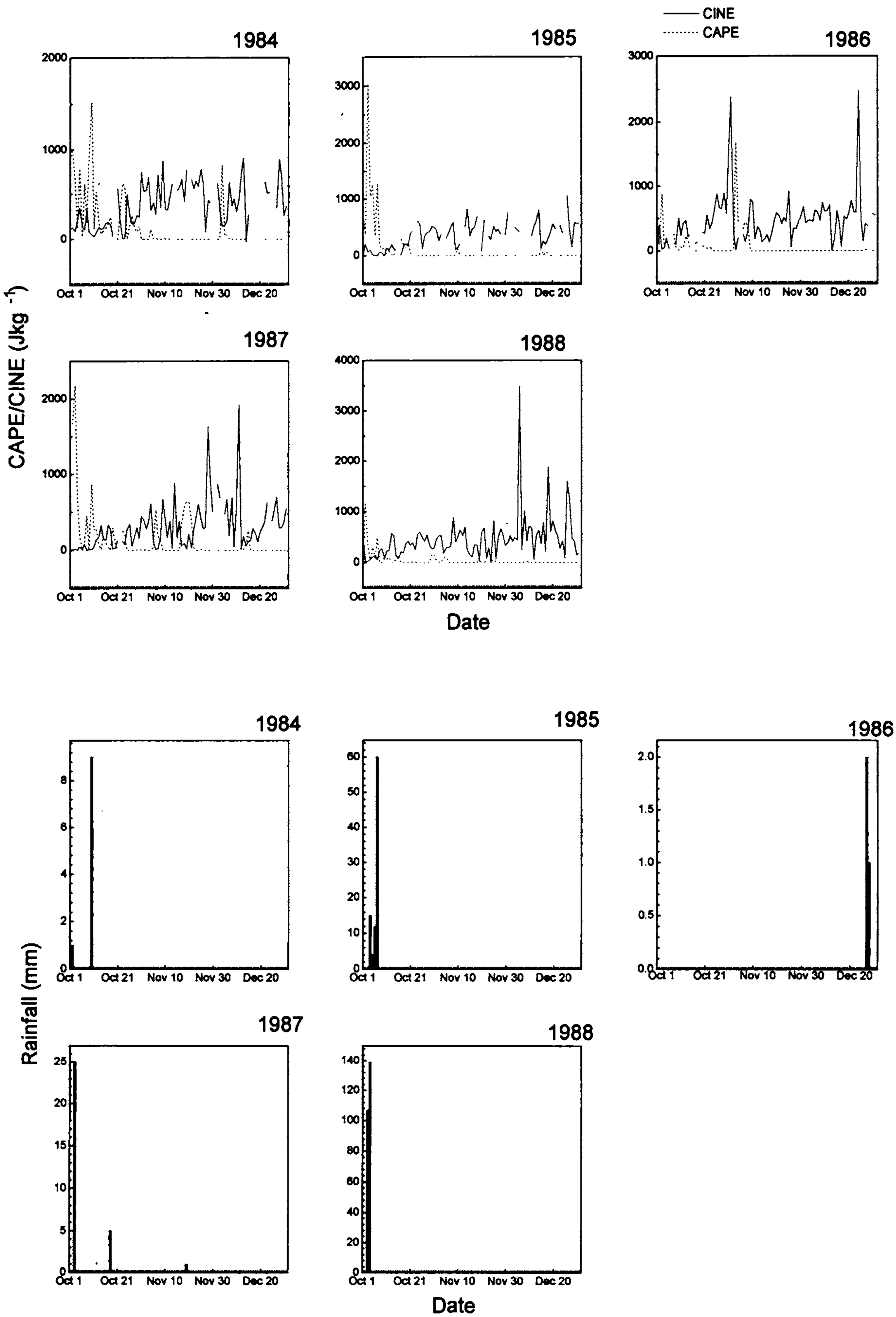


Fig. 5.2a CAPE/CINE and rainfall associated with thunderstorms during October - December, 1984 -1988 at Bombay

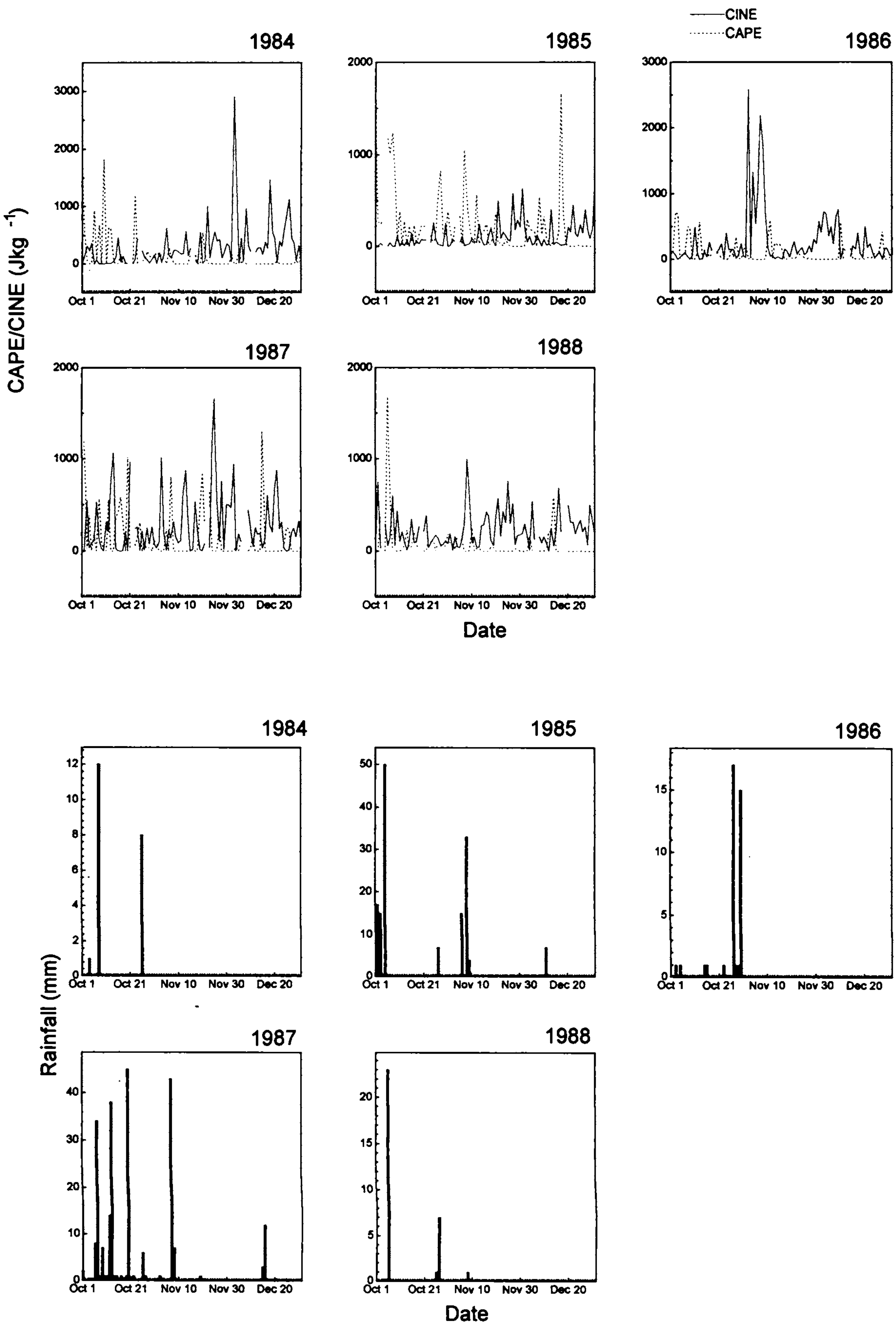
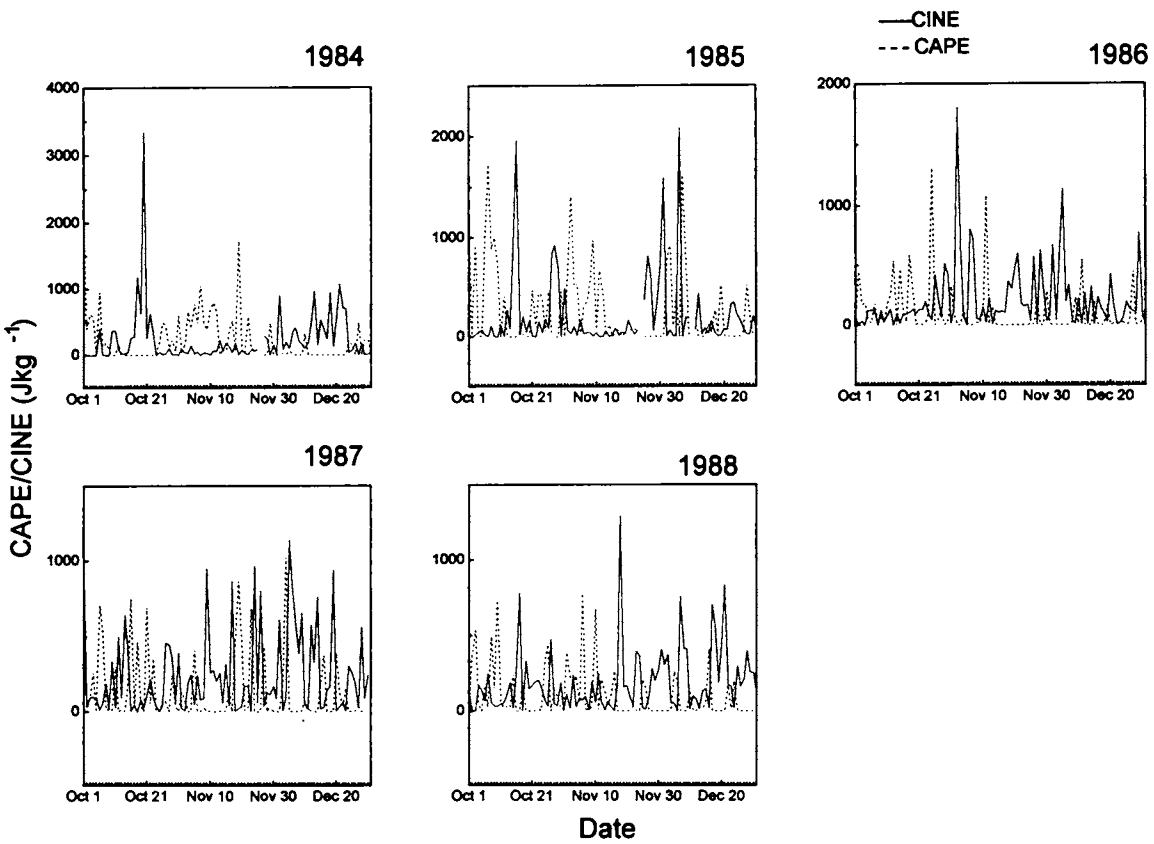


Fig. 5.2b CAPE/CINE and rainfall associated with thunderstorms during October- December, 1984 - 1988 at Mangalore



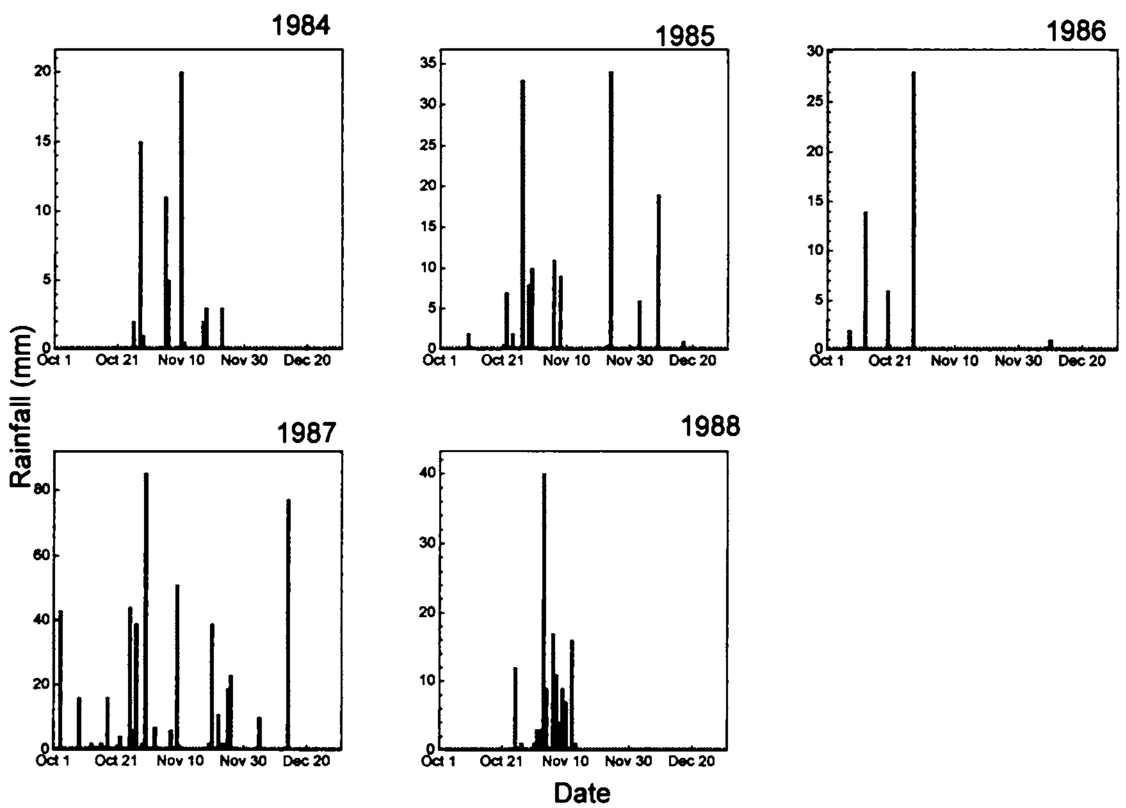


Fig. 5.2c CAPE/CINEand rainfall associated with thunderstorms during October-December, 1984 - 1988 at Trivandrum

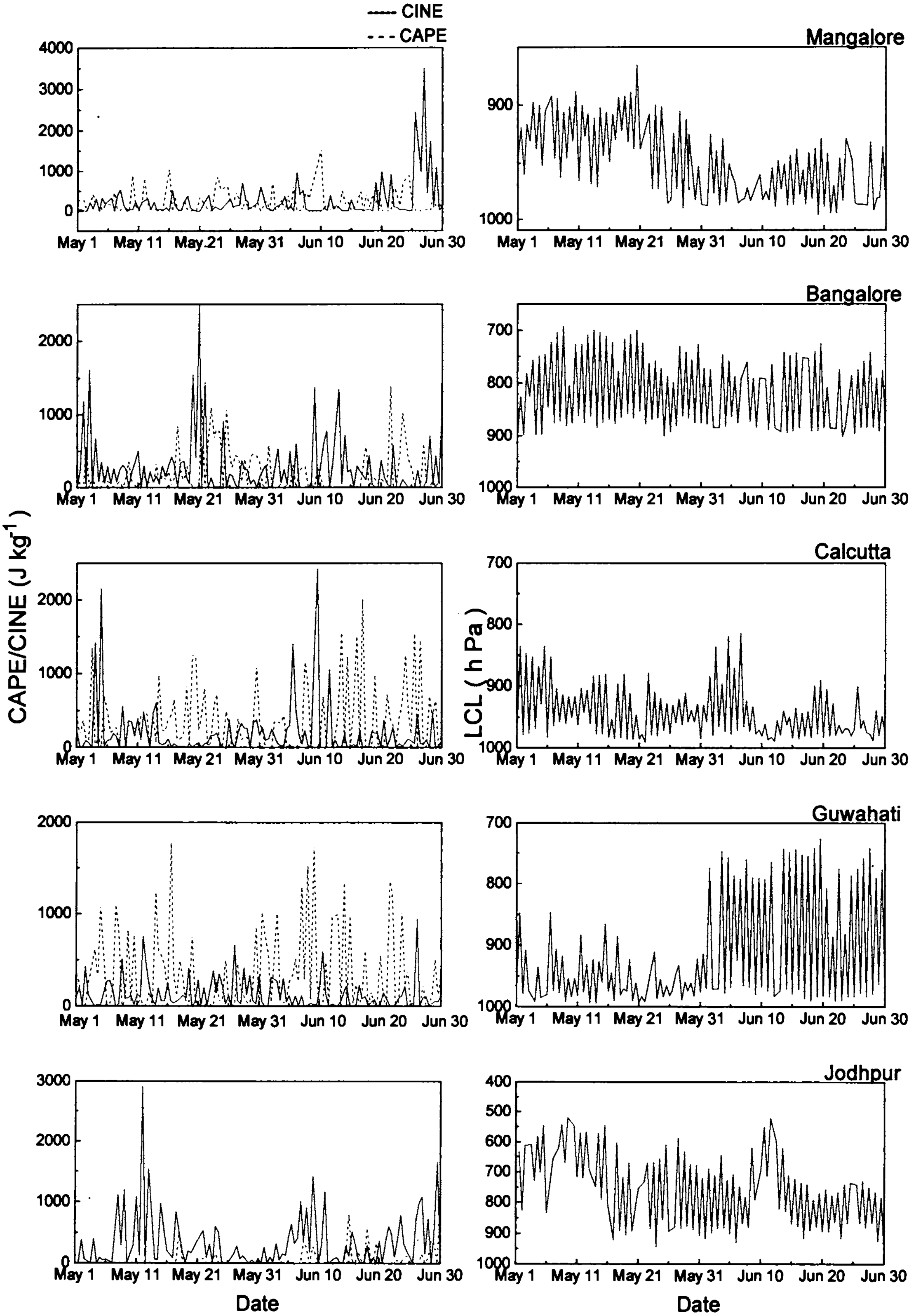


Fig. 5.3 CAPE/CINE and LCL during May-June 1988 at Mangalore, Bangalore, Calcutta, Guwahati and Jodhpur

values are much lesser in most of the days therefore a slight lifting will help the air parcel to reach LFC. As the air parcel reaches LFC it can rise by its own buoyancy force which is supplied to the parcel from the environment. At Mangalore and Guwahati the orography will help the air parcel to reach LFC. Calcutta is near the eastern end of monsoon trough, which is also called as a dynamic trough where the convergence reaches up to mid troposphere. Therefore the air parcels are lifted upwards by the large-scale convergence in the region. Jodhpur is located near the western end of monsoon trough, which is a heat low region with shallow ascent, and stable condition prevails usually. Thus generally dry convection with shallow clouds occurs over Jodhpur. Bangalore is located at the leeward side of Western Ghats. Hence the monsoon flow which cross the Western Ghats will sink and hence stable conditions prevails there and therefore convective activity is very less at the station.

Variation of Lifting Condensation Level

The daily values of Lifting Condensation Level in hPa during April, May and June at Bombay, Mangalore and Trivandrum are given in figure (5.4a) and that during October, November and December are given in figure (5.4b) respectively. Lower LCL values are found during the pre-monsoon period when compared to the onset period at all the three stations. Thus a lowering of LCL occurs by the monsoon onset. As the monsoon is reached over a station the atmosphere becomes highly humid and the air reaches near the saturation point so that a slight dry adiabatic lifting of the air parcel will leads to saturation and then condensation. The LCL is also referred as the cloud base. Hence during the pre-monsoon season high clouds are noticed in the atmosphere and during the monsoon period low and medium clouds are formed. During pre-monsoon at Bombay the LCL reaches 920hPa and occasionally above that where as by the onset the maximum reach of LCL is upto about 940hPa. LCL is at higher levels prior to the onset of monsoon. At Mangalore LCL reaches up to about 920hPa in the pre-monsoon period and it ranges between 980hPa and 920hPa. It lowers by onset and is found to be below 980hPa mostly and it reaches up to 950hPa occasionally. The range of LCL at Trivandrum during the pre-monsoon period is mostly between 990-960hPa. After the onset it is found to be below 980hPa most of the days with occasional rise in level and the maximum height it is found to reach is about 960hPa. In all the three stations the LCL reaches higher levels during May.

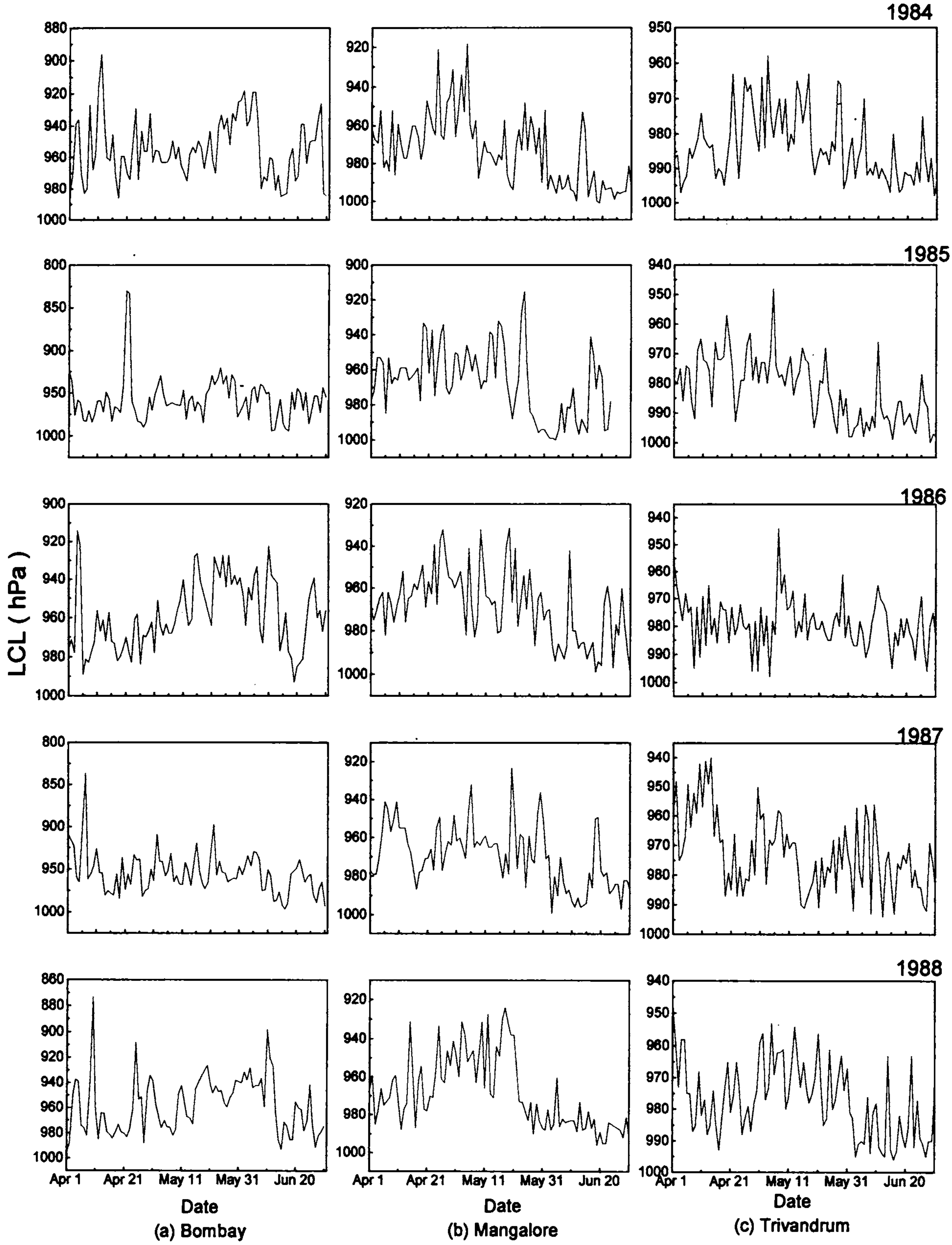


Fig. 5.4a Lifting Condensation Level during April - June,1984 - 1988 at (a) Bombay, (b) Mangalore and (c) Trivandrum.

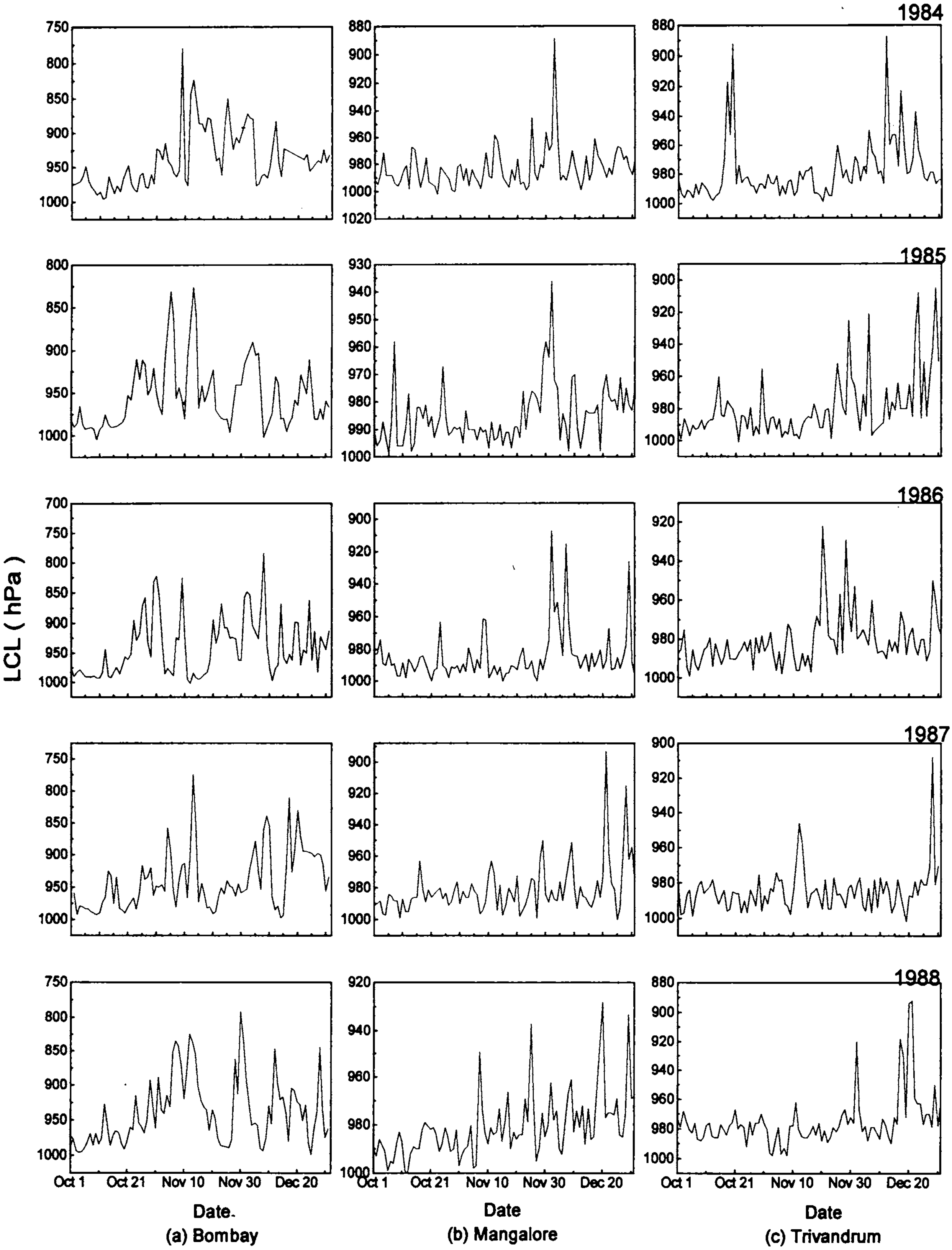


Fig. 5.4b Lifting Condensation Level during Oct.- Dec.,1984 - 1988 at (a) Bombay, (b) Mangalore and (c) Trivandrum.

The LCL, which is at the lower levels in the month of October gradually rises by November and reaches higher levels at Bombay. In early days of October the LCL is noticed below 950hpa which rises later and reaches up to about 850hPa or above during November and December. At Mangalore the LCL which is below 980hPa during most of the days in October and early November gradually rises above that by December. It may reach up to about 960hpa and occasionally above that. At Trivandrum the LCL is below 980hPa during October and early November rises gradually and reaches up to about 960hPa and occasionally to higher level. This shows that after the monsoon the atmosphere becomes less humid so that an air parcel needs to be lifted to higher levels for saturation and condensation. This indicates that the monsoon activity affects the thermodynamic structure of the atmosphere by lowering the lifting condensation level.

LCL during May and June 1988 at Mangalore, Bangalore, Calcutta, Guwahati and Jodhpur are given in fig. (5.3). LCL lowers at all the stations except Guwahati by the onset of monsoon. At Bangalore only a slight lowering of LCL is noticed and also the variation between the values at 00 and 12UTC is large. LCL ranges from 960hPa to 700hPa before onset, which becomes 900hPa to 775hPa after the onset. At Calcutta the LCL which may reach up to 850hPa before onset lowers and its maximum height is up to 900hPa. At Guwahati after the onset large variation is noticed between the 00 and 12UTC LCL values. It reaches even up to 700 hPa at 12UTC after the onset but at 00UTC its height is near 975hPa or at lower levels. Before the onset it reaches only up to 850hPa or below. At Jodhpur the range of LCL is between 900hPa and 550hPa before the onset but it lower and the range becomes between 900hPa and 800hPa after the onset. The LCL is at higher levels during 12UTC when compared to that at 00UTC.

Wind Structure during the onset of monsoon

The vertical structure of zonal wind over Bombay, Mangalore and Trivandrum during May and June for five years from 1984 to 1988 are shown in figures (5.5a to 5.5e). The westerly wind speed and depth increases prior to or at the time of onset of monsoon over each station. The onset day of monsoon over the subcontinent was on May 30th in 1984. At Trivandum the height of westerly wind regime gradually increases and by end may it reaches even up to 250hPa. The westerly depth and speed gradually increases

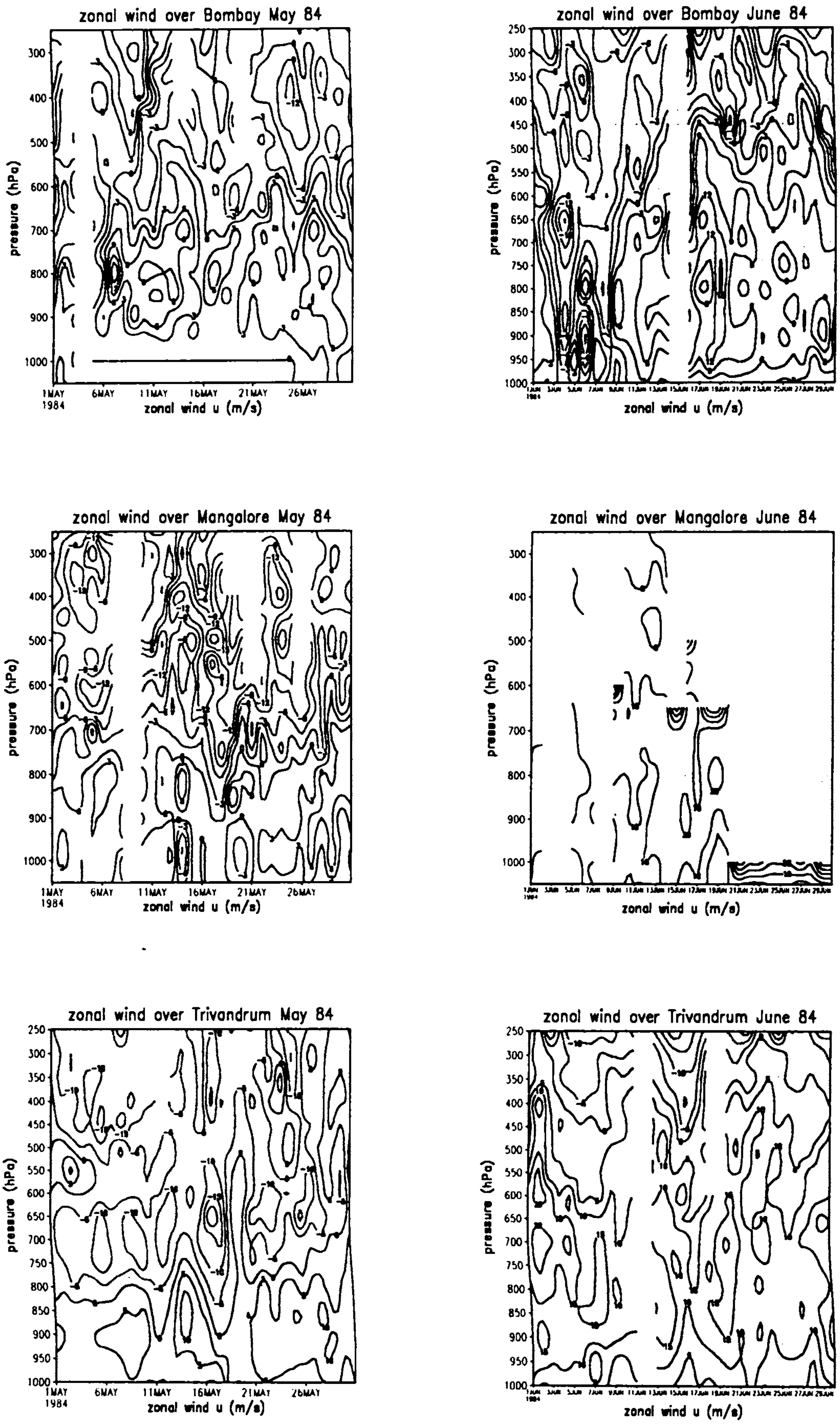


Fig. 5.5a Zonal wind structure over Bombay, Mangalore and Trivandrum during May-June 1984

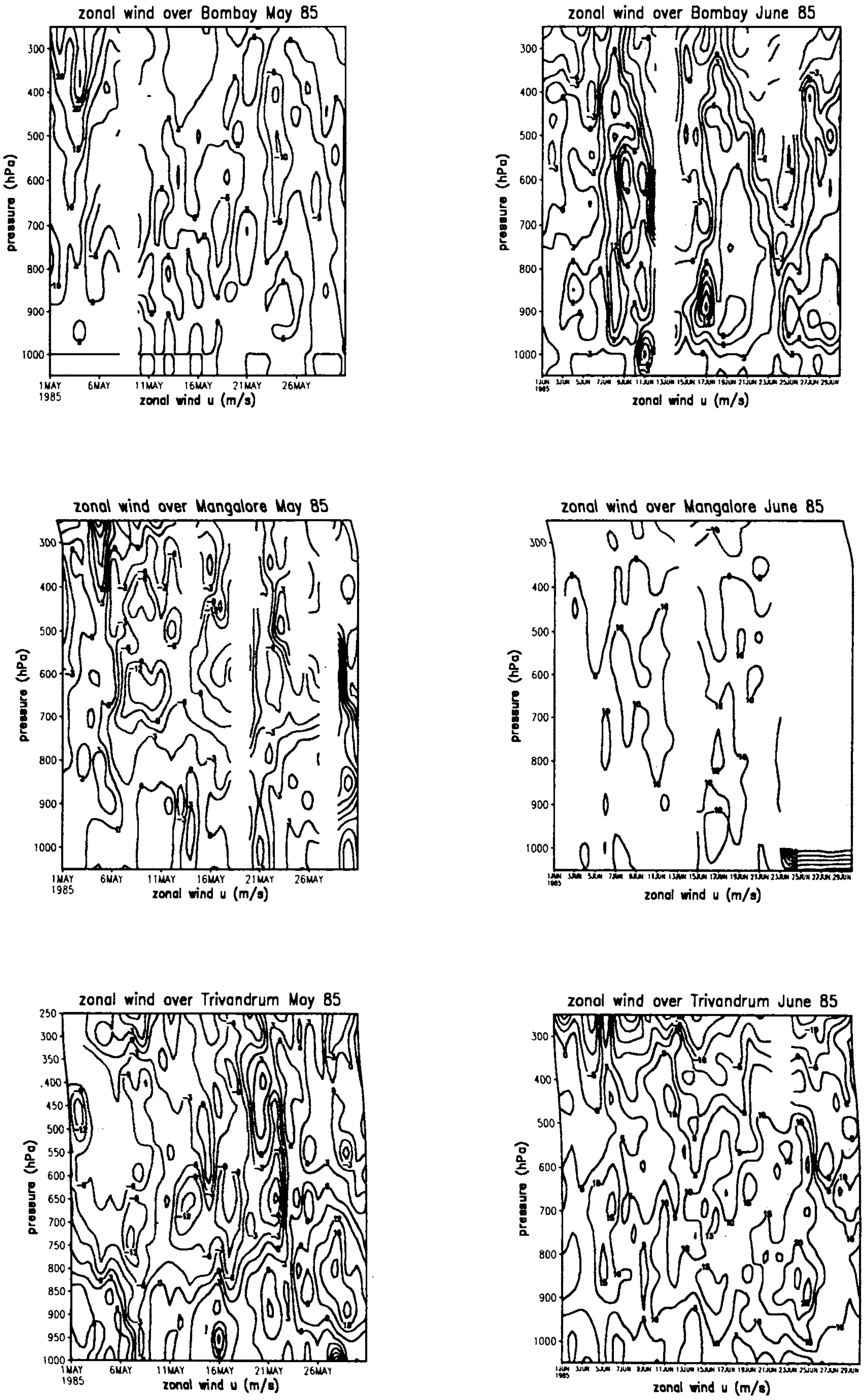


Fig. 5.5b Zonal wind structure over Bombay, Mangalore and Trivandrum during May-June 1985

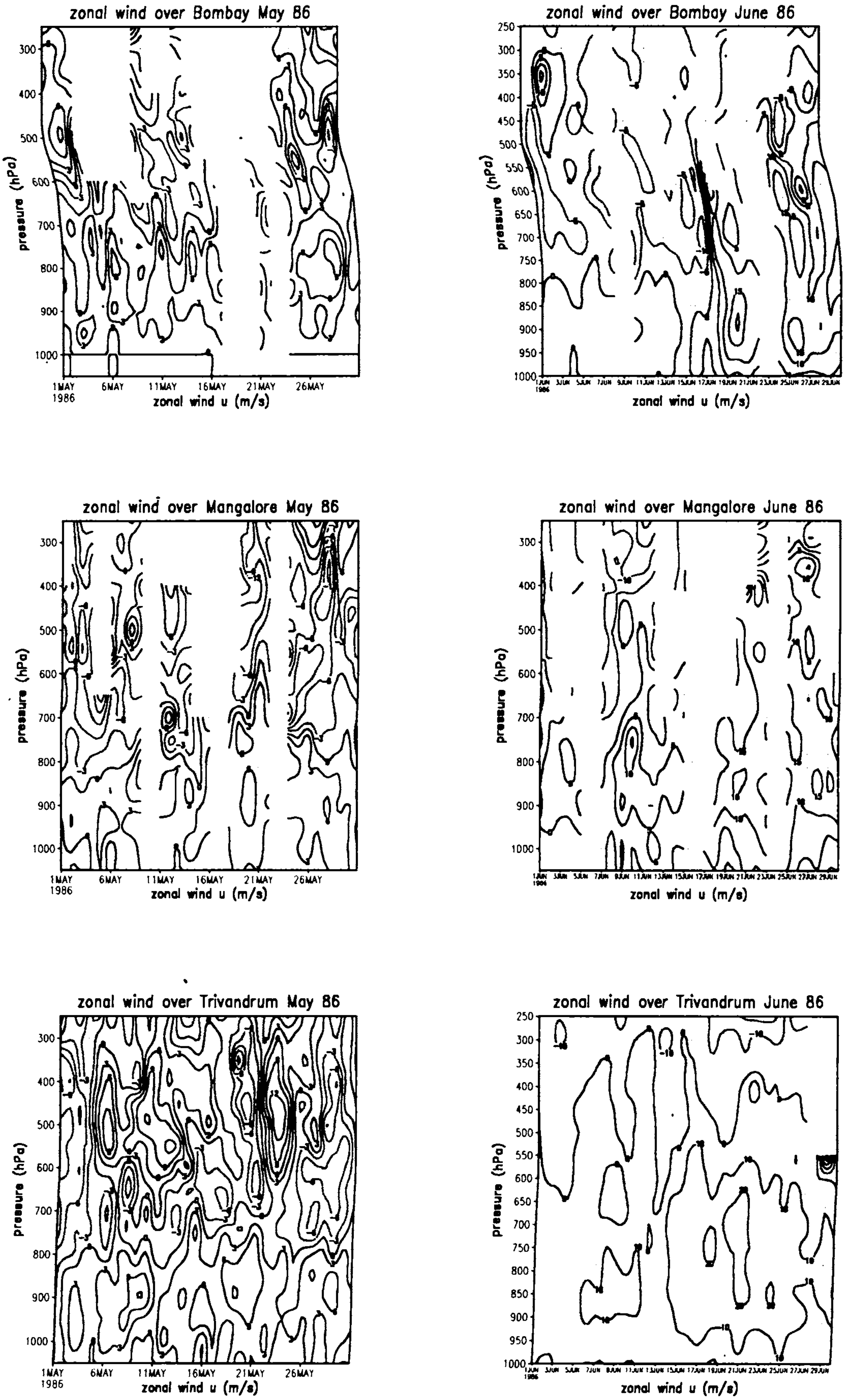


Fig. 5.5c Zonal wind structure over Bombay, Mangalore and Trivandrum during May-June 1986

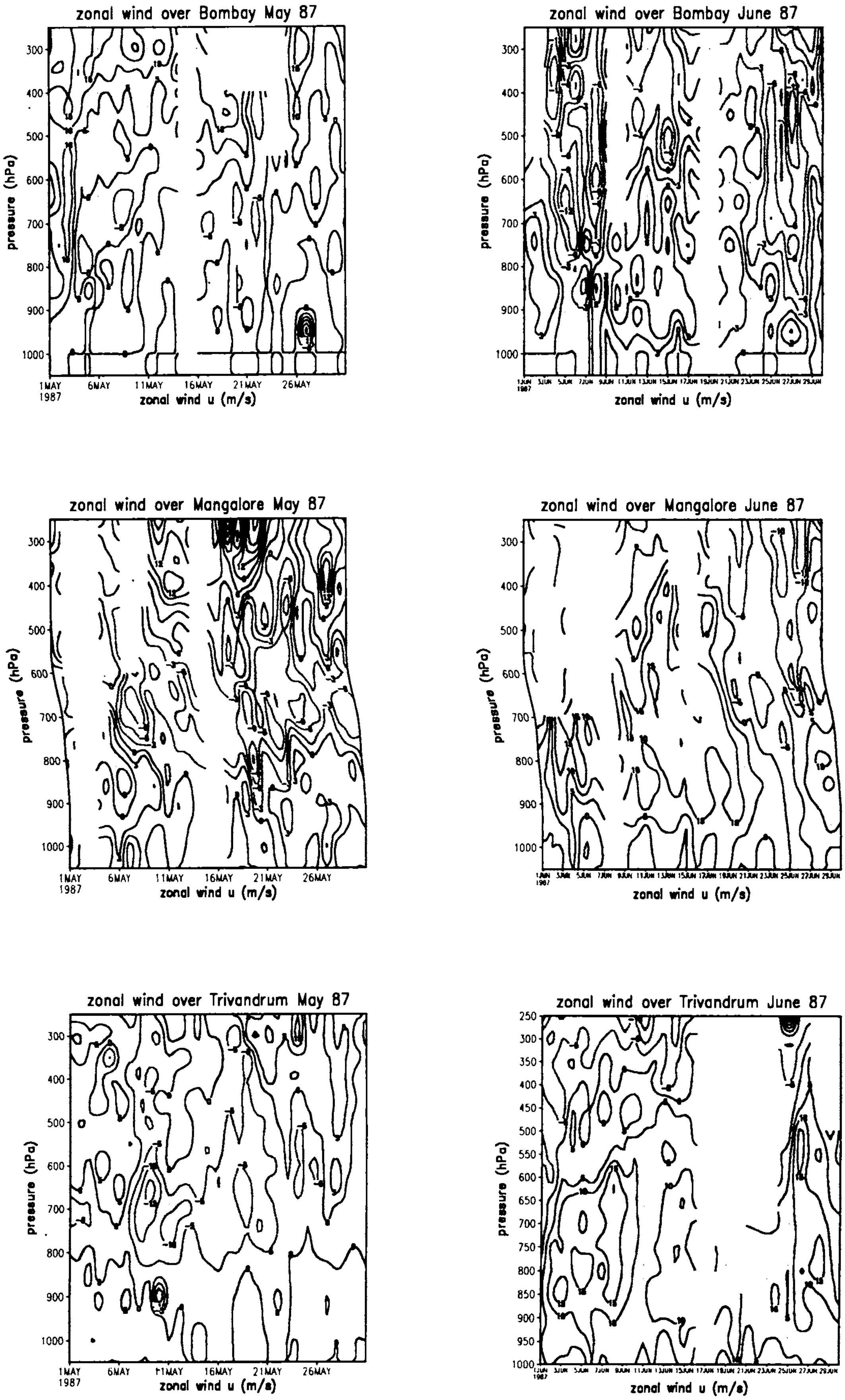


Fig. 5.5d Zonal wind structure over Bombay, Mangalore and Trivandrum during May-June 1987

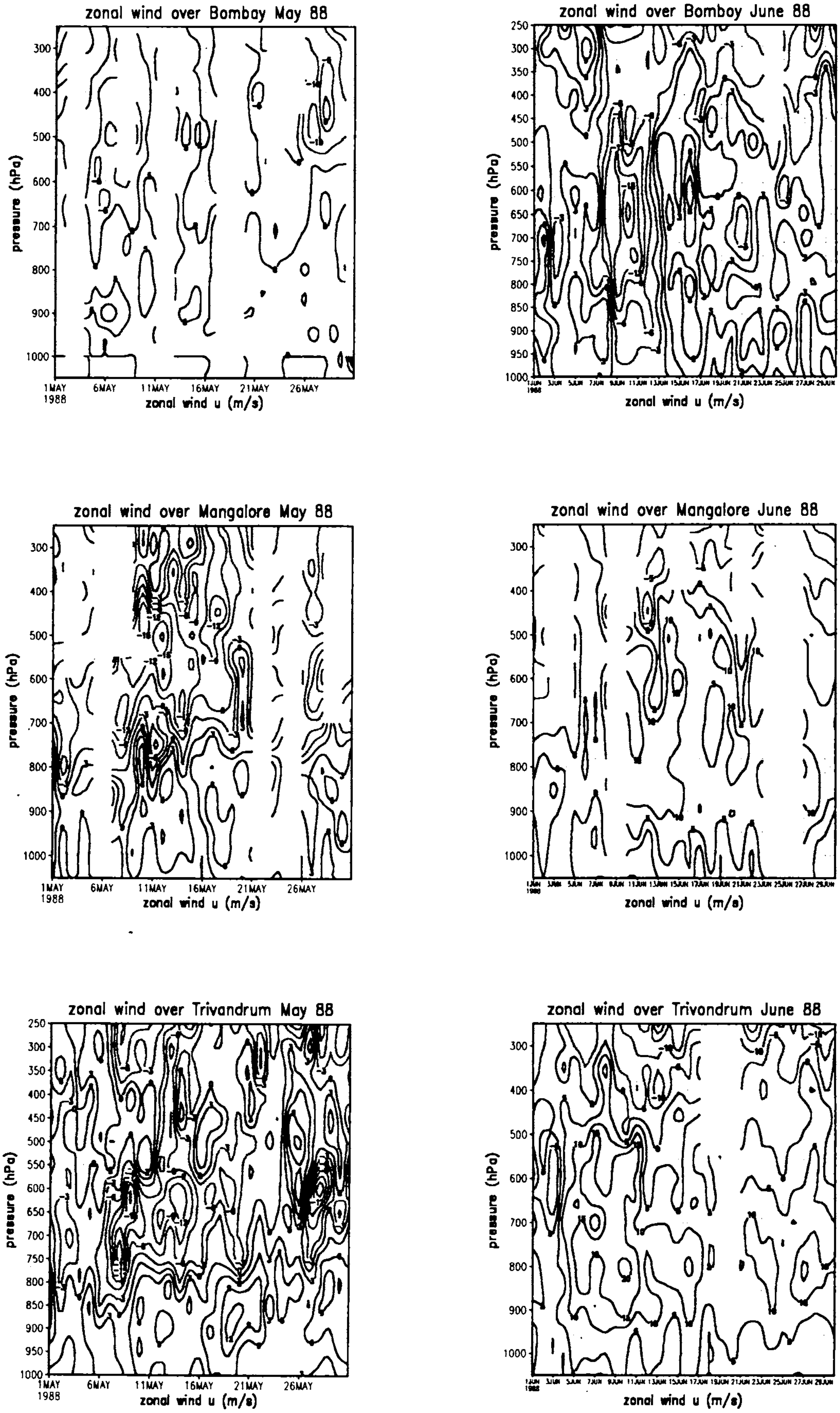


Fig. 5.5e Zonal wind structure over Bombay, Mangalore and Trivandrum during May-June 1988

during May over Mangalore. The wind structure during June is not clear because the data is missing over the station in most of the days. At Bombay the onset was on June 9th on which the westerly wind reaches up to 600hPa and the depth increases further. The easterly wind from surface to 250hPa prior to onset is because of the depression in the Arabian Sea during the period. In 1985 the onset date was on May 28th. Westerly reaches even up to 250hPa or above one week prior to onset over Trivandrum. A maximum speed of $15ms^{-1}$ or more is noticed around 850hPa on the onset day. At Mangalore also the westerly depth increases prior to onset and by onset it reaches up to about 400hPa. By the onset of monsoon the westerly speed and depth over Bombay also increases. A maximum speed of $12ms^{-1}$ is noticed on June 8th between 800hPa and 900hPa. Westerlies are seen up to 250hPa or above. In 1986 easterlies are seen from about 800hPa upwards in May at Trivandrum. The westerlies gradually deepened by onset, which was on June 4th. A maximum westerly speed of about $10ms^{-1}$ is noticed around 850hPa level soon after the onset. Before the onset the westerly depth and speed were less than that after the onset. At Mangalore the onset was on June 5th. By onset the depth of westerly wind increases. Before the onset easterlies prevail upward from about 700hPa which gradually changes and westerly reach higher levels. Westerly speed also increases and a maximum of $10ms^{-1}$ is noticed at 800hPa soon after the onset. At Bombay in 1986 the onset was on June 20th. By June 17th the westerly depth suddenly increases and reaches up to 250hPa or above. A maximum speed of $15ms^{-1}$ or more is noticed around 850hPa on the onset day. The lower tropospheric westerly depth increases suddenly by June 1st and reaches up to 400hPa or above at Trivandrum where the onset was on June 2nd in 1987. Westerly speed increases and a maximum of $15ms^{-1}$ is noticed at 850hPa during the onset day. In May the westerly wind which was seen up to 700hPa deepened by the onset which was on June 3rd over Mangalore. The speed, which was about 3ms-1 during late May, increases and reaches a maximum of about $15ms^{-1}$ at 750hPa during onset. At Bombay the onset was on June 15th. The westerly depth and speed increases by onset over the station. The maximum speed of about $9ms^{-1}$ is noticed at 800hPa level. In 1988 the onset was on May 26th. The westerly depth gradually increases from 800hPa level to 300hPa level prior to onset over Trivandrum. Maximum speed is reached after the onset and the speed is about $15\text{-}20ms^{-1}$ at 800hPa. At Mangalore the onset was on June 2nd in 1988. The westerly depth, which was up to 700hPa during end, May increases after the onset only. At the time of onset maximum of about $10ms^{-1}$ is noticed around 850hPa level. At Bombay easterlies are noticed prior to onset because of the depression in the Arabian Sea during the period. By the onset day the easterlies vanishes and westerly wind depth reaches 300hPa level. A maximum speed of $9ms^{-1}$ is noticed near 850hPa soon after the onset.

The vertical structure of zonal wind during three days before and after the onset and on the onset day for Mangalore, Trivandrum, Minicoy, Amini, Bangalore, Goa, Bombay, Jodhpur, Calcutta, Guwahati and Lucknow are given in figures (5.6a to 5.6b). At Mangalore the westerly depth increases after the onset over the station, which was on June 2^{nd}. The speed increases and is $>10ms^{-1}$ at 850hPa soon after the onset. But at Trivandrum the depth of westerly wind increases before the onset and reaches 350hPa level by the onset day. At Minicoy the westerly depth and speed increases by onset. A maximum speed of $12ms^{-1}$ or more is noticed near the surface on the onset day and westerly even reaches 350hPa. But at Amini only slight increase in depth is noticed and the speed also does not vary much. At Bangalore which is located in the leeward side of Western Ghats the depth is not found to vary but the speed increases and is more than $12ms^{-1}$ at 850hPa on the onset day. Westerly depth increases soon after the onset at Bombay. Prior to the onset easterly prevails from surface to upper troposphere because of the depression in the Arabian Sea during 9 to 12 June. The westerly, which is up to 600hPa before the onset at Goa gradually decreases because of the easterlies as a result of the low pressure in the Arabian Sea which, developed into a depression. On the onset day the westerly speed increases and was more than $9ms^{-1}$ at 850hPa level. At Jodhpur which lies at the dry end of the monsoon trough also the westerly depth increases prior to onset. The speed also increases and a maximum of more than $12ms^{-1}$ are noticed on the onset day. Easterlies are noticed on the onset day at Calcutta because of the depression, which formed over the head Bay during June 9 to 10. After 10^{th} June westerly is noticed over the station. At Guwahati where the Bay of Bengal branch of monsoon current reached on May 30^{th} the westerly wind is noticed even at 250hPa prior to the onset. Wind speed of more than $10ms^{-1}$ is noticed at 850hPa on the day before onset. Westerlies, which prevailed prior to onset over Lucknow, weakened and easterlies appeared at the surface on the onset day. Weak westerlies appeared in the lower troposphere after the onset. Thus we can say that the westerly depth and speed increases at each station at the time of onset of monsoon over the station. Therefore as the monsoon advances over the country the surface easterlies are pushed up by westerlies and they even reach upto 250hPa or above prior to the onset or on the onset day over each

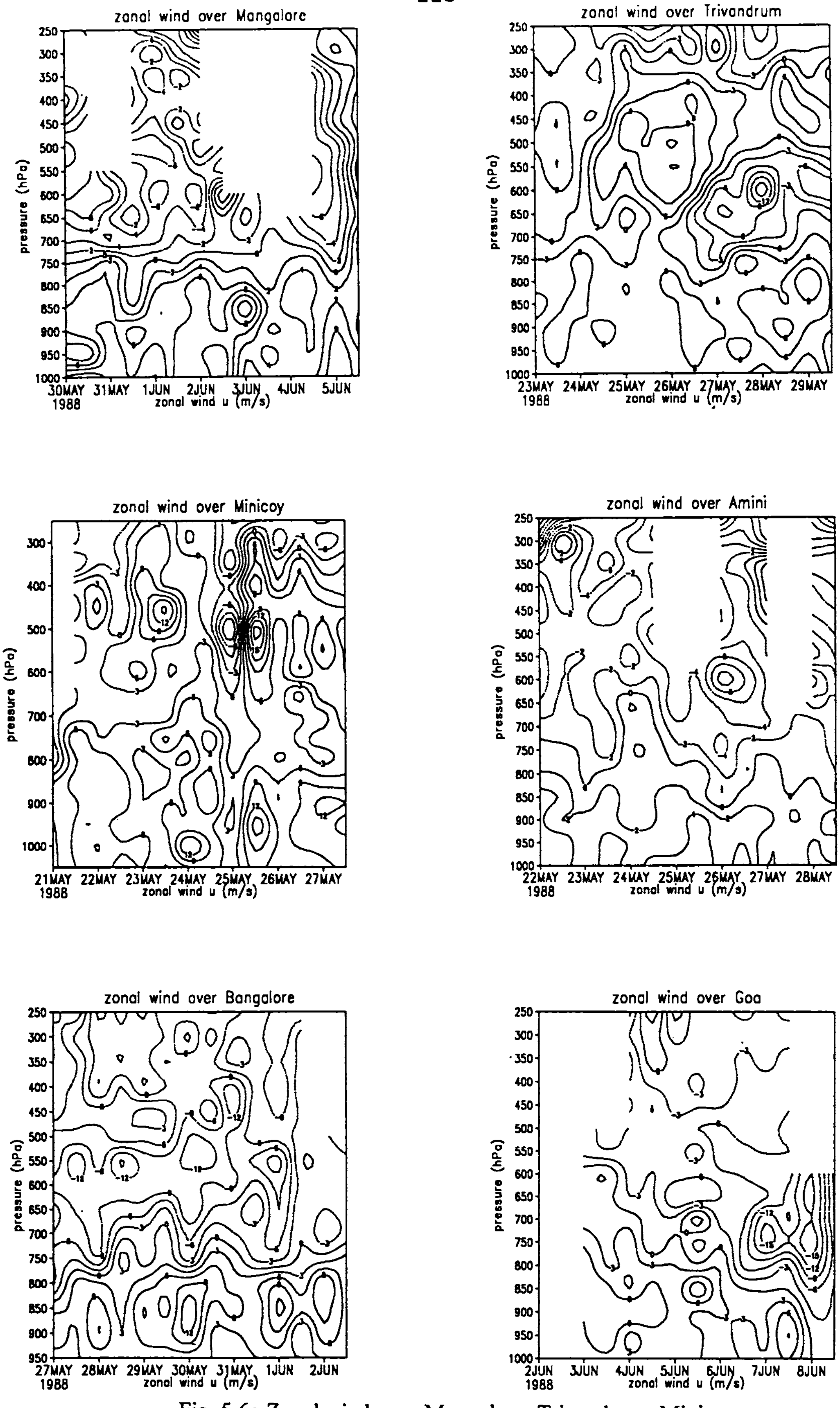


Fig. 5.6a Zonal wind over Mangalore, Trivandrum, Minicoy, Amini, Bangalore and Goa during onset

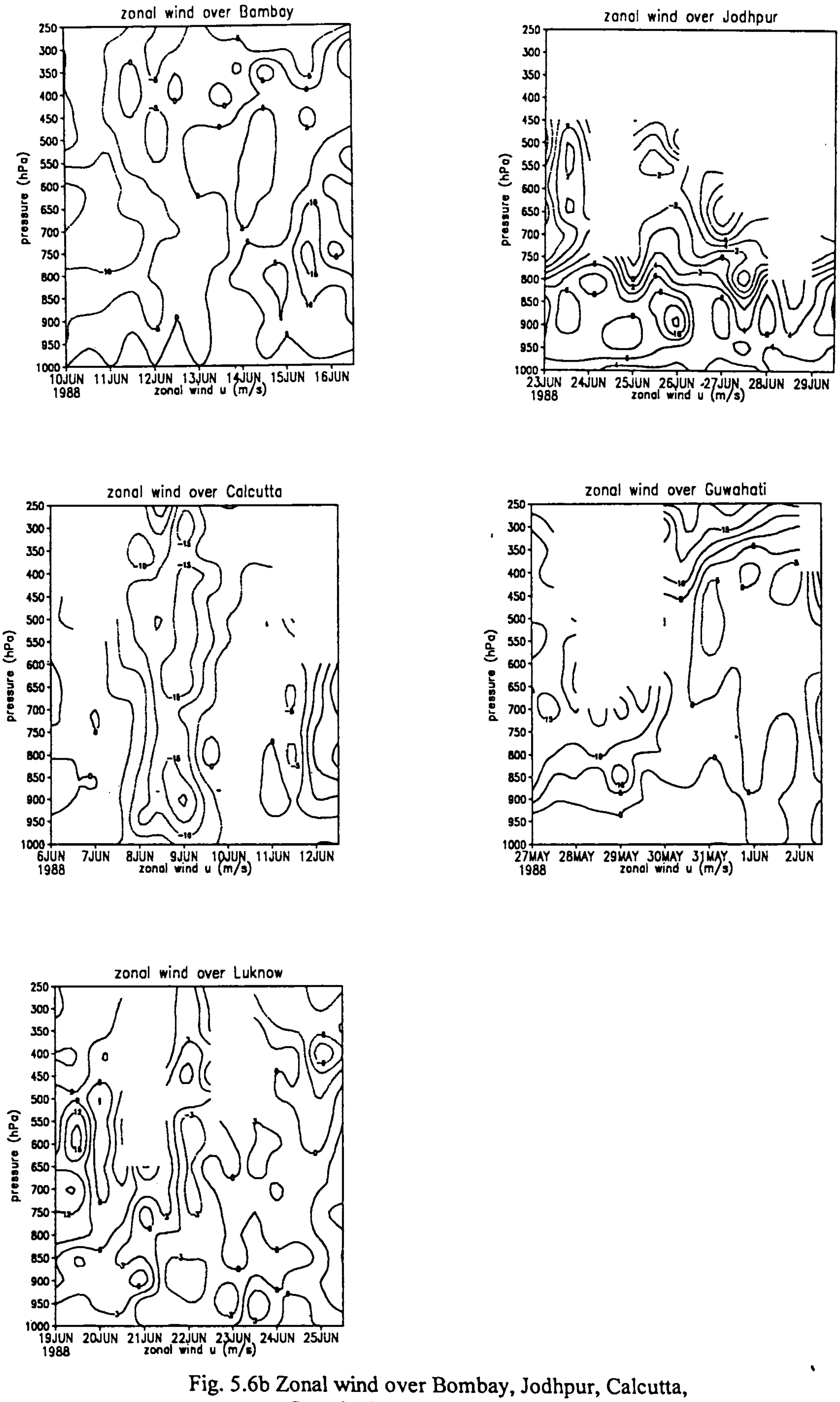
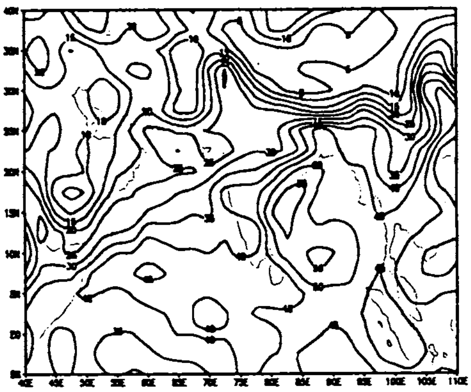


Fig. 5.6b Zonal wind over Bombay, Jodhpur, Calcutta, Guwahati and Lucknow during onset

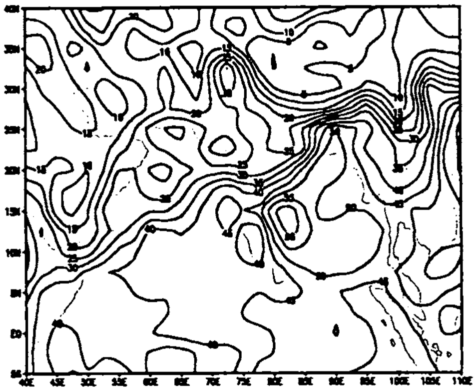
station. Also the westerly speed is found to become maximum at the time of onset over the station. Ananthakrishnan et al (1968) noticed this phenomena at Trivandrum during the onset but the speed and depth of westerly wind is found to increase at each station as the monsoon reaches the station.

Variation of precipitable water content during the onset of monsoon.

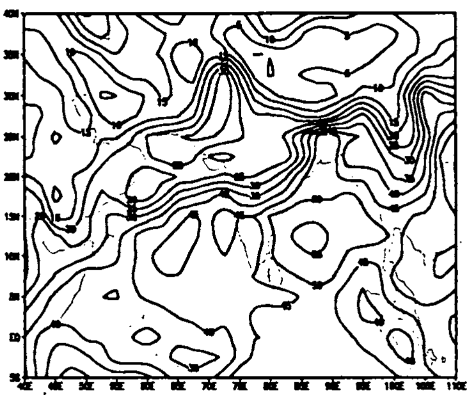
Total Precipitable water vapour content of the atmosphere from surface to 100hpa level over the Indian region during the onset for four years from 1988 to 1991 are given in figures (5.7a to 5.7d). The figures are for six days, which is for three days before the onset day, onset day and two days after the onset over the subcontinent. In 1988 the onset of monsoon was on May 26th. On 24th May the water vapour content in the Arabian Sea is about 40kgm^{-2}. At about 15°N near the coast the content is about 45kgm^{-2}. In the Bay of Bengal the water vapour content is found to be between 50 – 55kgm^{-2}. The content in the Arabian Sea region increases and on the onset day it is about 45kgm^{-2} over the south Arabian Sea. Not much variation is noticed over the Bay of Bengal region and over the subcontinent. The content further increases over the Arabian Sea after the onset. The onset was on 3rd June in 1989. On 31st May the precipitable water content in the south Arabian Sea is about 45kgm^{-2}. An amount of 50kgm^{-2} is noticed at two pockets, one near Myanmar coast and another to the south east of Srilanka. Along the east coast and southern most region of peninsula it is greater than 40kgm^{-2}. By June1st the amount increases and is about 55kgm^{-2} at central Arabian Sea and about 45kgm^{-2} at southernmost *peninsular region. To the south of 20°N over the subcontinent it is about 35kgm^{-2} or* more. The amount further increases over the Srilankan region on June 2nd and is about 50kgm^{-2}. On the onset day the area having 50kgm^{-2} water vapour increases and to the south of 20°N over the region is having an amount greater than 40kgm^{-2}. After the onset the amount over West Bengal region and over the Bay of Bengal region increases. In 1990 the onset of monsoon was on 25th May. From the figures it is clear that as the onset approaches the precipitable water vapour amount in the Arabian Sea and adjacent Indian Ocean region increases. The amount which is about 50kgm^{-2} three days before onset decreases and then increases to 55kgm^{-2} on onset day and after. In the subcontinent the water vapour content, which is about 35-40kgm-2 from 20°N downwards, increases and becomes about 40-45kgm^{-2}. The precipitable water content in the head Bay region also increases from 45kgm^{-2} to 50kgm^{-2} after the onset. The onset was on May 29th in 1991.



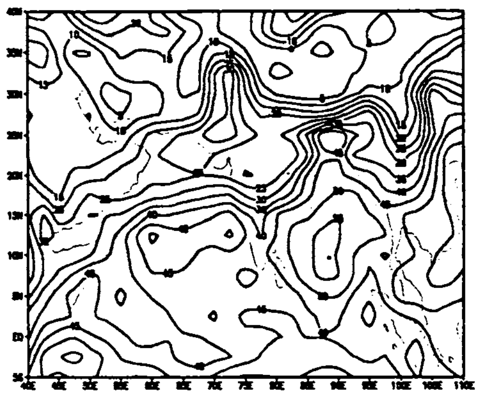
a) 3days before onset



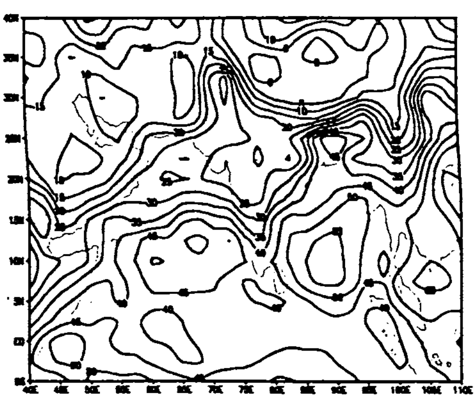
b) 2days before onset



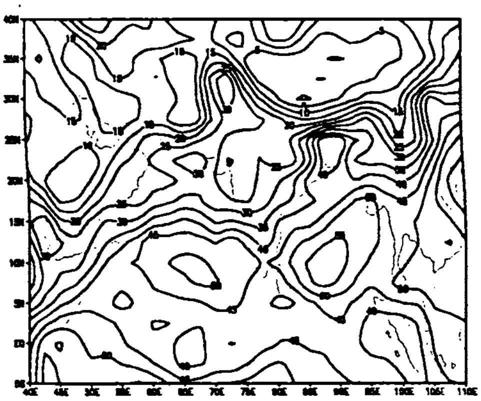
c) day before onset



d) onset day

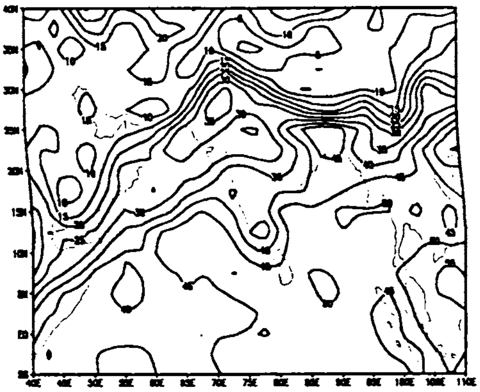


e) day after onset

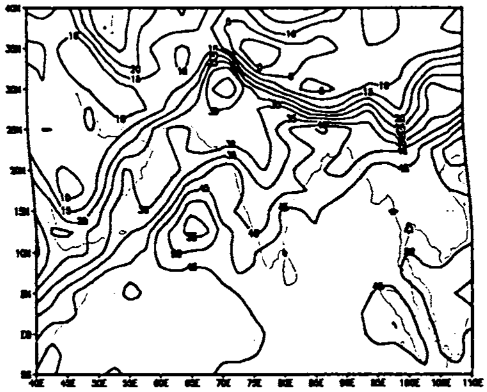


f) 2days after onset

Fig. 5.7a Precipitable water vapour content 1988



a) 3days before onset



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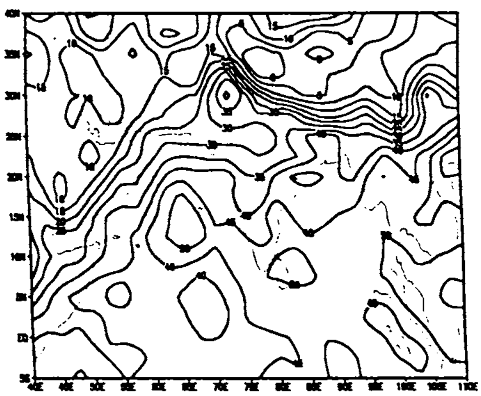
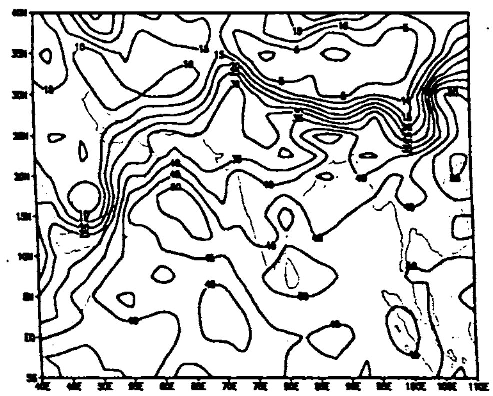
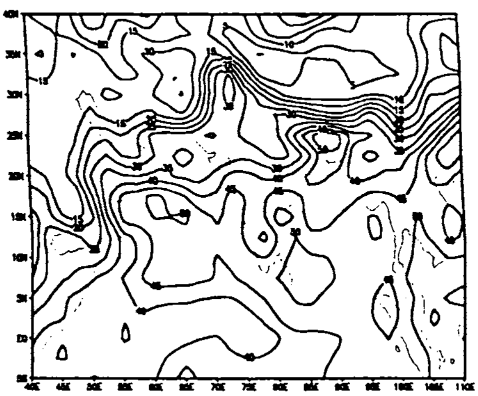


c) day before onset



d) onset day



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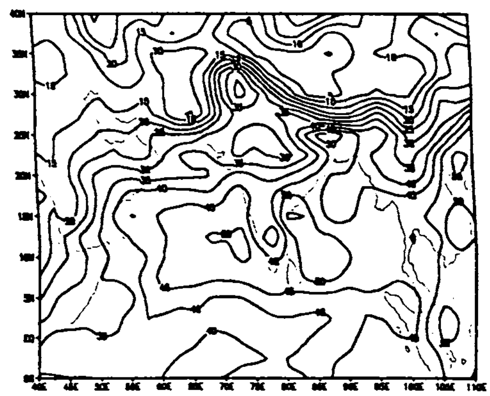
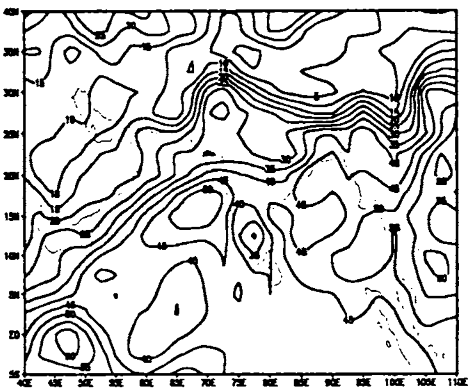
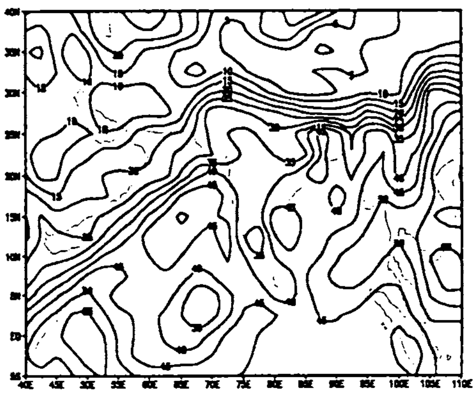


f) 2days after onset

Fig. 5.7b Precipitable water vapour content 1989



a) 3days before onset



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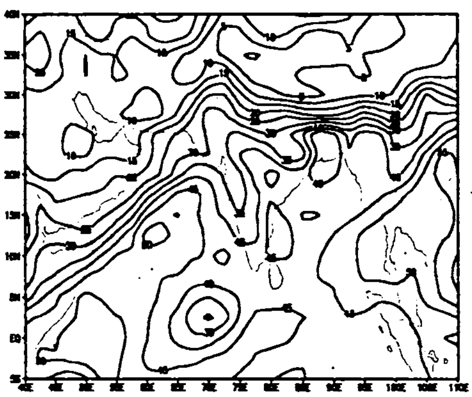

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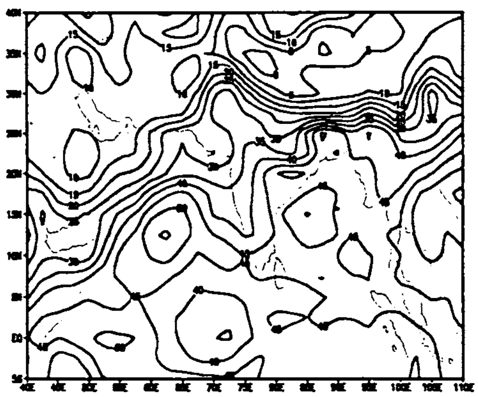
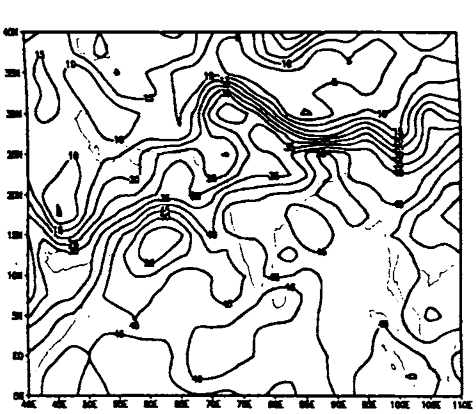


d) onset day



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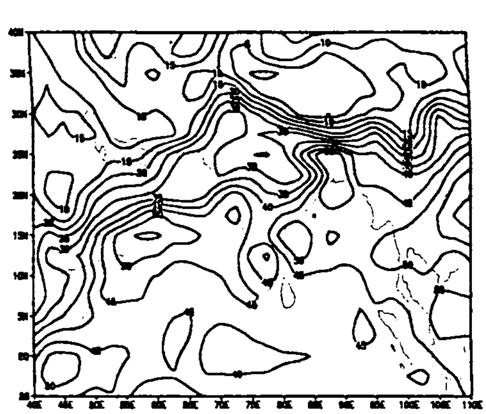
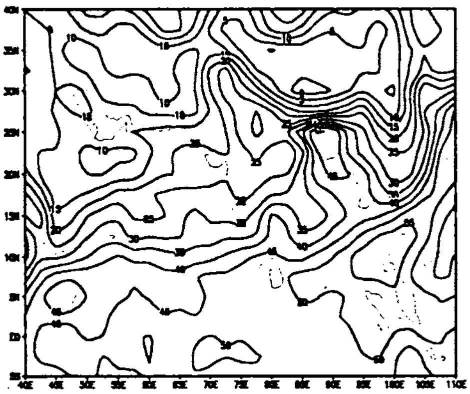
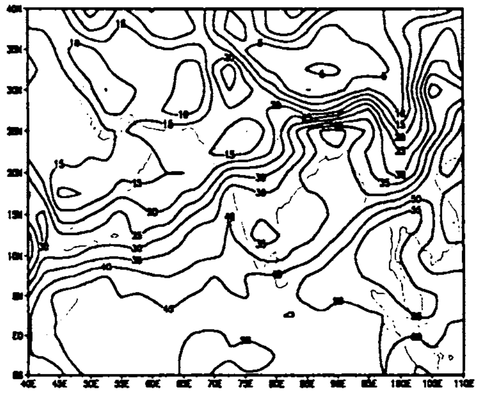


f) 2 days after onset

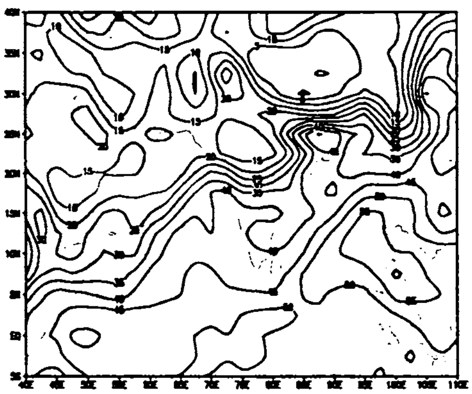
Fig. 5.7c Precipitable water vapour content 1990



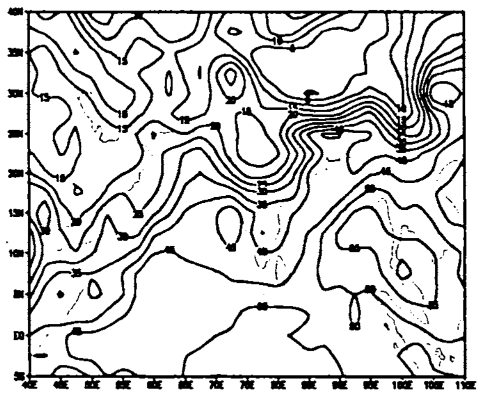
a) 3days before onset



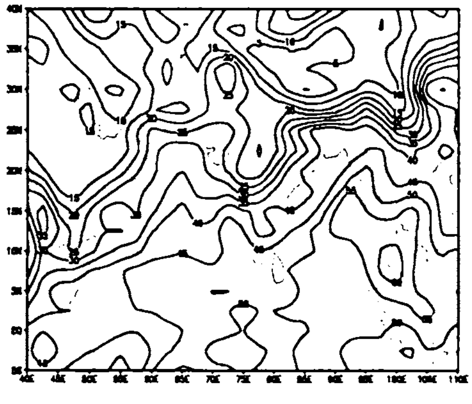
b) 2days before onset



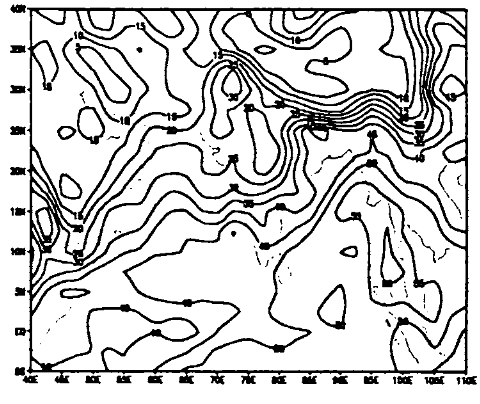
c) day before onset



d) onset day



e) day after onset



f) 2 days after onset

Fig. 5.7d Precipitable water vapour content 1991

Before the onset the amount of precipitable water vapour ranges between 25-45kgm^{-2} to south of 20°N on 26th May in the Arabian Sea. The amount increases in the Arabian Sea region off the west coast and becomes about 45kgm^{-2} over a large area. In the Bay of Bengal region the amount which is between 30 to 45kgm^{-2} near the east coast of India becomes greater than 40-45kgm^{-2} on the onset day. In the subcontinent south of 15°N the amount is about 30kgm^{-2} three days before the onset increases to about 35kgm^{-2} or more by the onset. The precipitable water content over Bay of Bengal and adjacent Indian Ocean region also increases after the onset. Therefore the precipitable water vapour content in the atmosphere is found to increase at the time of onset over the Indian region. The increase is more significant over the Arabian Sea, which is the region through which the monsoon current blows towards the sub continent. Over the land region the variation is not much significant as in the oceanic region.

The CAPE and CINE values during the various periods indicate that during the pre-monsoon and onset periods the atmosphere is highly favourable for the initiation of convection. During the post-monsoon season the atmosphere over the stations are highly stable most of the days. A positive correlation exists between CAPE and occurrence of thunderstorm activity. High CAPE is not the only necessary factor for the convection to set in, the CINE also controls the convective activity. CINE values increase in association with the surface stability after the rainfall from thunderstorms. The monsoon activity affects the thermodynamic structure by lowering the LCL. The vertical wind structure shows an increase in westerly depth and speed over each station, as the monsoon is onset over the station. An increase in precipitable water content of the atmosphere occurs over the Indian region especially the Arabian Sea area during the onset of monsoon.

CHAPTER 6

Conclusions

The surface layer fluxes and stability are found to vary during various seasons. Higher momentum flux is noticed during the pre-monsoon and onset periods than post-monsoon season. This is as a result of the high-speed winds during the pre-monsoon and monsoon season. Sensible heat flux is high during certain days in the pre-monsoon months because of higher temperature during the period. By the onset the increase in wind speed causes the increase in sensible heat flux. During the post-monsoon months the sensible heat flux increases from October to December due to increase in surface temperature after the monsoon activity. The latent heat flux is found to be higher during onset because of the increase in evaporation due to high-speed winds. In the post-monsoon period higher latent heat flux is noticed during October due to higher winds and rainfall in association with the withdrawal of monsoon. After that the latent heat flux decreases and turns downward due to the reduced availability of moisture at the surface for evaporation. The surface stability shows a tendency of the atmosphere to become less unstable during onset from highly unstable situation during pre-monsoon and it tends to become highly unstable after the withdrawal of monsoon. Less unstable or near neutral situation during monsoon and withdrawal period is due to radiative cooling of the surface because of overcast sky and moist winds during the period. This indicates the influence of monsoon activity on the surface layer characteristics.

The occurrence of mesoscale and synoptic scale disturbances and the surface layer characteristics are found to have close linkage. The atmosphere becomes highly unstable prior to the occurrence of the disturbances. Due to the increased instability the surface turbulence also increases. High surface fluxes are noticed during the occurrence of these disturbances due to the increase in turbulence. The impact of synoptic scale disturbance on the surface layer characteristics is not only near the region of its formation and along its track but also on the entire subcontinent.

The drag coefficient computed indirectly also shows variation during various seasons. Higher drag coefficient occurs during pre-monsoon and shows a gradual decrease by monsoon onset and it increases gradually during the post-monsoon after the withdrawal of monsoon. The decrease in drag coefficient during monsoon is due to the increase in wind speed. The drag is found to decrease with increasing wind speed. It is also found to decrease when the atmosphere becomes less unstable or near neutral. Steep increase in drag coefficient occurs when the wind speed is less then $4ms^{-1}$ at the west coast stations. The drag coefficient during less unstable or near neutral condition are much less than that given by Garratt (1977).

The surface fluxes are higher during most of the days during the northeast monsoon period over Bay of Bengal during the BOBMEX Pilot experiment. This is due to the increase in turbulence over the region due to the depression, low-pressure area and deep convective activity during the period. Though the fluxes computed by various methods shows more or less similar trend the magnitudes are not same. The eddy correlation method using fast response data from sensors which gives data at 0.1s interval will be more accurate than the other two methods. The computed drag coefficient gives a higher value than that used in the bulk aerodynamic method. Also the drag was higher during the occurrence of disturbances and deep convection. The drag coefficient over the marine boundary layer is much less than that over the land boundary layer.

The thermodynamic structure of the atmosphere shows variation during various seasons. The onset of monsoon causes lowering of the Lifting Condensation Level, which was at higher levels during pre-monsoon. During monsoon due to moist winds and rainfall the atmosphere becomes more humid therefore a slight lifting of air parcel will leads to saturation and then condensation. After the monsoon during post-monsoon the LCL is found at higher levels. That is the atmosphere will be less humid during post-monsoon so that an air parcel needs to be lifted to higher levels for saturation and condensation. During the pre-monsoon and onset period CAPE was higher than CINE which indicates that the atmospheric condition is favourable for supply of energy to an air parcel for initiation of convection. During post-monsoon CINE was high in most of the days, which indicates highly stable lower atmosphere. During the occurrence of thunderstorms high CAPE and low CINE values are noticed. This indicates that lesser amount of energy is required for the air parcel to reach LFC after which the parcel rises on its own and high amount of energy is supplied to the parcel by the atmosphere. This leads to convection and thunderstorm occurs.

The wind structure during monsoon onset shows an increase in depth and speed of westerly wind. The westerly wind reaches even up to 250hPa or above at the time of onset. As the monsoon advance over the country the surface easterly winds are pushed up the westerlies and they reach even upto 250 hPa prior to onset or on the onset day of monsoon at each station. The westerly speed is also found to increase on the onset day. By the onset the precipitable water content over Arabian Sea and adjacent region increases. The increase is significant over the oceanic region especially the Arabian Sea region through which the monsoon current flows towards the subcontinent.

Therefore in general the boundary layer and thermodynamic characteristics over the subcontinent varies during different seasons and during the occurrence of disturbances. This study gives a better understanding of the boundary layer and thermodynamic structure and their variations during various periods, which can be used in the modelling studies.

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