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 Meteorological 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 20 ms\(^{-1}\) 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

\[
\frac{dw}{dt} = g \left( \frac{T_p - T_e}{T_e} \right)
\]

(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 \( \text{CAPE} = (dw/dt) dz \) \hspace{1cm} (5.2)

But from hypsometric equation we can write

\[ dz = RT \ln(P_1/P_2)/g \] \hspace{1cm} (5.3)

where \( T \) is the temperature of the environment i.e. \( T_e \)

Therefore \( \text{CAPE} = R( (T_p - T_e) \ln(P_1/P_2) ) \) \hspace{1cm} (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 \( (T_p - T_e) \) 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

\[ \text{CAPE} = \int_{LFC}^{LOC} R(T_p - T_e) \frac{dp}{P} \] \hspace{1cm} (5.5)

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

CINE is computed using the following equation

\[ \text{CINE} = \int_{LFC}^{\text{surface}} R(T_p - T_e) \frac{dp}{P} \] \hspace{1cm} (5.6)

Computational procedure is same as that as CAPE by taking \( dp = 1 \) hPa 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 temperature \( \theta_e \) for \( \theta \) is obtained as (Holton 1992)
\[ \theta_e = \theta \exp \left( (L \frac{q}{(C_p T)}) \right) \] (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 \( \theta_e \) to an accuracy of 0.0018°C.

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

where \( T_{LCL} \) is the dry bulb temperature at LCL, \( \theta_{LCL} \) is the potential temperature at LCL.

\[ \theta_{LCL} = T \left( \frac{1000}{(p-e)} \right)^{0.2854} (T/T_{LCL})^{0.00028w} \] (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

\[ q = \frac{0.622e}{(P-0.378e)} \] (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

\[ e = 6.11 \exp \left( A( T - 273.16 )/( T-B ) \right) \] (5.11)

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

\[ A = 21.87 \; \text{when} \; T < 263 \]
\[ A = 17.26 \; \text{when} \; T \geq 263 \]
\[ B = 7.66 \] when \( T < 263 \)
\[ B = 35.86 \] 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

$$u = -ff \sin (dd)$$

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 precipitable 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 up to 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/CINE and 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 Trivandrum 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 15-20ms\(^{-1}\) 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 Sri Lanka. Along the east coast and southern most region of peninsula it is greater than 40kgm\(^{-2}\). By June 1st 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.
Fig. 5.7a Precipitable water vapour content 1988

a) 3 days before onset

b) 2 days before onset

c) day before onset
d) onset day

e) day after onset
f) 2 days after onset
Fig. 5.7b Precipitable water vapor content 1989
Fig. 5.7c Precipitable water vapour content 1990

a) 3 days before onset
b) 2 days 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

a) 3 days before onset

b) 2 days before onset

c) Day before onset

d) Onset day

e) Day after onset

f) 2 days after onset
Before the onset the amount of precipitable water vapour ranges between 25-45kgm\(^{-2}\) to south of 20°N on 26\(^{th}\) 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 subcontinent. 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.