CHAPTER 6

ASSOCIATION OF CIRRUS WITH UPPER TROPOSPHERIC TURBULENCE

6.1. INTRODUCTION

Strong convective turbulence and very low temperatures prevailing in the upper troposphere is conducive for condensation of vapours under favoured conditions leading to cirrus formation. Turbulence in cirrus clouds is linked to the dynamical state of the ambient flow field in which the clouds are embedded [Quante and Starr, 2002]. Turbulence is directly linked to the life cycles of the clouds through internal mixing and entrainment processes. Smith and Jonas [1996] suggested that there are a variety of atmospheric processes, which may contribute to variations in the vertical velocity at certain scales. These include gravity waves, Kelvin-Helmholtz waves and buoyancy effects. Shear production and buoyancy effects are believed to be the predominant factors in upper tropospheric turbulence associated with cirrus cloud systems. Turbulent Kinetic Energy (TKE) may be mechanically produced by extraction of kinetic energy from the mean flow in the presence of wind shear (i.e., Kelvin-Helmholtz instability) or by gravity wave breaking. Buoyancy production of TKE occurs through radiative heating and latent heat release that influence the heat flux.

The turbulent kinetic energy (TKE) dissipation rate, \( \varepsilon \), and eddy diffusion coefficient, \( K_m \), are indicative of the strength of prevailing tropospheric turbulence. The rate of turbulent kinetic energy dissipation has been estimated within the cirrus cloud systems by a number of workers. It has been found that \( \varepsilon \) is greater in the cirrus cloud compared to that in the cloud free region and strong turbulence in patches has been found to occur in the cirrus cloud [Quante and Starr, 2002]. Almost all these studies correspond to mid and high latitudes [e.g., Gultepe and Starr, 1995; Schumann et al., 1995]. Using a detailed cirrus cloud model to evaluate the physical processes responsible for the formation of sub-visual cirrus near tropical tropopause, Jensen et al. [1996b] found that two distinct formation mechanisms are viable: (i) remnants of the cumulonimbus out flow anvils and (ii) in situ nucleation of ice crystals by homogeneous nucleation or condensation. Of these, the second one i.e., the role of
shear driven turbulent mixing in the formation of cirrus clouds through condensation/nucleation is examined in the following. Model studies have indicated that turbulence plays a significant role in cloud evolution during the maintenance and dissipation period of the cirrus cloud life cycle [Gu and Liou, 2000]. Shear driven turbulence can be characterized by the eddy diffusion coefficient ($K_m$) for momentum [Hocking, 1989]. A negative altitude gradient of $K_m$ ($K_m$ decreasing with altitude) in the presence of a negative altitude gradient of the concentration of minor species (like water vapour, aerosols) leads to accretion/convergence of the species [Krishna Murthy, 1987]. Thus, altitude gradient of $K_m$ is a pertinent parameter to be examined in this regard.

Role of vertical wind velocity and turbulence in the upper troposphere on the cirrus formation are examined using lidar-derived cirrus parameters and MST radar derived vertical wind and turbulence in this chapter. The turbulence parameters like turbulent kinetic energy dissipation rate and the eddy diffusion coefficient are estimated to study the role of turbulence on cirrus formation mechanism. Of the 281 nights of lidar observations during March 1998 to June 2002 simultaneous MST radar observations were available on 120 nights. The lidar measurements reveal the presence of cirrus on 229 nights out of which simultaneous MST radar observations (of temperature) were available on 82 nights.

6.2. VERTICAL WIND IN THE UPPER TROPOSPHERE

Altitude profile of backscattered signal from MST radar, operated in vertical beam mode with an altitude resolution of 150 m, at every 45 sec for a duration of ~2 hrs (from 22:00 IST to 24:00 IST) are Fourier analysed to derive Doppler spectra. The first moment of the Doppler spectra provides the mean Doppler frequency from which the vertical wind velocity is estimated. This provides a time series of vertical wind velocity ($w$) sampled at 45 sec interval in the altitude range of 3.75 to 25 km at 150 m resolution. The details of the estimation of vertical wind velocity are presented in Chapter 2. This vertical wind velocity is an indicator of turbulence structure in troposphere. The altitude and temporal structure of $w$ is examined along with lidar data to study the association between tropospheric turbulence and cirrus formation.

Figures 6.1 and 6.2 shows the contour plots of backscatter ratio profiles for copolarized ($R_p$) and cross-polarized ($R_s$) components derived from lidar data along with the contour plot of vertical wind velocity derived from MST radar data (during the
The contour plots of backscatter ratio ($R_p$ and $R_s$) on 26 January 1999 depict the presence of cirrus in the altitude region 13-15 km. As described in Chapter 3, $R_p$ or $R_s$ $\geq 2$ is taken as the presence of cloud even though two levels of contours below 2 also are presented in these figures. It is seen that cloud is present throughout the period from 23:00 IST to 01:00 IST. After midnight the cloud is rather weak. The vertical wind velocity is shown in grey scale. It is seen that vertical velocity ($w$) is very high, in the altitude region where cirrus cloud is located, reaching a maximum value up to 0.25 m s$^{-1}$. The vertical wind velocity is not seen to be consistently high in the cloud altitude region. Intermittency of vertical velocity ($w$) is clearly seen from this figure (very high positive value and negative value, i.e., 0.25 to $-0.35$ m s$^{-1}$). Figure 6.2 shows the contour plots of $R_p$, $R_s$ and $w$ for the night 29 January 1999. The backscatter ratios show the presence of the cirrus in the altitude region 14-15 km, which is strong around 23:00 IST. The cloud is weak in the co-polarized channel. The vertical wind velocity ranges from $-0.45$ to 0.35 m s$^{-1}$ with high positive and negative values near the cloud base. This shows that a large shear in $w$ exists in the altitude region 12-16 km where cirrus clouds are seen frequently. At altitudes above and below this region, the temporal variation in $w$ is rather small.

Models [Jensen and Toon, 1994; DeMott et al., 1997] have demonstrated that ice number density generated by homogeneous freezing of aerosols increases rapidly with increasing vertical wind speed. It is expected that high ascent rates would lead to more ice crystals. Also large vertical wind speed would cool the tropopause region [Jensen et al., 2001a]. Gültepe et al. [1995] measured the vertical velocity of wind using a Doppler radar during First International Satellite Cloud Climatology Project Regional Experiment (FIRE) II, over Kansa region. They found intense turbulence and strong small-scale structures in cirrus cloud with vertical velocities of about ±1.5 m s$^{-1}$ as typical case. But vertical velocities from large-scale data analysis during FIRE II are found to be generally between 0.01 and 0.02 m s$^{-1}$ with high values reaching up to 0.1 m s$^{-1}$. Starr and Cox [1985a,b] and Zhang et al., [1992] have shown that large-scale vertical wind of about 0.01 to 0.02 m s$^{-1}$ can play an important role in cirrus development and showed that vertical velocity of this magnitude range can significantly change cloud radiative and microphysical characteristics. These studies indicate the importance of dynamical activity in the generation of cirrus.
Figure 6.1 Contour plots of backscatter ratios for co-polarized and cross-polarized components ($R_p$ and $R_s$) along with the vertical wind velocity ($w$) on 26 January 1999

Figure 6.2 Contour plots of backscatter ratios for co-polarized and cross-polarized components ($R_p$ and $R_s$) along with the vertical wind velocity ($w$) on 29 January 1999

The present study shows that the magnitude of vertical wind velocity observed at this tropical station is higher than that observed in the mid-latitudes.

6.3. Estimation of Turbulence Parameters from Radar Data

The vertical velocity is an indicator of turbulence intensity [Gultepe et al., 1995]. The turbulence parameters like turbulent kinetic energy (TKE) dissipation rate, $\varepsilon$, and eddy diffusion coefficient, $K_m$, are estimated using the vertical wind data obtained
from the MST radar. Owing to the high temporal resolution of radars, there are three methods of deriving the turbulent parameters from radar observations. These are (a) the Doppler spectral width method, (b) radar backscattered signal power method and (c) the wind variance method. Among these the wind variance method adopting the procedure developed by Satheesan and Krishna Murthy [2002] is followed in the present study, which is described briefly in the following. An advantage of the wind variance method over the other methods is that all the required data for estimation of $\varepsilon$ are available from the MST radar itself.

The vertical wind observation on each night is subjected to Fourier transform and the resulting amplitude-frequency spectrum is converted to power-frequency spectrum. The wild points in the data are removed by visual inspection and the data gap is filled by linear interpolation before applying Fourier transform. The power spectrum at each altitude is examined to identify the Brunt-Vaisala (B-V) frequency, $N$ (explained in Chapter 2). The power spectrum of the vertical wind consists of two distinct ranges of frequencies demarcated by the Brunt-Vaisala frequency. These are the inertial subrange of turbulence and the buoyancy range or the range in which wave propagation can exists. This method essentially consists of estimating the wind variance from the temporal spectra of vertical wind at frequencies greater than $N$, by integrating the power spectrum of the vertical wind from the B-V frequency to Nyquist frequency. The Nyquist frequency, $f_n$, is given by

$$f_n = \frac{2\pi}{2 \Delta t}$$

(6.1)

where $\Delta t$ is the sampling interval (45 s for the present case). The frequencies greater than the Nyquist frequency are not considered due to its negligible contribution [Satheesan and Krishna Murthy, 2002]. Altitude profiles of $\varepsilon$ and $K_m$ are obtained from the altitude profile of vertical wind variance ($\overline{v^2}$). The TKE dissipation rate $\varepsilon$ is obtained as [Hocking, 1989; Weinstock, 1981; Fukao et al, 1994, Satheesan and Krishna Murthy, 2002],

$$\varepsilon = 6.1F \overline{v^2} \frac{N}{2\pi}$$

(6.2)

where $F$ is the fraction of the measured velocity variance that resides in the inertial subrange and the rest in the buoyancy subrange. In the present study the vertical wind variance is obtained over the frequency range from $N$ to the Nyquist frequency i.e.,
the contribution from the inertial subrange alone is considered. Hence in the estimation of $\varepsilon$, the factor $F$ is taken as unity. The vertical eddy diffusivity, $K_m$ is estimated using the equation

$$K_m = \frac{0.8\varepsilon}{N^2} \quad (6.3)$$

From the altitude profile of $N$ (obtained from power–frequency spectrum analysis), the altitude profile of turbulence parameters like $\varepsilon$ and $K_m$ are estimated using Equations (6.2) and (6.3) respectively.

6.4. EXPERIMENT CAMPAIGN TO STUDY THE ASSOCIATION OF CIRRUS WITH TURBULENCE

Two intensive Tropical Cirrus Experiment (TCE) campaigns have been conducted to study the day-to-day variability in the characteristics of upper tropospheric turbulence and its association with changes in the altitude structure of tropospheric turbulence. During these campaigns lidar and MST radar are operated regularly on consecutive nights. The first campaign (TCE-1999) was conducted during January-March period of 1999 and the second campaign (TCE-2000) during February-March period of 2000. The TCE-1999 campaign was conducted on 47 consecutive nights during 18 January to 5 March in which lidar was operated for about 6 hrs from 22:00 IST and MST radar was operated for 2 hrs from 22:00 to 24:00 IST. The second campaign (TCE-2000) was conducted on 40 consecutive nights during the period 21 February to 31 March 2000 in which lidar was operated for about 3 hrs and MST radar was operated for 2 hrs from 20:00 IST. Lidar measurements were used to study the scattering properties of cirrus.

6.4.1 Frequency of occurrence of cirrus during the TCE campaigns

6.4.1.1. Tropical Cirrus Experiment-1999 (TCE-1999)

During TCE-1999 lidar observations were made continuously for ~6 hrs, which consists of more than 100 profiles (each 250 s of lidar observation yields one profile) on each night. Out of the planned 47 nights, lidar data was available on 41 nights. On the other six nights lidar could not be operated either due to the presence of low-level clouds or due to technical problems in the lidar system. The duration of cirrus cloud is indexed based on the number of profiles in which cirrus clouds were present. The percentage occurrence of cirrus on each night is obtained by taking the number of profiles in which cirrus clouds were observed to total number of profiles in that night.
Figure 6.3 shows the histogram of the percentage of occurrence of cirrus in lidar profiles during TCE-1999. Short hollow bars below the X-axis are used to denote the nights on which the lidar data were not available. Cross hatched bars indicate the percentage of occurrence for SVCs, light-shaded bars show the same for TCs and dark shaded ones for DCs. The frequency of occurrence of cirrus in January 1999 is significantly larger than those during the rest of the period. A marked difference was observed in the occurrence of cirrus clouds during the month of January and February. MST radar was operated in the vertical beam mode from 22:00 IST to 24:00 IST on all the nights. On four nights (18, 19, 23 and 29) in January 1999 the cirrus was of sub-visual type. Rest of the nights in January 1999 the clouds were either TC or DC type. The cloud is totally absent during the period 30 January 1999 to 2 February 1999. During the period 3 to 6 February 1999 the cirrus clouds reappeared in lidar signal even though their percentage of occurrences were low. Figure 6.3 shows that the clouds were strong and persistent for longer duration in January 1999 whereas in February-March period they were relatively weak and appeared only for a lesser duration.

Figure 6.3. Histogram showing the frequency of occurrence of cirrus during the period 18 January 1999 to 5 March 1999
6.4.1.2. Tropical Cirrus Experiment-2000 (TCE-2000)

During TCE-2000 on most of the nights the lidar observations were taken for ~3 hrs from 20:00 IST yielding ~45 profiles on each night. Figure 6.4 shows the histogram of percentage of occurrence of cirrus for different nights during this period. Cross hatched bars indicate the percentage occurrence for SVCs, light-shaded bars show the same for TCs and dark shaded ones for DCs. The hallow bars below the X-axis indicates the absence of lidar observations. During the period 25 February to 27 February 2000, lidar observation could not be taken due to the presence of thick low-level clouds. Unlike the case of TCE-1999, during the TCE-2000 period the cirrus clouds were present on most of the nights. During TCE-1999, February-March was a lean period as far as cirrus occurrence is concerned. But in TCE-2000, no such lean period is observed. Persistent cirrus is observed in February-March period. On a few nights, they were absent or intermittent in their occurrence. Only on 10 nights in March (3, 9, 10, 14, 15, 16, 19, 20, 23 and 25) the cirrus cloud was totally absent.

![Histogram showing the frequency of occurrence of cirrus during the period 21 February 2000 to 31 March 2000.](image-url)

Figure 6.4. Histogram showing the frequency of occurrence of cirrus during the period 21 February 2000 to 31 March 2000.
On seven nights they appeared for shorter duration. On most of the nights the cirrus observed were of SVC type with mean optical depth less than 0.03. On 4 March, 26-29 March and 31 March 2000 the observed cirrus were of TC type. DC type cloud is observed only on one night (7 March 2000) during this campaign period. On 13 nights cirrus persisted throughout the period of lidar observation. The average occurrence frequency of cirrus clouds in different months during TCE-1999 and TC-2000 does not show any marked difference even though the day-to-day occurrence pattern show some differences.

6.4.2. Cirrus Clouds and Altitude Structure of Turbulence Parameters

The MST radar is operated along with lidar on all the nights during the campaign period for ~2 hrs. The backscatter ratio profiles obtained from the lidar data and the turbulent parameters ε and K_m estimated from MST radar data are used to study the association of cirrus with tropospheric turbulence. Figure 6.5 shows the contour plots of R_p and R_s on 19 January 1999 along with the altitude profile of ε and K_m obtained from the MST radar data for that night (only for 22:00 to 24:00 IST). A thin cirrus cloud is seen in the contour plots around 16 km with a cloud depth of 0.8 km. It can be seen from Figure 6.5 that in this altitude region the profile of ε and K_m shows a prominent maximum. The peak value of ε exceeds 1\times10^3 m^2 s^{-3} and K_m exceeds 1.4 m^2 s^{-1} around 16 km, confined to a small altitude region of thickness less than a kilometre. At other altitudes (above and below the cirrus) the value of ε are significantly low.

Figure 6.5 Contour plots of backscatter ratio for co-polarized and cross-polarized components (R_p and R_s) and altitude profile of turbulent kinetic energy (ε) and eddy diffusion coefficient (K_m) on 19 January 1999

208
Figure 6.6 shows a similar plot for 20 January 1999 on which a thick cirrus cloud is observed in an extended altitude region of ~12 to 16 km. The contour plots of $R_p$ and $R_s$ are presented in this figure along with the mean profiles of $\varepsilon$ and $K_m$. The vertical extent of the cirrus is ~4 km. The peak in $\varepsilon$ and $K_m$ also is broad on this night covering the entire altitude region where cirrus is observed. The value of $K_m$ at the peak is greater (exceeds 5 m$^2$ s$^{-1}$) than that on 19 January 1999 (shown in Figure 6.5).

Figure 6.6 Contour plots of backscatter ratio for co-polarized and cross-polarized components ($R_p$ and $R_s$) and altitude profile of turbulent kinetic energy ($\varepsilon$) and eddy diffusion coefficient ($K_m$) on 20 January 1999

31 January 1999

Figure 6.7 Contour plots of backscatter ratio for co-polarized and cross-polarized components ($R_p$ and $R_s$) and altitude profile of turbulent kinetic energy ($\varepsilon$) and eddy diffusion coefficient ($K_m$) on 31 January 1999
Figure 6.7 shows contour plots of \( R_p \) and \( R_s \) along with the altitude profiles of \( \varepsilon \) and \( K_m \) on 31 January 1999 in which no cirrus clouds are observed in the lidar signal. Though altitude profiles of \( \varepsilon \) and \( K_m \) show a small peak around 13 km, it is not as prominent as that seen in Figures 6.5 and 6.6.

This study shows that the upper troposphere is generally turbulent (with high values of \( \varepsilon \) and \( K_m \)) and the strength of the turbulence increases significantly on nights in which cirrus clouds are observed. From the above illustrations it is seen that whenever the turbulence is strong and confined to small altitude region cirrus clouds are observed in that region and cirrus formation does not occur when the turbulence is relatively low.

The turbulent kinetic energy rates estimated in the present study is in agreement with earlier published works. The present study shows a range of values from \( 6 \times 10^{-4} \) to \( 8 \times 10^{-3} \) m\(^2\) s\(^{-3}\) for \( \varepsilon \) within the cirrus clouds and less than \( 6 \times 10^{-4} \) in the cloud free altitudes. The value of \( \varepsilon \) in the cloud free nights is always less than \( 6 \times 10^{-4} \) m\(^2\) s\(^{-3}\) (in the altitude region from 3.75 to 20 km). Gultepe and Starr [1995] reported that the \( \varepsilon \) ranges from \( 10^{-6} \) to \( 10^{-3} \) m\(^2\) s\(^{-3}\) within the cloud and the value is generally less than \( 0.5 \times 10^{-6} \) m\(^2\) s\(^{-3}\) in clear air (over Wisconsin and Minnesota in 1986 during FIRE I). Pinus and Litvinovo [1980] reported a much lower value of about \( 10^{-6} \) m\(^2\) s\(^{-3}\) for clear air in the free troposphere. Turbulence Kinetic energy rates estimated within the cirrus were of the order of \( 10^{-4} \) m\(^2\) s\(^{-3}\) [Chan et al., 1998] during SUCCESS (over Kansas). Dmitriyev et al. [1986] estimated \( \varepsilon \) of about \( 10^{-5} \) m\(^2\) s\(^{-3}\) for cirrostratus clouds. All these estimated values of \( \varepsilon \) within cirrus are from midlatitude locations. The study of upper tropospheric turbulence within cirrus over the tropics is rare.

6.4.2.1. Cirrus Clouds and Altitude Profiles of Turbulence Parameters on Different Nights During the Campaign Period

In the above, a few typical cases depicting the association of cirrus cloud with the altitude profiles of \( \varepsilon \) and \( K_m \) are presented and discussed. This aspect has been examined in detail by considering all the nights during the two-campaign period. Figure 6.8 shows the altitude profile of TKE dissipation rates for six typical nights in which cirrus clouds were present, which includes 3 nights in January 1999 (19, 20 and 29) when the cirrus persisted throughout the lidar observation period and 3 nights in February 1999 (3, 17 and 22) when the cirrus was intermittent, along with three nights of February 1999 (1, 12 and 13) in which the cirrus cloud was totally absent. The cold
The point tropopause level ($h_{tp}$) identified from the altitude profile of temperature derived from MST radar data is marked by an arrowhead on the left Y-axis. The cloud base ($h_b$) and cloud top ($h_c$) are marked on the right hand side Y-axis for those nights in which the clouds were observed in the lidar data.

Figure 6.8. Altitude profiles of TKE dissipation rates, $\varepsilon$ (m$^2$s$^{-3}$), on 19, 20 and 29 January 1999 when the cirrus was strong and persistent and on 3, 17 and 22 February 1999 when cirrus was weak and intermittent and on 1, 12 and 13 February 1999 when cirrus is totally absent. The arrowhead on left Y-axis indicates the tropopause level ($h_{tp}$). Mean cloud top ($h_m$) and cloud base ($h_{cb}$) are indicated by the two heads on the right Y-axis for those nights in which cirrus cloud was present.
As seen from Figure 6.8, on all the three nights in January $E$ shows well-defined peak in the altitude region where cirrus clouds were observed. On 29 January 1999, the peak in $E$ is at an altitude slightly below the cloud. For the two nights in February 1999 (February 3 and 22) when the cloud is relatively weak and intermittent in its appearance, $E$ shows a broad peak in the altitude region 10 to 18 km. These peaks are not sharply defined. On 17 February 1999 no cirrus is observed in the early part of the night up to 03: 45 IST (during MST radar observations) and after this the cirrus formation is observed ~16 km, which is just above the less prominent peak in $E$. For the remaining three nights in February (1, 12 and 13) in which the clouds are not observed, the altitude profile of $E$ is highly structured and the peaks are rather broad. These figures show that, in general, strong turbulence exists in the altitude region 10 to 18 km, which under favourable conditions can lead to cloud formation. The cloud could be strong and persistent at the altitude region where $E$ shows a sharp well-defined peak.

The turbulent parameters are also examined for different nights during the TCE-2000 campaign period. Figure 6.9 shows the altitude profile of TKE dissipation rate for six typical nights: on 23 February and 26 March 2000 when the cirrus was strong and persistent and on 8, 9, 11 and 14 March 2000 when cirrus was either weak/intermittent or totally absent. The arrowhead on left Y-axis indicates the tropopause level ($h_{tp}$). Mean cloud top ($h_{ct}$) and cloud base ($h_{cb}$) are indicated by the two heads on the right Y-axis for those nights in which cirrus clouds were present during the pre-midnight (MST radar observation period) period. On 23 February 2000, $E$ profile shows a pronounced peak in the altitude region 12-17 km up to the tropopause. On this night the cirrus was observed in the altitude region 11.5 to 15.3 km and persisted throughout the lidar observation period. On 26 March 2000, $E$ profile shows high values at lower altitude between 10 and 12.5 km. The cirrus was also observed at lower altitude but just above the sharp peak region of $E$. The altitude profiles of $E$ for the other four nights (8, 9, 11 and 14 March 2000) do not show any sharp well-defined peak.

The TKE dissipation rate, $E$, profiles observed on both the years 1999 and 2000 showed a well-defined peak whenever the cirrus is strong and persists for longer duration. If $E$ profiles are broad and weak the cirrus was observed to be weak and persists only for a shorter duration, if not totally absent. It may be noted that the altitude resolution of MST radar data was 150 m during TCE-1999 and it was only 300 m during TCE-2000.
Figure 6.9. Altitude profiles of TKE dissipation rates, $\varepsilon$ (m$^2$s$^{-3}$), on 23 February and 26 March 2000 when the cirrus was strong and persistent and on 8, 9, 11 and 14 March 2000 when cirrus was either weak/intermittent or totally absent. The arrowhead on left Y-axis indicates the tropopause level ($h_{tp}$). Mean cloud top ($h_{ct}$) and cloud base ($h_{cb}$) are indicated by the two heads on the right Y-axis for those nights in which cirrus was present during the pre-midnight period.

The frequency of occurrence of cirrus is found to be generally high during January 1999 compared to February 1999 (as seen from Figure 6.3). During February the occurrence of cirrus is weak and intermittent in nature. In this context the monthly mean profiles of TKE dissipation rate ($\varepsilon$) and eddy diffusion coefficient ($K_m$) are examined. Altitude profiles of monthly average TKE dissipation rate ($\varepsilon$) and eddy diffusion coefficient ($K_m$) for January and February 1999 are shown in Figures 6.10. These profiles indicate an enhancement in $\varepsilon$ and $K_m$ (and thus turbulence) with a broad maximum in the altitude region ~10 km to 17 km. In January 1999 (when cirrus occurrence is high) $K_m$ profile shows a broad maximum in the altitude region 11 km to 17 km. Superposed on this broad peak, there is a sharp peak at ~13 km in January
1999. In contrast in February 1999 when cirrus clouds are weak and intermittent in their occurrences, the $K_m$ profile is much broader extending from 7 km to 16 km with no sharp peak superposed. Further, $K_m$ is smaller in February than in January (in the peak region). The TKE dissipation rates also shows similar feature with high values during January compared February in 1999.

![Graphs showing average TKE dissipation rate and vertical eddy diffusion coefficient](image)

Figure 6.10. Altitude profiles of average TKE dissipation rate ($\varepsilon$) and vertical eddy diffusion coefficient ($K_m$) [(a) and (c) for January 1999, (b) and (d) for February 1999] during TCE-1999.

The frequency of occurrence of cirrus observed during TCE-2000 does not show any marked difference between February and March. It is seen from Figure 6.4 that cirrus is found to be present throughout the period of lidar observation on most of the nights and some nights they are weak and intermittent or totally absent. In this context, average $\varepsilon$ and $K_m$ profiles are obtained for the nights on which cirrus is strong and persists for longer duration and for the nights on which the cirrus is weak and...
intermittent or totally absent (irrespective of month). The average $\varepsilon$ and $K_m$ thus obtained are presented in Figure 6.11. As seen from the Figure 6.11 a and c the value of $\varepsilon$ and $K_m$ are generally large in the altitude region 9-18 km with a sharp peak in $K_m$ ~13.2 km on those nights in which the cirrus were strong and long persistent. On nights in which cirrus clouds were weak or totally absent the peak in $\varepsilon$ and $K_m$ profiles are rather broad and poorly defined (Figure 6.11 b and d). No sharp peak is observed in the altitude profile of $K_m$. This indicates that whenever the turbulence is high (enhanced $\varepsilon$ and $K_m$) in the upper troposphere, the observed cirrus is found have longer lifetime.

![Figure 6.11. Altitude profiles of average TKE dissipation rate ($\varepsilon$) and vertical eddy diffusion coefficient ($K_m$) [(a) and (c) for the nights when cirrus is strong and persist for longer duration, (b) and (d) for the nights with weak and intermittent cirrus or cloud free nights] during TCE-2000.](image)
6.4.2.2. Day to day Variability in Cirrus Activity and Turbulence in the Upper Troposphere

In the above (Section 6.4.2.1) it is seen that, the turbulence in the upper troposphere (12-18 km) is generally high over this tropical station. This is conducive for cirrus formation, under favourable conditions. When the amplitude of the peak in $e$ and $K_m$ is significantly large cirrus clouds are observed. But on some nights the cirrus clouds are strong and persistent and on some other nights they are weak and intermittent. This is associated with sharpness of the peak observed in the altitude profile of $e$ and $K_m$. These observations indicate that the day-to-day variability in cirrus activity can be associated with corresponding variations in turbulence parameters in the altitude region 14-16 km. In order to examine this aspect the contour plots of $e$ and $K_m$ are generated for the two campaigns superimposing the histograms showing the frequency of occurrence of cirrus clouds (presented in Figures 6.3 and 6.4 respectively). This is presented in Figure 6.12 for TCE-1999 and in Figure 6.13 for TCE-2000 along with a bar chart showing mean cloud strength for each night. From these figures it is seen that $e$ and $K_m$ are generally high in the altitude region 12-18 km. Above and below this altitude region $e$ and $K_m$ are very small.

On January 28 and 29, 1999 (Figure 6.12) a strong turbulent band is observed in the altitude region 10 to 14 km. In this period the observed cirrus is strong and persistent. In February 1999 the turbulence in this altitude region is rather weak with relatively low values of $e$ and $K_m$. During the period 22-26 January 1999 when the observed cirrus is optically dense the value of $K_m$ also is very high. After 29 January 1999 when the values of $K_m$ are generally low cloud becomes weak (as seen from bar chart of $M_o$) and intermittent (as seen from the frequency of occurrence). The above features are observable in Figure 6.13 also (for TCE-2000). During TCE-2000, number of high turbulent nights is larger than that in TCE-1999. They are distributed during the entire period of observation. But during TCE-1999 the observed values of $K_m$ during the former part of the campaign were larger than those during the latter. On comparing the cloud strength ($M_o$) with the strength of the turbulence, it is seen that $M_o$ is high during the strong turbulence in the upper troposphere. During the nights 24, 25 and 26 February 2000, a strong band of turbulence is observed in the altitude region 8 to 20 km. This could be due to strong convection from the lower altitudes extending up to tropopause. In this period lidar observations could not be taken due
Figure 6.12 Contour plots of $\varepsilon$ and $K_m$ showing the day-to-day variability of turbulent kinetic energy dissipation and vertical eddy diffusion coefficient in the altitude region 8 to 20 km during the period 18 January 1999 to 5 March 1999. The frequency of occurrence of cirrus is superposed on $K_m$ contour to illustrate the correspondence. The lower panel shows the mean cloud strength observed on the nights during the above period.
Figure 6.13 Contour plots of $\varepsilon$ and $K_m$ showing the day-to-day variability of turbulent kinetic energy dissipation and vertical eddy diffusion coefficient in the altitude region 8 to 20 km during the period 22 February 2000 to 31 March 2000. The frequency of occurrence of cirrus is superposed on $K_m$ contour to illustrate the correspondence. The lower panel shows the mean cloud strength observed on the nights during the above period.
the presence of thick low-level clouds (open bars below the X-axis in Figure 6.13b). This low level cloud would have been associated with dense cirrus above. During the nights of 1-2 March the turbulence is found to be weak and associated with this the cirrus cloud became weak and intermittent. From 3 March 2000 to 7 March 2000 turbulence in this altitude region became strong as seen from the contour of $K_m$; associated with that the cirrus occurrence has also increased. From 8 to 10 March 2000, the turbulence became weak once again. The percentage occurrence of cirrus is low during this period. Such waxing and waning of turbulence is observed from the contour plot of $K_m$ (in Figure 6.13) on the following nights, and the cirrus occurrence also varies accordingly. When the turbulence is strong, the probability of cirrus occurrence is high. Moreover, on some nights percentage of occurrence of cirrus is found to be very low when strong turbulence is observed (in $K_m$) which would have been due to limitation in the duration of lidar observation (~3hrs) (cirrus would have formed in the later part of night) or due to poor availability of condensable vapour.

The association between turbulence and cloud strength is examined by considering the altitude integrated turbulence parameter and the cloud strength for each night. The mean cloud strength is estimated during the 2 hrs period (when the

![Figure 6.14 Scatter plot of cloud strength against the turbulent kinetic energy dissipation rate](image)

Figure 6.14 Scatter plot of cloud strength against the turbulent kinetic energy dissipation rate.
MST radar was operated) by integrating and normalizing the effective backscatter ratio along the vertical extent of the cloud (Equation 3.34 and Equation 3.39) and averaging it for the cloud duration. The TKE dissipation rate (\( \epsilon \)) is integrated vertically along the cloud altitude width (from the cloud base to cloud top) and normalized to the cloud width to estimate the mean TKE dissipation rate. A scatter plot of cloud strength with TKE dissipation rate is presented in Figure 6.14. It shows a moderate positive correlation between cloud strength and TKE dissipation rate with a correlation coefficient of 0.46. The slope of the best-fit regression line is \( 6.7 \times 10^{-5} \) \( (m^2 s^{-3}) \). This shows that the cloud strength is directly related to the prevailing turbulence in the upper troposphere.

6.4.2.3. Vertical Convergence and Cirrus

The dominant sources of water vapour in the stratosphere are upward transport of humid air across tropical tropopause. This can occur over particular regions at tropics where tropopause temperatures are significantly low [Newell and Gould-Stewart, 1981]. This mostly occurs in the stratospheric fountain region of the western Pacific, northern Australia and India [Jensen et al., 2001a]. The observed cirrus clouds (whose occurrence frequency is high over India) provide a possible mechanism for dehydration associated with large-scale ascent across the tropopause. When an air parcel with initial peak relative humidity with respect to ice (RHI) of 30% at 7 km ascend to 17 km, its RHI increases above 200%, if no cloud formation occurs. As soon as cloud formation occurs, continued increase in RHI is prevented by depletion of vapour due to ice crystal deposition and growth. This leads to the formation of cirrus clouds and initiates the dehydration process. Jensen et al. [2001a] simulated one-dimensional model for studying the formation mechanism of cirrus taking into account the physical processes such as ice nucleation, ice crystal growth and sublimation and vertical transport. In the present study the effect of vertical transport on cirrus formation is examined based on observations.

Among the tropical cirrus experiment campaigns, TCE-1999 has two periods (January and February) in which there was a distinct contrast in cirrus occurrence. It would be worth in this context to examine the mean profile of \( K_m \) in these two periods (Figure 6.10). Again convergence in vertical wind can lead to accumulation of condensable vapours and nuclei to accumulate leading a favourable condition for cloud formation. To investigate this aspect the altitude gradient of \( K_m \) (say \( K_m' \)) is
examined. Figures 6.15a and 6.15b show the altitude profiles of $K_{m}'$ for January and February 1999. In both these months, in general, $K_{m}'$ is positive in the altitude region 9-13 km and above which it is strongly negative up to 15 km. In January 1999 (Fig. 6.15a) the $K_{m}'$ is high and shows sharp positive peak at ~13 km and a sharp negative peak at ~13.2 km within a small altitude region of 0.2 km. In contrast, during February 1999 the values of $K_{m}'$ at different altitudes are less and the amplitudes of positive/negative peak, which are rather broad in nature, is small. A similar pattern is observed during the year 2000 for the case shown in Figure 6.11. Positive and negative $K_{m}'$ would lead respectively to divergence and convergence of species like water vapour and aerosols in the presence of a negative altitude gradient in their concentrations. It appears that these conditions are favourable for accretion of constituents like aerosol and water vapour at altitudes very close to the tropopause leading to the formation of cirrus clouds. Jensen et al. [2001a] simulated the formation of ice clouds in slowly rising air and treated the advection of aerosol particles and water vapour assuming that any vertical divergence/convergence is balanced by horizontal convergence/divergence. When this convergence/divergence is strong the cloud is relatively strong and persists for a longer duration and when it is weak the chances of cloud formation are small, and even when it forms the cloud is very weak and intermittent in nature (the cloud gets dissipated faster).

![Figure 6.15 Altitude gradient of vertical eddy diffusion coefficient, $K_{m}'$, [(a) for January 1999, (b) for February 1999]](image-url)
Smith and Jonas [1996] during European Cloud and Radiation Experiment (EUREX) campaign (Scotland) observed the occurrence of turbulence in cirrus and found that they are patchy, weak and rather two-dimensional with most of turbulent kinetic energy being contained in the horizontal wind components [e.g., Dmitriyev et al., 1984; Quante and Brown, 1992]. The occurrence of intermittent turbulence in the cloud region can be attributed to strong wave interactions [Weinstock, 1985] or intermittence in the convective activity. Quante [1989] and Gultepe and Starr [1995] also confirmed the presence of gravity waves, quasi-two dimensional waves, and intermittent turbulence in the spatial and temporal structure of cirrus. The mean vertical velocity in the troposphere is about ±0.25 m s\(^{-1}\) and high magnitudes are observed very close to the cloud altitude region. Based on the detailed microphysical and dynamical measurements in cirrus clouds Heymsfield [1977] found that cirrus ice crystal habit, size distribution and number concentration depend strongly on the ambient vertical velocity and temperature. Satheesan and Krishna Murthy [2002] observed low values of Richardson number at 14-16 km at Gadanki and suggested that the low values of Richardson number could be due to dynamical instability driven by wind shear. Hence they concluded that the strong turbulence in the upper troposphere could be due to dynamic instability driven by vertical gradient of the horizontal wind. Fujiwara et al. [2003] observed significant turbulence in the altitude region 15-17 km at Sumatra Island (0.2°S, 100.32°E) using radar measurements. They attributed that the enhanced turbulence was convectively generated in the breaking phase of an equatorial Kelvin wave.

The present study shows that the region below tropopause (14-16 km) is highly turbulent and the eddy diffusion has a negative altitude gradient, which is favourable for accumulation of species (like water vapour and aerosols) having a negative altitude gradient for their concentrations. This is favourable for cirrus formation. This supports the formation of optically thin cirrus through in situ nucleation near tropopause. Supersaturation required for the in situ nucleation could be provided by the turbulence. Gu and Liou [2000] pointed out that the primary role of turbulence is to modulate the supersaturation conditions, which affects the rate of phase change between water vapour and ice crystals. The occurrence frequency of the cirrus is also found to be high in the altitude region 14-16 km as discussed in Chapter 4.
6.6. Summary

This study shows that the region below tropopause (14-16 km) is highly turbulent (with high values of TKE dissipation rates and Eddy diffusion coefficient). When the amplitude of the peak in $\epsilon$ and $K_m$ is significantly large cirrus clouds are observed. The persistent and intermittent nature of cirrus is associated with sharpness of the peak observed in the altitude profile of $\epsilon$ and $K_m$. The eddy diffusion has a positive altitude gradient followed by a strong negative altitude gradient. This is conducive for accumulation of trace species like aerosols and water vapour. This feature under favourable conditions leads to the formation of cirrus clouds. If the gradient in $K_m$ is large, cirrus clouds form and persist for a longer duration. When this gradient is small, the cirrus clouds are observed to be weak and intermittent in their appearance, if not totally absent. The formation and persistence of cirrus however, also depends on the available amount of water vapour also.