Lidar observations of aerosol layers just below the tropopause level during IFP-INDOEX

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A lidar system has been used at Gadanki (13.5°, 79.2°E) to study the characteristics of aerosol layer (cloud) occurring just below the tropical tropopause. The preliminary results of the lidar observations indicate that the cloud occurs ~ 2 km below the tropopause. The top and bottom edges of the cloud have propensity for ice crystal presence with liquid droplets/ vapours in-between. The clouds show temporal fluctuations (in their backscattering ratio) with temporal scales of the order of 30–90 min.

TROPICAL tropopause is a very cold region with temperatures as low as 190 K in some months of the year[1]. Such low temperatures are favourable for nucleation, condensation and formation of clouds provided vapours, e.g. H_2O/H_2SO_4 are present with sufficient pressure. In fact, thin persistent ice clouds (of H_2O/H_2SO_4) have been detected near the tropical tropopause and reported as early as 1977 from ground-based lidar observations[2]. From satellite-based measurements, Prabhakara *et al.*[3,4] found that the seasonal average cloud cover produced by thin cirrus (near tropopause) is as large as 50% over the tropical western Pacific in all seasons. They also reported that the thin cirrus occurrence frequency peaks sharply over the western Pacific particularly during northern winter when the tropopause is coldest. SAGE II solar occultation particle extinction measurements[5] also revealed optically thin cirrus clouds near tropopause in excess of 50% of the time in the western Pacific region and that they occur within top 2 km of the tropopause. Lidar measurements at Kwajelein[2] in the Pacific region showed the presence of these clouds persisting even for days. Air-borne lidar measurements during the Central Equatorial Pacific Experiment[6] (CEPEX) revealed the cloud layer with a thickness of the order of 200 m. These measurements showed that the cloud layer sometimes occurred not only in the outflow region of deep convective systems but are also present in regions apparently not affected by convection.

Optically the thin cirrus clouds are very thin and it is inferred that their optical depth[6] is less than 0.03. Gage *et al.*[7] suggested that the subvisible cirrus cloud may be responsible for local radiative energy absorption and hence, responsible for the observed upward vertical wind speeds of 0.3 cm/s near tropopause with implications to lower stratospheric circulation.

Jenson *et al.*[6] evaluated two mechanisms for the formation of the thin cirrus clouds: (i) Dissipation of optically thick cumulonimbus outflow anvils leaving behind an optically thin layer of ice crystals, (ii) *In situ* nucleation of ice crystals near the tropopause due to homogeneous freezing of sulphuric acid particles. The supersaturation required for the ice nucleation in the second mechanism may be due to either slow synoptic scale uplift or shear-driven turbulent mixing. Jenson *et al.* found that both the above two mechanisms are plausible for the formation of the thin cirrus clouds.

The microphysical properties of the clouds and their temporal variations are crucial information to understand and gain insight into the cloud formation mechanisms. For this ground-based lidar offers an excellent means though limited to nighttime operation. At Gadanki (13.5°N, 79.2°E), a tropical station, where a powerful MST radar has been in operation for the last few years, a lidar capable of operation in the Mie and Rayleigh scatter modes has been installed recently. The lidar is capable of providing signal strength measurements of the cross polarized component along with the parallel component in the Mie scatter mode, so that depolarization ratio which is a very important factor in the cloud microphysics can be obtained. During the Indian Ocean Experiment-Intense Field Phase (INDOEX-IFP) the lidar experiment at Gadanki has been conducted during nighttime from 18 January 1999 to 5 March 1999. We present here the preliminary results of an analysis of the lidar data on aerosol (cloud) layers just below the tropopause level. (We are using the more general term 'aerosol (cloud) layer' than the usual term 'cirrus cloud' as cirrus clouds usually comprise ice crystals whereas our observations reveal the presence of liquid phase also as shown later in the paper.)

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Lidar description and method of analysis

The lidar employs a Nd : YAG laser as the transmitter and it is operated at its harmonic wavelength of 532 nm. The laser pulse energy is 0.4 J and the pulse width is 7 ns. The laser beam is directed vertically upwards by an optical mirror system. The receiving telescope primary mirror is ~ 30 cm dia parabolic for Mie mode of operation (for aerosol studies in the troposphere and lower stratosphere). The transmitter beam and the receiving telescope are separated by a horizontal distance of ~ 2 m. The system is operated with an altitude resolution of 300 m and a pulse repetition frequency of 20 Hz. The system provides backscattered signal strength integrated over 5000 transmitted pulses for both polarizations (i.e.) parallel and perpendicular (to the laser beam polarization) polarization components. The integration over 5000 pulses corresponds to a time of 250 s. Three such integrated backscatter signal strength profiles are averaged providing a profile averaged over 750 s (12.5 min) and these profiles are the basic data used for subsequent analysis. The lidar has been operated on almost all the nights during 18 January 1999 to 5 March 1999 from ~ 2200 h to ~ 0500 h with few interruptions.

The backscatter signal strength profiles for the parallelly polarized (main) component are used to obtain the backscatter ratio. The backscatter ratio (BR) is given by

$$\mathrm{BR} = (b_\mathrm{R} + b_\mathrm{A})/b_\mathrm{R},$$

where b_R and b_A are the backscatter coefficients for air molecules (Rayleigh scatter) and aerosols respectively. The reference height for this evaluation is taken as 40 km where the aerosol number density can be assumed to be nil and the air density profiles for estimation of Rayleigh backscatter coefficient are taken from the model of Sasi and Sengupta[8]. The depolarization factor (D) which is essentially the ratio of the backscatter signal strength of the perpendicular to parallel component is also evaluated.

The lidar data obtained during 18 January 1999 to 5 March 1999 have been analysed to obtain the above parameters.

Results

We have examined the backscatter ratios derived from the lidar data for the presence of aerosol layers (clouds) below the tropopause level. The clouds will be characterized by high backscatter ratios. We have found that aerosol clouds are present just below the tropopause (which is taken to be at ~ 16 km) more frequently in January than in February and also in March till 5th during the period under examination. This probably is due to the generally colder tropopause (and upper troposphere) temperatures in January compared to February (and March). The aerosol clouds (hereafter referred as aerosol clouds than the more frequently used term cirrus clouds) are also stronger in January and last longer than in February when they occurred for short durations and were weaker.

Figure 1 shows the profile of backscatter ratio (BR) on the night of 20–21 January averaged over the period 2240 h to 0440 h. The aerosol cloud is present (Figure 1) between 12 and 16 km with two peaks in-between. In order to quantify the cloud characteristics we defined three parameters namely cloud strength, m_0; mean height of the cloud, m_1; and vertical thickness of the cloud, m_2. These are given by

$$m_0 = \sum \mathrm{BR}(h)$$

$$m_1 = \sum \mathrm{BR}(h).h/m_0$$

$$m_2 = \{\sum (h - m_1)^2 \mathrm{BR}(h)/m_0\},$$

where h_1 and h_2 are the lower and upper height limits of the cloud.

The above parameters are similar to the first three moments of frequency spectrum which give the total power, mean weighted frequency and the spectral width.

We have obtained the parameters m_0, m_1 and m_2 taking h_1 and h_2 as 12 and 18 km respectively (as the lower and upper height of the cloud). Based on the cloud strength m_0, the clouds can be broadly classified into strong ($m_0 > 20$) and weak ($m_0 < 20$) though the demarcation is not very distinct. Quite often it is found that the cloud strength (backscatter ratio) fluctuates rapidly with a time scale of 30 min. We present in the following features of four different cases of clouds which occurred on four different

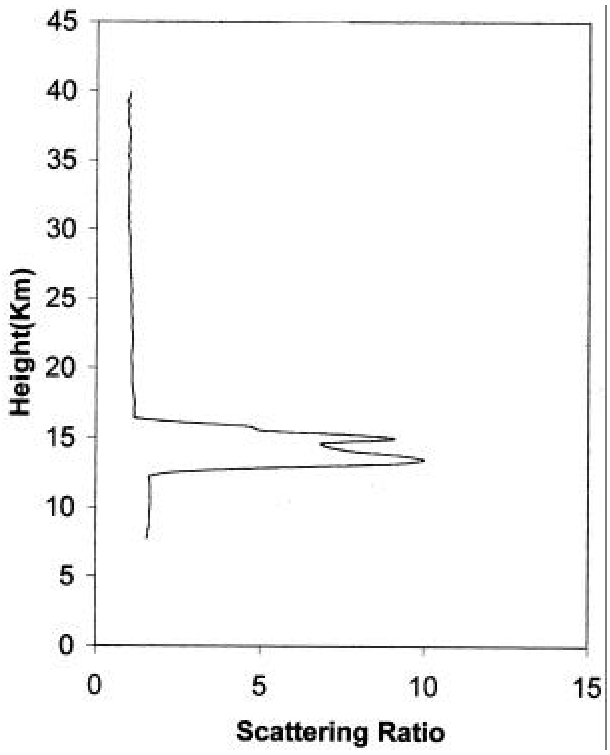

Figure 1. Backscatter ratio (parallel component, P) vs height on the night of 20–21 January 1999. This is the average over the period from 2240 h to 0440 h.

nights. These are typical of the clouds observed during the period under study.

18–19 January 1999

Figure 2 shows an example of a fluctuating cloud, which occurred on the night of 18–19 January. The figure shows m_0, m_1 and m_2. The cloud strength (m_0) shows rapid fluctuations with time with scales as short as ~ 30 min. The mean height m_1 (km) is around 14.9 km and appears to fluctuate similar but opposite in sense to m_0. The cloud vertical thickness varied between 1.45 km and 1.7 km and the cloud appears, in general, to be thinner when optically strong as seen from a comparison of the top and bottom panels of the figure.

Figure 3 shows the depolarization factor D of the cloud. It is known[8,9] that the depolarization D of moist air is less than that of dry air and that it is more for solid crystals (H_2O/H_2SO_4) of non-spherical shape (compared to dry air). Examining Figure 3 we find that D in the cloud is more than that in the ambient (above and below the cloud) till about midnight. Later it is high in the top and bottom edges of the cloud with lower values in-between. So, it appears that the top and bottom edges of the cloud have strong ice crystal (of presumably H_2O/H_2SO_4) presence while within the cloud liquid phase dominates. Further, D appears to be more, thus indicating stronger ice crystal presence, in the top edge than in the bottom edge. D does not show the rapid fluctuations of m_0.

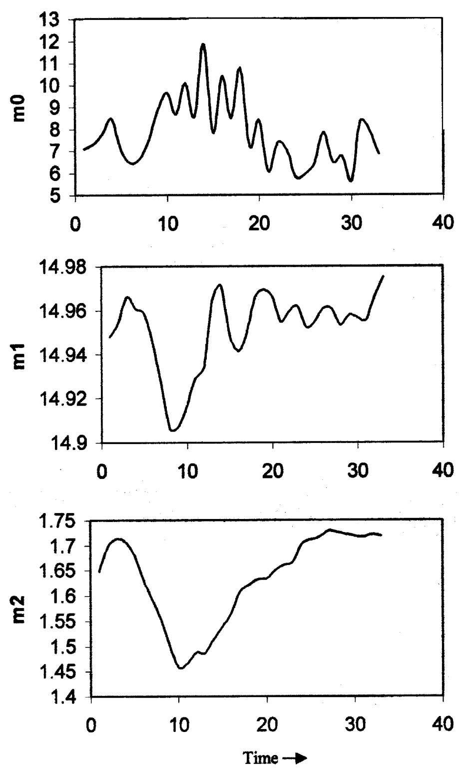

Figure 2. Cloud parameters m_0 (top panel), m_1 (middle panel) and m_2 (bottom panel) on the night of 18–19 January 1999 for the main (parallel) component. The starting time (corresponding to 0 on the time axis) is 2240 h. One unit on the *x*-axis corresponds to 750 s (12.5 min).

20–21 January 1999

Figure 4 shows features of a strong cloud which occurred on the night of 20–21 January. The value of m_0 (top panel) is between 20 and 50 with superposed fluctuations till ~ 0400 h (20 on the *x*-axis scale in the figure). Then it raises sharply to a very high value > 100 followed by a sharp decrease. The total duration of this excursion is ~ 1 h. The mean height of the cloud m_1 (middle panel) is between 14.3 km and 15 km till 0340 h in the night. Later, when the cloud strength rose sharply the mean height decreased rapidly to 13.8 km. The cloud thickness (m_2)

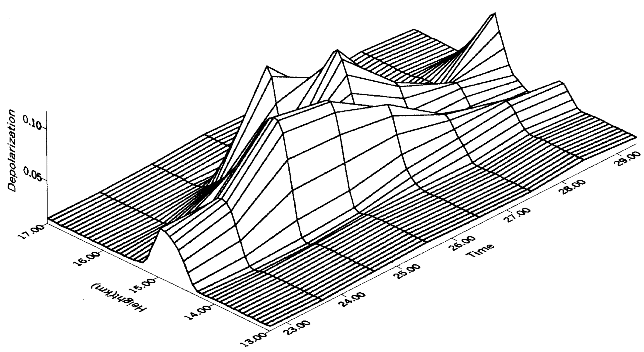

Figure 3. Depolarization ratio on the night of 18–19 January 1999.

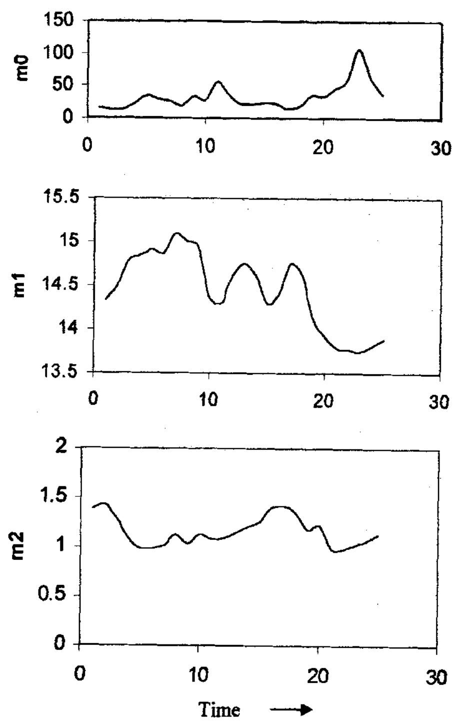

Figure 4. Same as Figure 2, but for the night of 20–21 January 1999 and starting time 2352 h.

varied between 1 to 1.4 km. It appears that the cloud is thinner when strong.

Figure 5 shows the depolarization factor *D*. It is higher in the top and bottom edges of the cloud with a minimum in-between. The minimum is not deep. An interesting feature is the development of a region of high *D* at lower heights (12.5–13.0 km) in the early morning hours. There is again a region of low *D* (rather broad) between this cloud and the bottom edge of the main cloud.

26–27 January 1999

Figure 6 shows the three parameters on this night. The cloud on this night is not as strong as in the previous case but moderately strong. The cloud becomes stronger in the early morning hours as revealed by m_0. It fluctuated between 20 and 40 till early morning hours. The mean height of the cloud remained around 14.4 km for most of the night. In the early morning hours when the cloud became stronger there is a sudden decrease in its mean height to 13.7 km. The cloud thickness remained around 1 km except for some fluctuations.

Depolarization factor *D* is shown in Figure 7. *D* is quite low within the cloud compared to the ambient. This means the cloud is mainly in the vapour/liquid droplet state. In the early morning a region (broad) of high *D* developed below the bottom edge of the cloud as in the previous case. So, in the early morning hours there is a development of ice crystal layer below the bottom edge of the main cloud. This is also the period of high m_0.

3–4 February 1999

Figures 7 and 8 show the three parameters and *D* respectively. The cloud on this night is quite different from the earlier ones. It shows two bursts of higher m_0 values lasting for 30–90 min before midnight. Except for these bursts the cloud is very weak throughout the night with minor fluctuations. The mean height is steady around 1.5 km except for sharp troughs corresponding to the peaks in m_0. The cloud thickness also is high on this night (~ 1.75 km) except that it is thinner during the bursts.

The depolarization *D* is low within the cloud throughout the night compared to the ambient. This indicates that the cloud is in vapour/liquid droplet state.

Discussion and conclusions

The aerosol clouds appeared approximately 2 km below the tropical tropopause. The fact that they appeared more strongly in January than in February could be linked to the lower tropopause temperatures (and also upper troposphere) in January compared to in February. Low temperature is one of the conditions favourable for the presence of these clouds.

The strong clouds showed high depolarization factors in the top and bottom edges of the cloud with low values within the cloud. This indicates that the cloud edges have propensity for solid ice crystal formation. When the cloud

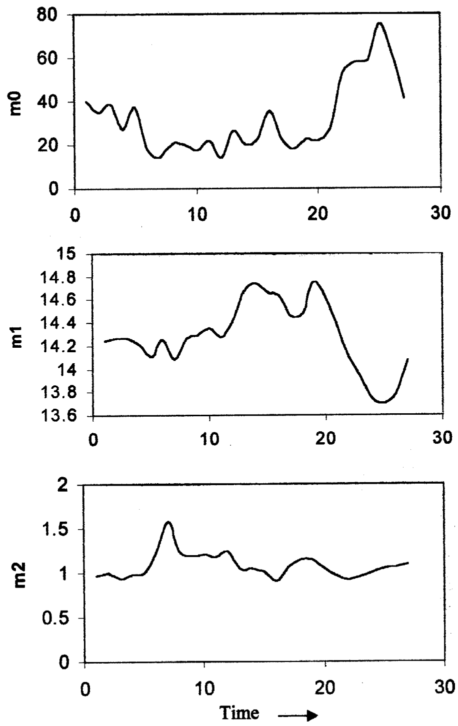

Figure 6. Same as Figure 2, but for night of 26–27 January 1999 and starting time 2205 h.

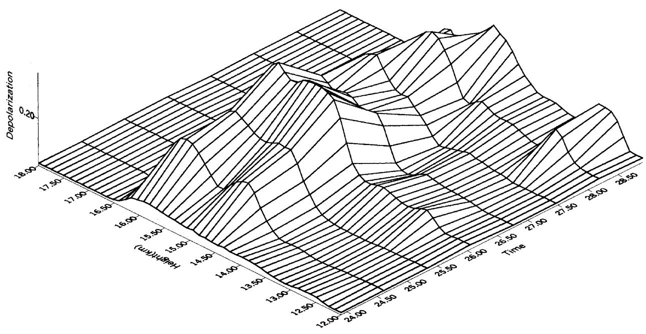

Figure 5. Same as Figure 3, but for night of 20–21 January 1999.

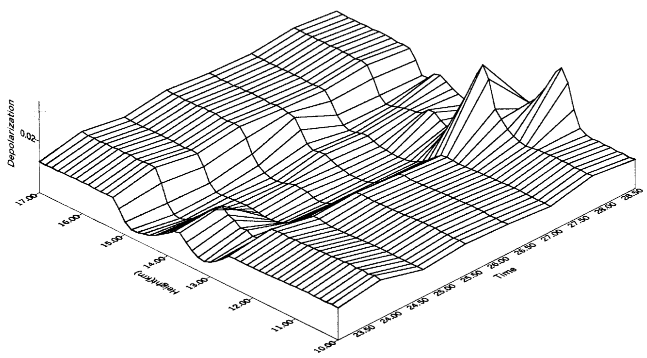

Figure 7. Same as Figure 3, but for night of 26–27 January 1999.

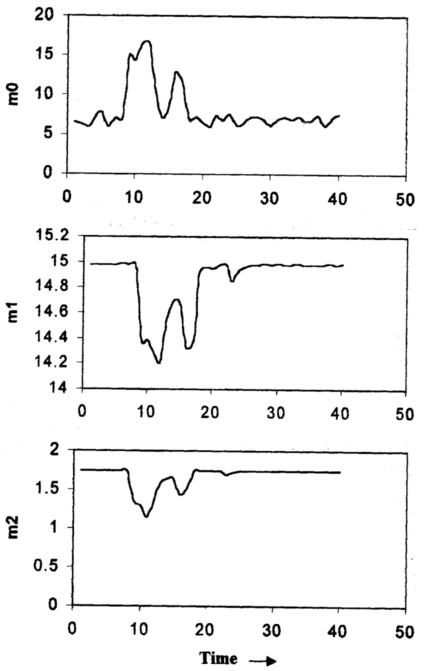

Figure 8. Same as Figure 2, but for night of 3–4 February 1999 and starting time 2206 h.

is wear/moderately strong, it is basically in the vapour/ liquid droplet state.

Jenson *et al.* suggested that dissipation of optically thick cumulonimbus outflow anvils leaving behind ice crystal layer and *in situ* nucleation of ice crystals near the tropopause due to freezing of sulphuric acid particles are two plausible mechanisms for cirrus cloud formation near the tropopause. In the case of cloud formation due to the first process, which is a large-scale process, the cloud dissipation would be due to adiabatic process and the clouds are likely to persist for days. (Large wind shears can also dissipate through diffusion.) We have found that the clouds often show very deep fluctuations with time scales of the order of 30 to 90 min. Thus, it appears that the source of the clouds is of variable strength (and/or the dissipation process is quite rapid). In the second process of cloud formation, turbulence plays a major role in transporting the vapours from lower heights to near tropopause heights where *in situ* nucleation can occur. It is quite likely that the turbulence strength itself is quite variable in time, thus modulating the cloud strength. Internal cloud dynamics/microphysics (including turbulence within the cloud) also can cause the fluctuations in the cloud strength. Thin layers of turbulence interspersed with layers of stability may occur below the tropopause. The thin stable layers where the mixing will be weak could facilitate formation of ice crystal layers due to high density of cloud condensation nuclei/aerosols. In between these stable layers of ice crystals, there can be condensed vapour in liquid state (or high vapour density without condensation) as observed in the present study when the clouds are strong. Further work is in progress on these lines.

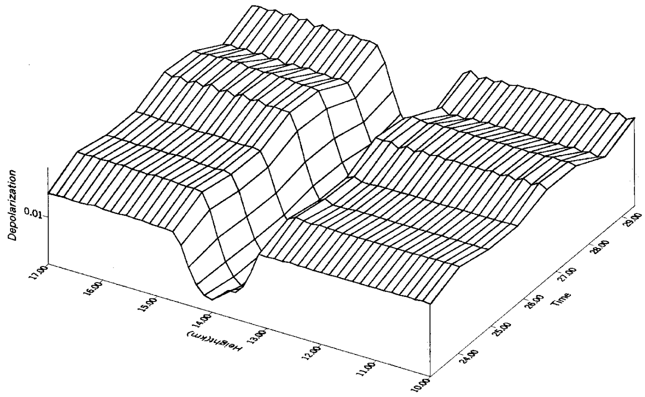

Figure 9. Same as Figure 3, but for night of 3–4 February 1999.

1. The following are the main conclusions from the present study.
2. Thin aerosol clouds occur ~ 2 km below the tropical tropopause. They occur more frequently and strongly when the tropopause (and upper tropopause) is colder.
3. As the cloud becomes stronger it comes down in height and becomes thinner, may be due to cloud mass content.
4. The top and bottom edges of the cloud have propensity for presence of ice crystals with liquid phase in between.

The strength of the cloud shows very prominent fluctuations with time scales of the order of 30 to 60 min.

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