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Atmospheric Chemistry and Physics Discussions

Suppression of warm rain by aerosols in rain-shadow areas of India

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Abstract

Aircraft observations of clouds and aerosols were conducted during the Cloud Aerosol Interaction and Precipitation Enhancement Experiment (CAIPEEX) executed by the Indian Institute of Tropical Meteorology over the Indian subcontinent during the period of May–September 2009. Existence of aerosol layer with large concentrations of cloud drop condensation nuclei extended up to heights of 4 to 5 km was observed over the rain shadow areas to the east of the Western Ghats over central India. The thick aerosol layers were observed to suppress the formation of warm rain in convective clouds up to heights of about 7 km, where mixed phase precipitation formed. This prevented clouds that did not exceed this height from precipitating significantly. This might invigorate the very deep clouds on expense of the smaller clouds. The aerosol radiative effects are suspected to decrease the surface heating and hence the available energy for propelling the convection. The net effect of the aerosols on the rainfall amounts is unknown due to the complexity of the effect, but it is suspected to be detrimental in an area where the rainfall is critical to the livelihood of the inhabitants. This requires continuation of this research.

1 Introduction

Heavy aerosol loading in the atmosphere can have both hydrological and radiative cloud-mediated effects on the climate. Aerosols can block the incoming solar radiation thereby inhibiting the convection process and can slow down the monsoon circulation (Satheesh and Ramanathan, 2000; Ramanathan et al., 2001; Chung and Ramanathan, 2004). Large concentrations of cloud drop condensation nuclei (CCN) aerosols nucleate large number of small cloud drops. The smaller drops are slower to form precipitation from shallow and short living clouds (Gunn and Phillips, 1957; Twomey, 1977; Albrecht, 1989; Rosenfeld, 2000; Ramanathan et al., 2001; Rosenfeld et al., 2001; Rosenfeld and Givati, 2006; Qian et al., 2009). The delay of rain in warm based deep

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convective clouds can lead to their invigoration and eventual heavier rainfall (Andreae et al., 2004; Rosenfeld et al., 2008). Areas within the downwind leeward side of the Western Ghats along the west coast of central India do not get enough rainfall unless they are under the synoptic forcing of depressions within the monsoon circulation. Given the opposite effects under different conditions, the role that aerosols could play in the rainfall deficiency over these regions remains unclear. The water shortage leads to demand for cloud seeding over these areas, despite being still unproven for adding water in India. Figure 1 shows part of the rain shadow regions of India in which the observations for this study were collected. To understand the aerosol and cloud microphysical properties over these areas, aircraft observations were conducted from May to September 2009 under the Cloud Aerosol Interaction and Precipitation Enhancement Experiment (CAIPEEX), which was executed by the Indian Institute of Tropical Meteorology (http://www.tropmet.res.in/~caipeex/).

2 The aircraft measurements

A Piper Cheyenne instrumented twin engine turboprop aircraft was utilized during the CAIPEEX program. This pressurized aircraft can climb up to a maximum altitude of 8 km. During the observations the mean true air speed of the aircraft was approximately 100 m/s. Most of the observations over the study areas were carried out during weak or break monsoon conditions. We utilized the aerosol and cloud condensation nuclei (CCN) concentrations (cm^{-3}) at 0.35–0.45% super saturation (SS) outside the cloud and lapse rate of environmental temperature (dT/dH, $^{\circ}C\,km^{-1}$) to explore: how aerosols influence (1) stability of the atmosphere and (2) cloud microphysical properties. The aerosol concentration was measured by Passive Cavity Aerosol Spectrometer Probe (PCASP) of diameter ranging from 0.1 to 3 µm in 30 bins. The CCN concentrations were measured by CCN counter, manufactured by Droplet Measurement Technology (DMT). It was set at SS cycle of 0.2%, 0.4% and 0.6%, which takes approximately 2 min to step to a new SS. The CCN counter operates on the principle that diffusion

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of heat in air is slower than diffusion for water vapor (Roberts and Nenes, 2005). Activated droplet counting is done with an Optical Particle Counter. The Cloud Droplet Probe (CDP), manufactured by DMT, measured cloud droplet concentrations of diameter range from 2.5 to 50 µm in 30 range bins. The hydrometeors images were obtained by the DMT cloud image probe (CIP). Satellite retrieved Aerosol Optical Depth (AOD) was obtained from The Moderate Resolution Imaging Spectroradiometer (MODIS) and Aerosol Index (AI) was obtained from total Ozone Mapping Spectrometer (TOMS). These aerosol parameters were utilized to estimate the aerosol loading and type of aerosols in the atmosphere. The MODIS provides AOD of coarse and fine particles. Positive AI indicates absorbing whereas negative AI indicates non absorbing type of aerosols.

3 Aerosol radiative effects

Total eight cases of aircraft measurements of cloud and aerosols vertical profiles over these areas are considered here. The temperature (°C), relative humidity (RH, %), aerosol and CCN concentration at 0.4% SS profiles of three contrasting days, two with heavy aerosols (21 June and 24 September) and other with relatively less aerosols (22 June) are shown in (Fig. 2a–f). The temperature profiles are characterized by lapse rates ranging from –4.9 to 1.5 °C/km at different altitudes. The rain shadow areas were under the influence of heavy concentrations of aerosol particles extended up to 5 km for the polluted cases. Significantly, the stable layers of several hundred meters of depths were accompanied by heavy aerosol and CCN concentrations. Aerosol concentrations greater 500 cm^{-3} and CCN concentrations at 0.35 to 0.45% SS in excess of 1000 cm^{-3} were observed up to heights greater than 4 km. Relatively high AOD of the range 0.40 to 0.66 obtained from MODIS over these areas agree with the presence of heavy aerosol concentrations (see Table 1). However, the association of stable layers with aerosols is not necessarily the result of aerosols radiative warming, because stable layers occur frequently regardless of the aerosols. The AI obtained from TOMS was

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in the range from 0.64 to 1.67. The positive AI suggested that the aerosols were of absorbing type. For the relatively clean environment at Raichur there were no stable layers (Fig. 2c, d). The aerosol and CCN concentrations were comparatively low both at the cloud base and at higher altitudes. The AOD was only 0.12 and AI was 1.12 which revealed that the aerosol loading was relatively low and the aerosols were of absorbing type.

The aerosols reflect some of the solar radiation back to space and so cause surface cooling (Satheesh and Ramanathan, 2000; Padma Kumari et al., 2007). In addition, some of the solar radiation is absorbed in the aerosols and warms the lower troposphere on expense of the surface temperature, thus causing low level stabilization (Kaufman et al., 2006) along with some destabilization above the aerosol layer. This inhibits the convection, especially when AOD exceeds about 0.25 (Rosenfeld et al., 2008; Koren et al., 2008), which was the prevailing situation in this study. We found a weak positive correlation between increasing AOD and the convective available potential energy, CAPE (Fig. 3). As these observations were during weak or break conditions of monsoon, the CAPE was not strongly influenced by the large scale monsoon circulation. Vertical growth of convective clouds is slowed or completely stopped by the stable layers. The detraining of the moisture at these heights might give rise to the increase in RH in the stable layers (e.g. Fig. 2). The moisture available in the atmosphere and increase in RH at the stable layer can produce haze particles. The stable layers could also delineate dry and wet air above and below the stable layers in the atmosphere by not allowing vertical mixing. Such atmosphere with moderate to high cape, but with convective inhibition (CIN), are conducive to thunderclouds when the CIN is broken (Colby, 1984; Lutgens and Tarbuck, 1986).

4 Aerosol microphysical effects

We present here the cloud microphysical properties of the convective clouds developed in the heavily aerosol loaded atmosphere. The convective clouds were profiled from

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the cloud base to the cloud top or vice versa. The cloud drop size distributions (DSDs) of growing convective clouds at different altitudes are presented in (Fig. 4a–e). The measurements took place in growing convective elements where precipitation could not have fallen from above, which was typically near the cloud tops. The cloud DSDs of the rain shadow areas were narrow at all altitudes, while in relatively clean environment cloud DSDs broadened faster with height. The narrowness and deficiency of large cloud droplets in the DSD spectra near cloud base revealed that giant CCN did not affect significantly the DSD. The scatter plot between the mean the maximum CDP concentration (cm^{-3}) above, and CCN concentrations (cm^{-3}) at 0.4% SS below the cloud base is shown in Fig. 5. The CCN at 0.4% SS was found out from the CCN-SS relation below the cloud base. Increase in CCN concentrations results increase in CDP concentrations of cloud drops. The narrow spectrum of the cloud droplets and large amount of CCN particles at the cloud base suggested that the clouds were super continental in nature. In case of relatively less polluted environment with small CCN concentrations, the DSD widened faster with height and formed large cloud droplets favorable for collision-coalescence process (e.g. Fig. 4b). The onset of warm rain processes could be inferred from the effective radius (R_e) of cloud droplets. The effective radius (R_e) is the ratio of sum of volumes (3rd moment of DSD) to the sum of surface areas (2nd moment of DSD) of the droplets in the measured cloud volume. It was observed elsewhere that when water vapor condenses on the cloud droplets and R_e reaches $\sim 12\,\mu m$, the cloud droplets start to form rain by collision and coalescence processes (Andrae et al., 2004; Rosenfeld et al., 2008b). This is also the case in these clouds, but a full treatment of this issue for CAIPEEX is done in another study.

Figure 6 presents the profile of R_e with respect to T (°C) for the 8 cases of convective clouds. We present here the R_e of the mean DSDs of the horizontal cloud penetrations with corresponding T. The R_e at the cloud base must start at the smallest size by which cloud drops are nucleated, but diverges with height according to the CCN and hence to cloud drop number concentrations. It is interesting to note that over Nashik, to reach R_e of 12 µm the cloud had to grow till 6.5 km (–6 °C). Over Nanded, the threshold

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$R_e \sim 12\,\mu m$ for rain initiation was not achieved even at ~ 7.5 km which correspond to the temperature of $-15\,^{\circ}C$. The smallness of the cloud droplets inhibits formation larger size cloud droplets as the coalescence process becomes inefficient. This in turn inhibited the formation of significant warm rain by drop coalescence. In such clouds most precipitation forms as ice particles. Images of such hydrometeors are shown in Fig. 7 for the convective cloud over Nanded (24 September). The images obtained from the cloud image probe shows presence of ice and graupel hydrometeors at temperature $-11\,^{\circ}C$ at 7.2 km. It was mentioned earlier that the haze layers over these regions were extended to colder temperature. The aerosol particles could act also as ice nuclei and initiate ice-phase precipitation at the higher altitudes (Sassen, 2002; Uno et al., 2009). The small cloud droplets in such low temperatures rimed to form graupel ice hydrometeors. Such convective clouds produce much lightning when becoming deep. Over central India frequent lightning activities are reported (Lal and Pawar, 2009). The direct transformation of cloud droplets to ice hydrometeors could indicate a plausible reason why lightning activities occur over these regions.

In relatively clean environment and with the absence of strong stable layers, warm rain form early as revealed from the R_e-T over Raichur and shown in Fig. 6. The threshold R_e of $12\,\mu m$ was reached at much warmer temperature of $4\,^{\circ}C$. However, in this case drizzle actually started forming only when Re reached $14\,\mu m$ at height of 5.8 km and temperature of $-7\,^{\circ}C$ (see Fig. 8). This means that even the relatively less polluted clouds warm rain was inhibited substantially, although to a lesser extent than in the more polluted cases.

These cases show the important role of large concentrations of CCN aerosols in suppressing the rain forming processes in the convective clouds of the rain-shadow areas, even when becoming quite deep with tops exceeding the $-10\,^{\circ}C$ isotherm level. This means that there is a net suppression of rainfall from polluted clouds with relatively high and cool bases ($T <\sim 15\,^{\circ}C$) and with tops that do not exceed by much the height of about 6 to 7 km. The invigoration effect might dominate the very deep cloud, especially when cloud bases are warmer (Rosenfeld et al., 2008a). The radiative suppressive

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effect might still dominate this invigoration effect for polluted conditions with AOD > 0.3 (Rosenfeld et al., 2008a; Koren et al., 2008). This AOD threshold is often exceeded in the rain shadow area, especially during the monsoon break conditions.

5 Summary and conclusions

Pollution aerosols appear to modify precipitation in the rain shadow areas of India in several ways, often to the detriment. Thick layers of heavy concentrations of aerosols were found and are suspected to affect the cloud formation, especially when the monsoon circulation is weak and rainfall is needed most. The aerosol layers can directly influence the instability of the atmosphere by partially absorbing and reflecting the solar radiation. This cools and reduces surface evaporation and stabilizes the lower troposphere. This slows at least the initiation of convective clouds. The large concentration of CCN aerosols slows the formation of warm rain and suppresses precipitation from clouds that are not very deep, i.e., with tops less than at least 7 km. The prevention of rain from the smaller clouds and from the lower portion of the deep clouds likely incurs invigoration of the deep (i.e., >10 km) clouds. This means that the pollution aerosols redistribute the rainfall such that less rainfall occurs in areas where rain amounts are modest anyway, and even more rainfall occurs where it already rains heavily. This can potentially explain the findings of Goswami et al. (2006), showing (i) significant rising trends in the frequency and the magnitude of extreme rain events and (ii) a significant decreasing trend in the frequency of moderate events over central India during the monsoon seasons from 1951 to 2000.

Theoretical considerations (Rosenfeld et al., 2008a) and observations elsewhere (Koren et al., 2008) suggest that under heavy aerosol loading (AOD > 0.3), which are typical mainly for the monsoon break at the rain shadow areas, the radiative suppressive effects dominate over the microphysical invigoration effects.

This study provides us just a first look at the possible effects of the pollution aerosols on precipitation in the rain shadow area in India, where severe water shortage occurs.

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Much more research needs to be done for a full understanding of the complex ways by which aerosols affects clouds and precipitation under the various meteorological conditions there and elsewhere.

Acknowledgement. CAIPEEX is funded by Ministry of Earth Science, Govt. of India. We sincerely acknowledge the effort of CAIPEEX team members for successfully conducting the experiment. Thanks to the pilot Captain Mark Hampton for safely maneuvering the aircraft into the convective clouds. We also acknowledge the MODIS, TOMS mission scientists and associated NASA personnel for the production of the data used in this research effort.

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Table 1. Details of flight sorties, Aerosol and Cloud microphysical properties as measured by instrumented aircraft. Aerosol Optical Depth (AOD) from MODIS, Aerosol Index (AI) from TOMS and Convective Available Potential Energy (CAPE) obtained from radiosonde observations.

Date	Time of flight (UTC) Start	End	CAPE J/kg	AOD	AI	CDP cm^{-3}	CCN at 0.4% SS cm^{-3}	$CCN = aSS^b$ a	b
12/06/2009	07:04	09:55	1487	0.81	0.67	2062	3405	10 705	1.25
15/06/2009	08:12	10:10	221	0.40	1.28	948	1614	2849	0.62
16/06/2009	08:19	10:57	325	0.43	1.06	1379	2377	7405	1.24
21/06/2009	07:58	10:46	1523	0.64	0.81	1046	1669	4784	1.15
22/06/2009	07:53	10:46	439	0.15	1.12	388	657	1939	1.18
06/07/2009	07:46	10:44	1961	0.66	1.17	543	596	1856	1.24
16/09/2009	09:05	11:52	1070	0.47	0.75	1088	2989	18 681	2.00
23/09/2009	09:49	12:53	3138	0.52	0.64	1005	3858	9309	0.96
24/09/2009	09:21	12:40	2684	0.31	1.67	1076	2772	8884	1.27

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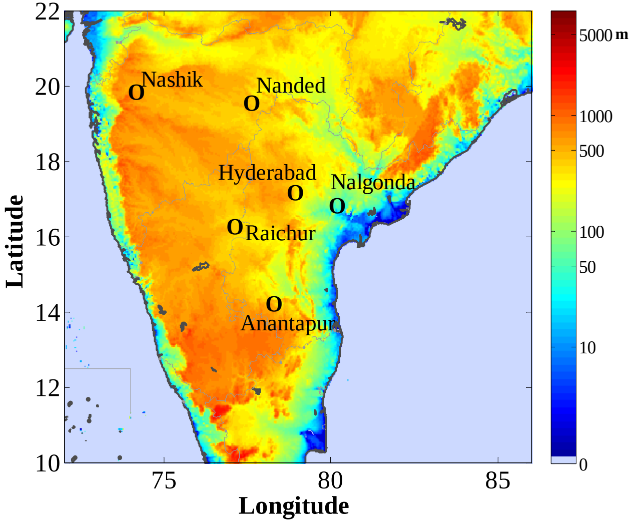

Fig. 1. Part of the rain shadow areas where aircraft observations of cloud and aerosols were conducted during CAIPEEX program 2009.

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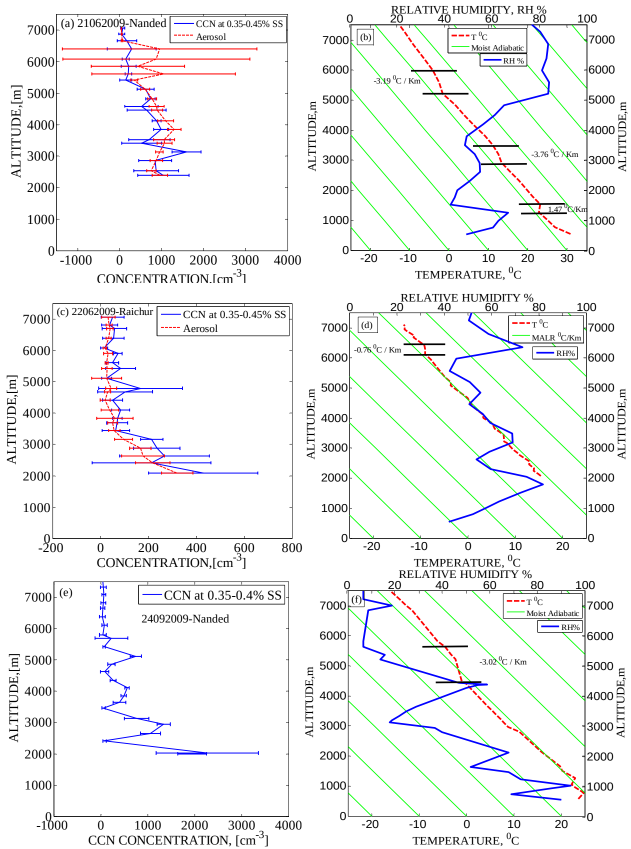

Fig. 2. Mean Aerosol and CCN concentration at 0.35 to 0.4% SS and Temperature (°C) profile as obtained from the aircraft over **(a, b)** Nanded 21 June, **(c, d)** Raichur 22 June and **(e, f)** Nanded 24 September The mean andstandard deviations at a resolution of 250 m are indicated by the horizontallines. The lapse rates of T at thestable layers are indicated in the figures. The CCN and aerosol concentrationsare considered outside the cloud.

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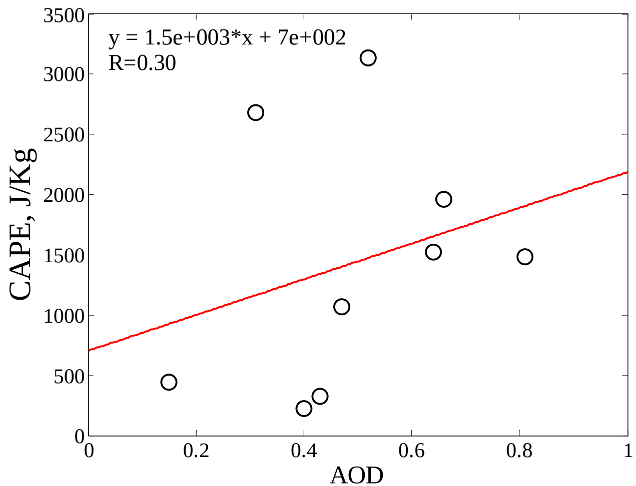

Fig. 3. Scatter plot between CAPE and AOD, weak correlation coefficient of 0.30 found between them.

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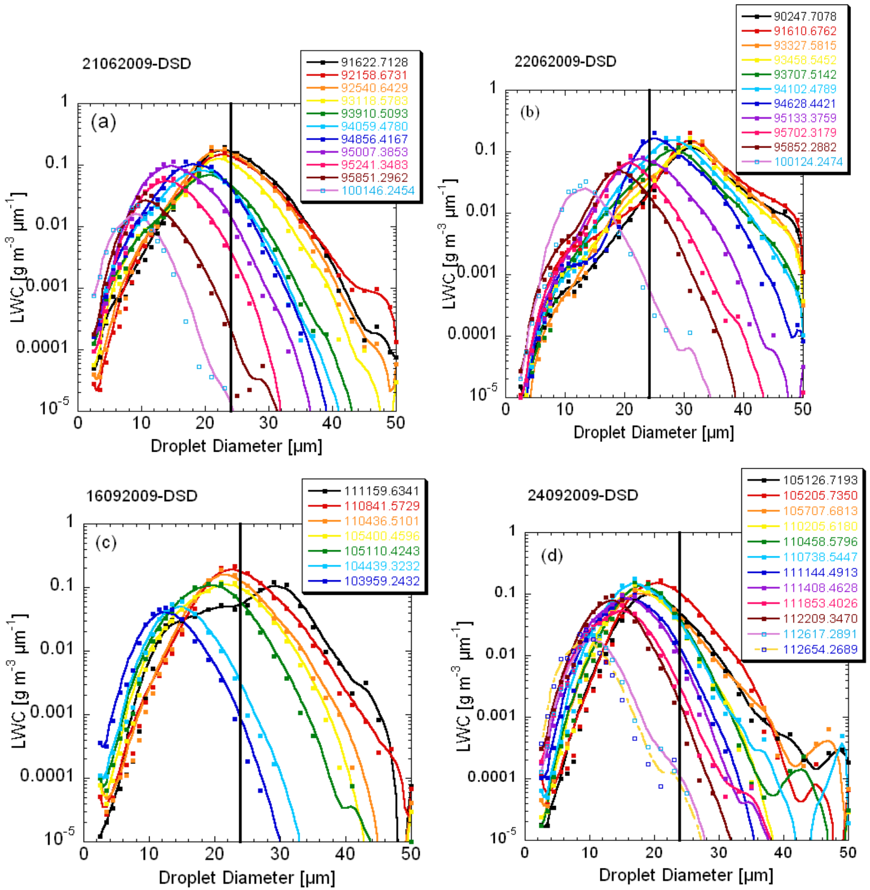

Fig. 4. Evolution of cloud droplet size distributions over **(a)** Nanded, **(b)** Raichur, **(c)** Nassik and **(d)** Nanded. The threshold droplet diameter ~24 µm for warm rain process is marked by the vertical solid line. The legend of each line is the GMT time and mean penetration height in m above sea level.

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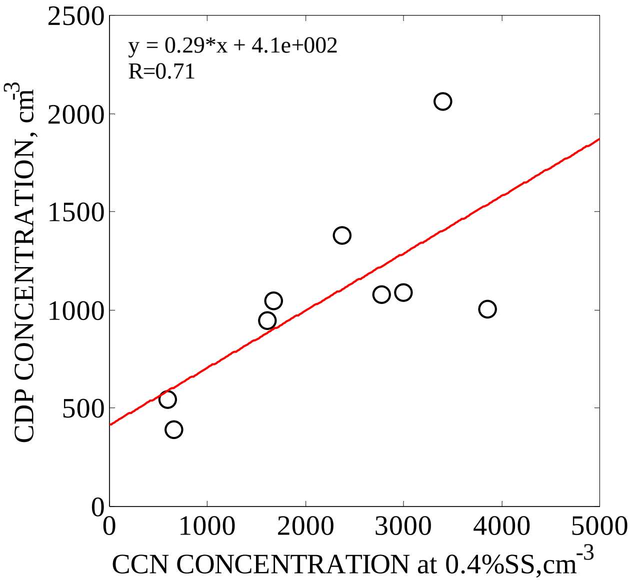

Fig. 5. Scatter plot between cloud droplet concentration and cloud condensation nuclei concentration at 0.4% of super saturation. The CCN concentration at 0.4% SS is calculated by using CCN-SS relation at the cloud base (see Table 1). The cloud droplet concentration is the mean of few maximum cloud droplet concentrations above the cloud base. The number of cloud droplets increase with increase in CCN concentrations.

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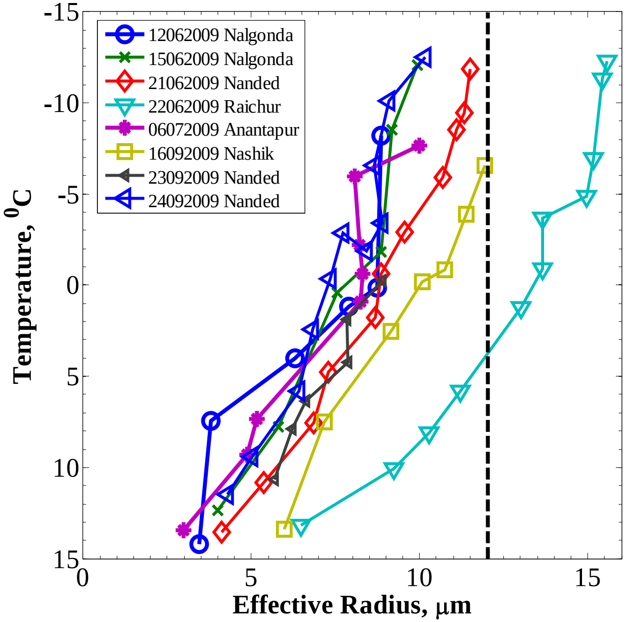

Fig. 6. The Temperature-Effective radius (R_e-T) plot over some of the rain shadow areas in India. The measurements were made from the cloud base to cloud top of the growing convective clouds. The observed R_e were less than critical radius of 12 μm for initiation of warm rain.

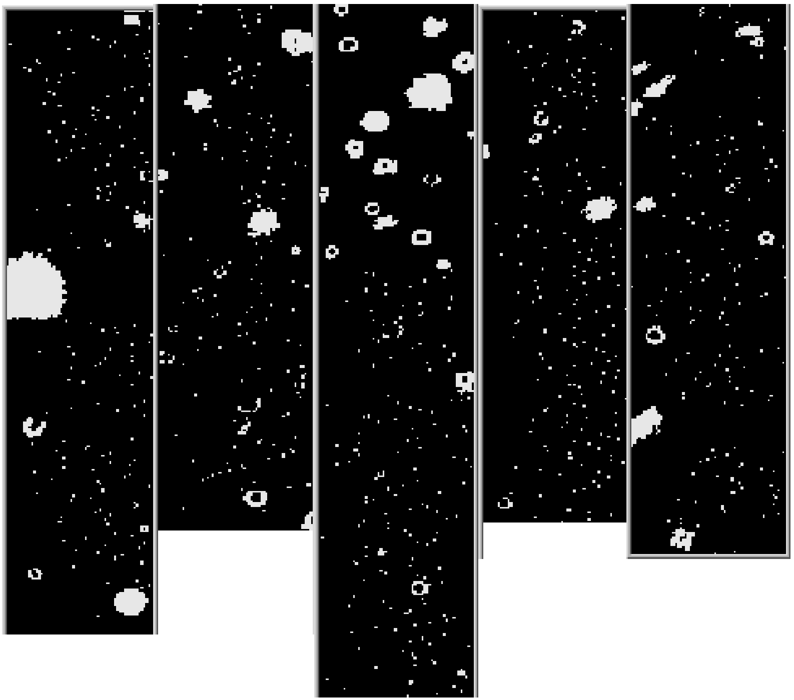

Fig. 7. Hydrometeor CIP images at ~7.2 km of a cloud pass on 24 September 2009 over Nanded The images of cloud droplets, ice crystals and graupels could be seen. The width of the view is 1.5 mm.

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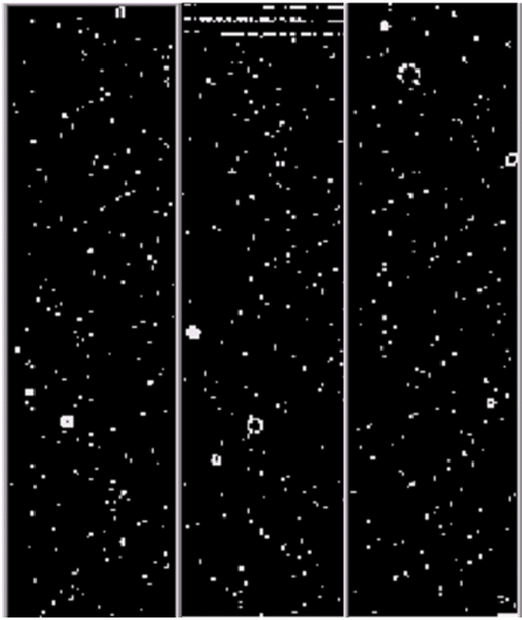

Fig. 8. Hydrometeor CIP images at ∼5.8 km of a cloud pass on 22 June 2009 over Raichur. The images of cloud droplets and drizzle that just start forming at this height could be seen. The width of the view is 1.5 mm.

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