Is the atmospheric boundary layer altitude or the strong thermal inversions that control the vertical extent of aerosols?

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**HIGHLIGHTS**

- ABL altitude is not the deciding factor for the vertical distribution most of the time.
- The vertical distribution of aerosols is restricted by the strong temperature inversion.
- Trapping of aerosols fully occurs up to particular strength and depth inversion only.
- These findings will help in modeling the diffusion and transport of air pollutants.

**GRAPHICAL ABSTRACT**

**ABSTRACT**

It is well known that the atmospheric boundary layer (ABL) plays a significant role in controlling the variability of atmospheric constituents such as aerosols and trace-gases. Hence, significant diurnal and seasonal variation in these will be observed as the ABL altitude does. However, on several occasions, high aerosol concentration in the lidar measurements is observed even above the ABL altitude. This raised a question that up to what extent ABL altitude acts as a capping layer for these pollutants? From the detailed analysis carried out using long-term (2010–2018) lidar observations and simultaneous radiosonde profiles obtained from Gadanki, India, we show that ‘there exist thermal inversions (TI), which are stronger than the ABL inversions, that fully control the vertical extent’. The detailed characteristics of TI (inversion strength (IS) and inversion depth (ID)) are also obtained. The results revealed that aerosol concentrations below the TI altitude increases with IS (ID) up to 3–4 K (300–400 m) during winter whereas in pre-monsoon it increases up to 2–3 K (100–200 m). Thus, IS of up to 2–4 K is required to fully trap the aerosol concentrations and this TI coincide with the ABL inversions for 51.7% only, particularly during the winter and pre-monsoon seasons. This analysis is further extended to different geographical locations of India using the aerosol profiles obtained from CALIPSO and a network of 23 radiosonde stations. The observed results provided further evidence that the vertical distribution of aerosols is restricted to the maximum extent by the TI but not the ABL altitude. These observations lead us to propose a hypothesis that ‘trapping of aerosols fully occurs up to particular IS and ID only and the ABL altitude is not the deciding factor most of the time for capping the aerosol vertical distribution’. These findings will greatly help in modeling the diffusion and transport of air pollutants in the lower troposphere.

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**1. Introduction**

Atmospheric aerosols, both natural and anthropogenic, play a significant role in the air quality and climate change (Boucher et al., 2013).
Aerosols and its optical properties as well as trace-gases exhibit large temporal and seasonal variation in their concentrations and are closely related to prevailing meteorological conditions, regional and long-range transport (Jain et al., 2021a; Nandan et al., 2021; Zhang et al., 2021). The increase in aerosol concentration near surface is not only attributed to sudden changes in the emission sources but mainly associated with changes in the meteorological conditions (Cheng et al., 2007). The meteorological variables such as atmospheric boundary layer (ABL), thermal/temperature inversions (TI) and atmospheric stability are the most important drivers for controlling atmospheric pollution dispersion (Turner, 1970). In particular, ABL is one of the crucial parameters which is directly coupled with the earth surface by exchanging heat, moisture, aerosols and other pollutants (Deardorff, 1972; Stull, 1988). The aerosols produced from different sources are largely confined below the ABL than the rest of the atmosphere. Most of the studies have revealed a pronounced interaction between aerosols and the ABL, which could have a significant influence on air quality (Guo et al., 2016, 2019; Li et al., 2017; Miao and Liu, 2019; Quan et al., 2013). Dedicated efforts are made to investigate the changes in the surface and vertical distribution of aerosols based on the evolution of the ABL (Gupta et al., 2021; Jain et al., 2021b; Lou et al., 2019; Nandan et al., 2021; Prasad et al., 2019; Ratnam et al., 2018, 2021).

It is well reported that, the surface level aerosol (for example Black Carbon) and trace-gases (for example CO2 and CH4) concentrations can vary significantly within a day (Jain et al., 2021b; Ravi Kiran et al., 2018). Nandan et al. (2021) has reported that diurnal and seasonal variation of surface aerosol number concentration, scattering and absorption coefficients are also strongly depend on the ABL altitude and other meteorological parameters. All these studies have mentioned that, changes in the surface aerosol concentrations and coefficients are mainly controlled by the dynamics of ABL in the lower troposphere. The Aerosol Optical Depth (AOD) is another important aerosol optical property that represents the total columnar concentrations of aerosol from surface to top of the atmosphere. The diurnal variation in AOD has been studied over different regions and found weak diurnal variation unlike surface concentrations (e.g. Lennartson et al., 2018; Madhavan et al., 2021; Pandithurai et al., 2007). Pandithurai et al. (2007) have found slight decrease in the AOD around noon, followed by a slight increase in the evening during winter. Whereas in pre-monsoon, low AOD are observed from the morning to the noon and thereafter started increasing over Pune, India. Similarly, Madhavan et al. (2021) also noticed a slight decrease in AOD till noon followed by an increase until evening during winter and pre-monsoon seasons over Gadanki, India. Whereas no change in the AOD is observed during monsoon and post-monsoon seasons up to 16:00 IST (=UTC + 05:30 h) with a small deviation of about ±2–3%, afterwards a sudden decrease is noticed till 17:00 IST during post-monsoon. In another study, Lennartson et al. (2018) also found very less deviation in the AOD (10% in coastal rural sites and 20% in coastal urban sites) from the daily average using long-term AERosol RObotic NETwork (AERONET) observations over South Korea. They attributed the observed diurnal variation in AOD to the changes in the meteorological conditions such as TI, ABL, humidity, winds and other factors. Zhang et al. (2021) has showed the evolution of vertical distribution of aerosol in the ABL and found a distinct structure under haze weather conditions over Wuhan, China. They observed accumulation of large number of particles below 0.5 km during morning hours due to stable conditions including a strong surface-based inversion, late development of the convective boundary layer and weak wind. A large reduction in the boundary layer aerosols is also noticed during afternoon hours due to improved ventilation. Further, few studies have discussed the role of inversions on the surface and vertical distribution of aerosols. Wallace and Kanaroglou (2009) has investigated the effect of TI on the ground-level NO2 and PM2.5 and found 49% and 54% increase in their concentrations, respectively, during nighttime pollution episodes. Wang et al. (2018) have also reported that the TI occurred in the boundary layer leads to the accumulation of aerosols within the inversion layer during two air pollution events Beijing, China, by using specialized air-craft Lidar observations. In addition, field studies conducted at various locations over China also revealed that aerosols will accumulate mainly at the bottom of the inversion layer (e.g. Sun et al., 2012; Zhang et al., 2011).

From all these studies, the impact of ABL and/or TI on the surface concentrations is well understood but its role on the vertical distribution of aerosols remains unexplored. Moreover, most of these studies focused on individual air pollution episodes with limited time period observations. Hence, it is necessary to resolve the relation between ABL and TI on the vertical distribution of aerosols and decide which one is the dominant factor that controls the vertical extent of the aerosols. To understand this relation, long-term, high-quality, simultaneous radiosonde and aerosol observations are required. In this study, we obtained aerosol extinction profile from ground-based Micro-pulse Lidar (MPL) during 2010–2018 and temperature profile from high-resolution radiosonde observations over Gadanki (13.5°N, 79.2°E), India. Further, we make use of Cloud Aerosol Lidar Infrared Pathfinder Satellite Observations (CALIPSO) and simultaneous radiosonde measurements from India Meteorological Department (IMD) to investigate their relationship spatially. Such investigations will greatly help in modeling the diffusion and transport of air pollutants in the lower troposphere (Li et al., 2015).

2. Data

2.1. Micro pulse Lidar (MPL)

For obtaining the aerosol properties (aerosol extinction profile), we used Micro Pulse Lidar (MPL) being operated at National Atmospheric Research Laboratory (NARL) since 2010. This MPL consists of Nd:YAG laser operated at 532 nm wavelength with an energy of 4 J/pulse and pulse repetition rate of 2500 Hz. This Lidar is operated at night times mostly after sunset during clear sky and partially clear sky conditions from August 2010 to December 2018. MPL provides raw photon counts from the surface to 45 km stored at 1-minute temporal resolution and 30 m vertical resolution. To remove the errors in the aerosol extinction profiles, we have performed noise and range correction to the raw photon counts as already discussed in the earlier literatures (Gupta et al., 2021; Prasad et al., 2019). By using Fernald inversion method (Fernald, 1984), we estimated aerosol extinction from the Lidar by considering Lidar Ratio as 40 and reference height at 8 km. The complete technical details of MPL and methodology of estimating aerosol extinction profiles is provided in the previous studies (Gupta et al., 2021; Prasad et al., 2019; Ratnam et al., 2018). The hourly averaged aerosol extinction profiles derived from MPL during 2010–2018 is used in this study. We also estimated AOD by integrating the aerosol extinction from surface to the desired altitudes (ABL and top of the atmosphere).

2.2. Sky radiometer

AOD is also obtained from the Sky Radiometer being operated at NARL since 2008. The Sky Radiometer (POM-01L, Make: Prede Co. Ltd., Japan) is an automatic ground-based scanning spectral radiometer. It measures the direct and diffuse sky radiance at 5 different wavelengths (400, 500, 675, 870, and 1020 nm) at 1 min and 10 min intervals, respectively. The field of view of this instrument with respect to the Sun is 1°. Sky Radiometer provides various aerosol optical properties such as AOD, asymmetry parameter, and single scattering albedo etc., based on the SKYRAD software (Nakajima et al., 1996). In this study, we used AOD at 500 nm during 2010 to 2018. More details of this instrument and retrieval methods are described in Madhavan et al. (2021).

2.3. CALIPSO measurements

The ground-based Lidars will provide the columnar distribution of aerosols with good vertical and temporal resolution but it is difficult
to obtain profiles spatially. In order to obtain the vertical distribution of aerosols spatially, Cloud-Aerosol Lidar with Orthogonal Polarization (CALIOP) onboard Cloud Aerosol Lidar and Infrared Pathfinder Satellite Observations (CALIPSO) are used. CALIPO is a space-borne elastic backscatter Lidar designed to study the vertical distribution of aerosols and clouds (Winker et al., 2009). It is a nadir viewing dual wavelength (532 and 1064 nm) satellite inclined at 98° at an altitude of 705 km with a 16-day repeat cycle launched in April 2006 (http://www.calipso.larc.nasa.gov/). Level 2, version 4.20, 5-km aerosol profile products from the CALIPSO are used in the present study. By applying the Cloud Aerosol Discrimination (CAD) score ranging from (−100 to −20), it discriminates the aerosols from the clouds and other layers. All the quality control flags are applied to the Level 2 data prior to use in the analysis (Winker et al., 2013). One of the main limitations of the space-borne lidars is the signal contamination from the surface which reflects in the aerosol extinction profile (Tackett et al., 2018). Due to this reason, we have considered aerosol extinction profile from –100 m above from the surface. Previous studies have showed the aerosol extinction profile derived from spaceborne lidar CALIPSO compares very well with ground-based MPL measurements over Gadanki (Gupta et al., 2021; Prasad et al., 2019; Ratnam et al., 2018). The CALIPSO measurements have large bias in the day time measurements than the night time due to contamination from several other sources (Mamouri et al., 2009). In the present study, we utilized aerosol extinction profile at 532 nm only at night time overpass from CALIPSO observations over a given radiosonde station of IMD (Table S1). The grid size of ±0.5” to the latitude and longitude of corresponding radiosonde station is chosen from CALIPSO observations.

2.4. Radiosonde measurements

For obtaining the information on the thermal inversions and boundary layer altitude, we used high-resolution GPS radiosonde launched regularly at 12:00 UTC (17:30 IST) from NARL since April 2006. These radiosondes provided profiles of atmospheric parameters such as Temperature (T), Pressure (P), Relative Humidity (RH) and horizontal winds up to an average altitude of ~30 km. The raw data were recorded at sampling interval of 1 s with vertical resolution of 5–6 m (Ratnam et al., 2014). In this study, we have considered total of 2181 Meisei radiosonde observations launched at 1200 UTC over Gadanki during 2010–2018. The quality checks have been applied to maintain continuity and consistency in the measurements (Ratnam et al., 2014). The radiosonde profiles are interpolated to 30 m to remove the outliers and to match with the MPL measurements. Further, we have considered data from the 23 IMD radiosonde stations over India covering seven different geographic regions based on the spatial heterogeneity of AOD and topography as shown in Fig. S1. The accuracy of the temperature profile measured from radiosonde is less than 0.1 K for Gadanki radiosonde and 0.5 K for IMD radiosonde. This AOD (combined Deep Blue (DB) and Dark Target (DT) product) is obtained from MODerate resolution Imaging Spectroradiometer (MODIS) satellite averaged during 2001–2018 (https://giovanni.gsfc.nasa.gov/). From this figure it is clear that there exists large spatial heterogeneity in the AOD over India and surrounding regions with higher (lower) AOD over Indo-Gangetic Plain (IGP) (south India and north west India) regions. The available IMD radiosonde stations cover most of this heterogeneous variability including the coastal regions. Since, we have selected CALIPSO night time –20 UTC (−01:30 IST) overpass observations, the closest time of IMD radiosonde data is at 00 UTC (05:30 IST). We have obtained quality controlled IMD radiosonde observations data from the Integrated Global Radiosonde Archive (IGRA) (Durre et al., 2006). Since this dataset is available with coarser resolution, we have used 100 m interpolated temperature data. Total number of profiles available from Gadanki and IMD radiosondes is illustrated in Fig. S1(b) and (c), respectively.

3. Methodology

3.1. Estimation of inversion height and its parameters

We have estimated inversion parameters, such as base and top of the temperature inversion layers and their corresponding magnitude of temperatures, from the radiosonde. For this analysis, we have considered only single (and strongest) inversion days in the radiosonde launched at 17:30 IST. The altitude at which the temperature start increasing is defined as inversion base height (IBH) and the altitude where temperature begins to decrease as inversion top height (ITH) (Fig. 1). The altitude difference between the inversion base (Hb) and top (Ht) is termed as inversion depth (ID) and difference in the temperatures between the respective heights is termed as inversion strength (IS). Inversions are identified by taking first derivative of the temperature profile (Kahl, 1990; Serreze et al., 1992). We have considered an inversion only when ΔT ≥ 1 K and Δz ≥ 60 m (twice the uncertainty in the measurement and the vertical resolution). Typical temperature and refractivity profiles along with their gradients observed on 16 January 2015 at 17:30 IST is depicted in Fig. 1. In this case, an inversion is identified with base (top) height at 2.44 km (2.65 km) and corresponding temperature is 283.4 K (289.1 K). The IS and ID for this example is ~5.7 K and ~210 m, respectively. Similar analysis is performed for all the radiosonde profiles during 2010 to 2018 and the number of profiles that have statistically significant inversions during different seasons are provided in Fig. S1(b). We have also estimated the IS and ID over other parts of India using IMD radiosonde network data by considering the minimum IS and ID of about 0.5 K and 100 m, respectively. The number of profiles that satisfy this criterion is shown in Fig. S1(c). We also tested the results by changing the IS and ID criteria with Gadanki radiosonde data, but the result does not show any significant change except in the number of available days. Due to this reason, we have chosen IS and ID of 0.5 K and 100 m for other radiosonde stations in this study.

3.2. Detection of the ABL altitude

For ABL altitude detection, we used gradient in the refractivity (Basha and Ratnam, 2009) and its method of detection is shown in Fig. 1(c). Since the refractivity consists of both temperature and water vapor content, and water vapor show the sharp reduction even at the small temperature inversion, it was shown that this parameter is the most suitable for detection of ABL altitude (Basha and Ratnam, 2009). Note that the virtual potential temperature also consists of both temperature and water vapor content, but the contribution of water vapor in this parameter is much less than the refractivity (Basha and Ratnam, 2009) (see supplementary information for more details on the ABL detection). Thus, maximum negative gradient in the refractivity is identified as ABL altitude. Fig. 1(c) and (d) shows the typical profiles of temperature, gradient in temperature and refractivity obtained on 3 January 2018 and 21 December 2015 at 17:30 IST launched from NARL, Gadanki, respectively. Strong negative gradient in refractivity at ~2.61 km and ~1.42 km corresponds to the ABL altitude for these two typical cases, respectively. Note the ABL altitude detected using wavelet covariance transform technique (Ratnam and Basha, 2010) is also shown and there exists small difference between the two.

4. Results and discussion

4.1. Gadanki location and background meteorological conditions

The observational site NARL, Gadanki (13.5°N, 79.2°E) is a tropical rural location situated about 120 km north-west of Chennai and about 130 km from the Bay of Bengal (BoB) in the southern peninsular India. Gadanki is small village surrounded by the agricultural fields where burning of wood and forest fires are observed occasionally (Jain et al.,
Fig. 1. (a) Typical vertical profile of temperature (red) and refractivity profile (blue) observed on 16 January 2015 at 17:30 IST over Gadanki. The dashed black represents the base and top of the inversions, respectively. (b) Temperature (dashed red line) and refractivity gradient (dashed blue line) observed for the same profiles. Typical profiles of (c) temperature (red line) aerosol extinction (black line) and gradient in temperature (dashed red line) and refractivity (dashed blue line) profiles observed on 03 January 2018 (17:30 IST) over Gadanki. (d) Same as (c) but observed on 21 December 2015. Blue, black and red solid circle represents ABL estimated by the refractivity, wavelet covariance technique (WCT), and inversion height respectively. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
A National Highway (NH-18) passes near to the measurement site and experiences moderate to heavy traffic. The climatological monthly mean meteorological parameters (temperature, relative humidity, zonal and meridional winds) observed over Gadanki using radiosonde measurements averaged from 2010 to 2018 is shown Fig. 2. Though radiosonde data available up to 28 km on an average, we show up to 5 km in this figure that is relevant to the present study. The maximum temperature has been noticed up to ~1 km during pre-monsoon and minimum during winter months (Fig. 2(a)). The maximum relative humidity is observed during monsoon and post monsoon seasons with peak between 2 and 4 km (Fig. 2(b)). During winter months, relative humidity is restricted to lower altitudes (<2.5 km). Strong westerlies are observed in zonal wind (Fig. 2(c)) during monsoon season which indicates presence of low-level jet (LLJ). Most of the moisture will be transported from Arabian Sea through LLJ which leads to high relative humidity during monsoon season (Ratnam et al., 2018). A strong wind (>2 m s⁻¹) in the meridional winds is noticed during pre-monsoon and very weak northerly winds are observed in other seasons (Fig. 2(d)). Gadanki experiences both southwest and northwest monsoon which attribute to wet scavenging of aerosols mainly below the boundary layer which leads to clean environment (Ratnam et al., 2018). Based on the prevailing meteorological conditions over the observational site, the seasons are classified as winter (December–January–February (DJF)), pre-monsoon (March–April–May (MAM)), monsoon (June–July–August (JJA)), and post-monsoon (September–October–November (SON)).

4.2. The role of ABL on the surface and vertical distribution of aerosols

Before proceeding further, we briefly discuss the role of ABL on the surface aerosol and trace gases at different time scales. Monthly mean values of scattering coefficient, absorption coefficient and aerosol number concentration obtained from Nephelometer, Aethalometer and Aerodynamic particle sizer, respectively, from NARL, Gadanki is shown in Fig. 3(a). These observations are averaged during 2008 to 2015 (re-plotted from Nandan et al., 2021). Higher (lower) coefficients and number concentrations are observed during winter (monsoon) months. Similar behavior is observed not only in the aerosol concentrations but also in the trace gas (CO₂ and CH₄) concentrations shown in Fig. 3(b). These are obtained from NARL, Gadanki using Cavity Ring-Down Spectrometer (CRDS) during 2016 to 2019 (re-plotted from Jain et al., 2021a). These variations match well with the ABL altitude shown in Fig. 3(d) with higher (lower) altitude during pre-monsoon to monsoon (winter) months obtained from radiosonde observations during 2010 to 2018. Thus, it is clear that surface aerosol and trace gas concentrations are mostly modulated by the ABL dynamics.

We further investigated the role of ABL altitude on these aerosol and trace gases at sub-daily scales. Annual mean diurnal variation of scattering coefficient, absorption coefficient, and Aerosol number concentration obtained from the instruments mentioned above is shown Fig. 3(e). We also included diurnal variation in the Black Carbon (BC) mass concentration obtained from Aethalometer. A clear increase (decrease) in all these concentrations during night (day) times can be noticed. A small increase in all these concentrations around 07:00–08:00 IST is due to fumigation effect (Ratnam et al., 2021; Talukdar and Ratnam, 2021). Similar behavior is also seen in the diurnal variation of trace gas (CO₂ and CH₄) concentrations (Jain et al., 2021a) shown in Fig. 3(f). These variations match well with the annual mean diurnal variation observed in the ABL altitude shown in Fig. 3(h). This figure clearly shows the shallow ABL altitude during morning hours and attains maximum altitude in the afternoon hours and gradually decreases in the evening. ABL altitude is almost close to ~1 km during night time and reaches maximum up to 2.5 km during the day time. The number of radiosondes used to estimate the ABL altitude for the respective hour is also superimposed in Fig. 3(h). The changes in the observed aerosol coefficients and trace gas concentrations show inverse relation with the diurnal variation in the ABL height. The higher surface concentrations and coefficients during night to morning and evening hours are mainly attributed to the low ABL altitude. Whereas, in the afternoon hours due to strong vertical mixing, the boundary layer starts to evolve and leads to decrease the surface level concentrations and coefficients during the daytime. Thus, it is clear that the ABL altitude play a significant role in controlling the variability of surface aerosol and trace gas concentrations.

In contrast, such interplay between ABL altitude evolution and AOD is not observed. To our surprise, monthly mean AOD obtained from MPL during 2010–2018 shown in Fig. 3(c) (re-plotted from Prasad et al., 2019) does not show this behavior. Though the higher ABL height is noticed during pre-monsoon to monsoon season (Fig. 3(d)) but also higher AOD has been noticed (Fig. 3(c)). This could be due to two things: either the ABL altitude might not affect the aerosol concentrations during these months or these aerosols might be due to long-range transport from the altitude above ABL. Using long-term (2010–2018) lidar observations, Ratnam et al. (2018) clearly shown the presence of the elevated aerosol layer during monsoon months over Gadanki due to long-range transport through low-level jet (LLJ) from Arabian sea. Further, Prasad et al. (2019) showed that aerosols present above the ABL altitude provides about 80–90% contribution to the total AOD during these months. The higher AOD during pre-monsoon may be due to increased concentration of continental aerosols.

Fig. 2. Composite monthly mean (a) temperature (b) relative humidity (c) zonal wind (d) meridional wind obtained using GPS radiosonde measurements over Gadanki averaged from 2010 to 2018. Composite monthly mean ABL altitude (black line) along with standard deviation (vertical bar) is also superimposed in (a).
due to strong surface winds which lifts the aerosols from the loose soil (Raja Obul Reddy et al., 2016). Further, high surface temperatures during this season may lead to increase the photochemical process which favors the hygroscopic growth of aerosols and leads to formation of secondary aerosols (Babu et al., 2013). Thus, aerosol concentrations during these months do not coincide with the variations in the ABL altitude. Note that the time period of the observations used in Fig. 3 from various instruments is not the same but they reasonably represent the diurnal and monthly/seasonal variations since large data base has been used.

We further investigated the relation between ABL altitude and AOD using the observations obtained from Gadanki during 2010 to 2018. The diurnal variation in AOD is obtained by combining the MPL (during night time between 19:00 IST and 06:00 IST) and Sky Radiometer (during day time between 07:00 IST and 17:00 IST) measurements and is shown in Fig. 3(g). The estimated AOD does not show any significant diurnal variation (less than 10% variation) and show almost constant AOD value around 0.5 except at 06:00 IST and 18:00 IST, where uncertainty in the AOD estimation is high. Thus, it is clear that the AOD at both monthly and sub-daily scales do not coincide with the variations in the ABL altitude. Note that AOD is an integrated aerosol property that represents the columnar aerosol concentrations from surface to the top of the atmosphere. This creates a basic curiosity to investigate the actual role of ABL on the vertical distribution of the aerosols.

4.3. Role of ABL altitude on the vertical distribution of aerosols at different temporal scales

In order to further investigate the role of ABL altitude on the vertical distribution of aerosols, we show temporal evolution of the ABL altitude and its effect on aerosol vertical distribution in Fig. 4. ABL altitude is shown for the every 3 h from the radiosonde observations conducted over Gadanki as a part of tropical tropopause dynamics (TTD) campaign (Hemanth Kumar et al., 2015; Ratnam et al., 2014) on 18–19 January 2012. Although radiosonde launches are initiated every 3-hour interval, we have considered observations only at 17:30, 20:30, 23:30, 02:30, 05:30 IST to match with the aerosol observations obtained from MPL during night times only on the same day. The ABL height is observed at ~2.5 km at 17:30 IST and after 20:30 IST, the stable boundary layer (SBL) is formed at ~1 km throughout night which matches well with the values shown in the diurnal variation of ABL in Fig. 3(h). During the night times, ABL altitude comes closer to the surface and forms a stable layer which leads to increase in the surface aerosol and trace gas concentrations (Stull, 1988).

In similar lines, we have investigated how this ABL altitude will modulate the vertical distribution of aerosols during night time. For this purpose, we have considered mean aerosol extinction profile 30 min before and after the radiosonde launch (except at 17:30 IST) and is shown in Fig. 4. At 17:30 IST, the aerosols are distributed uniformly (average extinction of about ~0.05 km$^{-1}$ throughout a profile) below the ABL i.e., at ~2.5 km and drops completely at this altitude (Fig. 4(a)). Similar features are also observed in the example shown in Fig. 1(c). Though inversion persist at 2.7 km even at 20:30 IST with stronger gradients, a small inversion develops at ~1 km at the same altitude (Fig. 4(b)) which is considered as SBL as explained in Mehta et al. (2017). The presence of this small inversion is reflected in vertical distribution of aerosol with a small drop at ~1 km but enhanced aerosol concentration still persists until 2.5 km (Fig. 4(b)). The aerosol extinction below SBL is almost doubled (0.1 km$^{-1}$) at 23:30 IST with stronger SBL but small concentrations of aerosol still persist even above the SBL (Fig. 4(c)). If we monitor surface concentrations in the aerosol and the trace gases during these timings, it is natural to observe enhanced concentrations. Note that the SBL formed after 20:30 IST remains at ~1 km and becomes stronger and stronger with time until 05:30 IST and significant reduction in the aerosol concentrations is being noticed at this altitude. However, large amount of aerosols still persists above this SBL and get trapped completely around ~2.8 km altitude where the stronger
temperature inversion persists. These typical examples clearly suggest that though ABL (SBL) altitude significantly affect the aerosol concentrations near the surface but unable to completely trap at this inversion. Similarly, in Fig. 1(d), we observed (on 21 December 2015) a sudden drop in the aerosol concentrations at ~1.4 km, but still higher concentrations are noticed even above but completely trapped around ~2.8 km.

The estimated AOD below and above the lower inversion is 0.02 and 0.10, respectively. The higher concentrations above the ABL or SBL are not only observed during particular night but also noticed on several days in all the seasons.

Fig. 5 shows typical profiles of temperature and gradients in the temperature and refractivity obtained from radiosonde launched over
Gadanki around 17:30 IST along with mean aerosol extinction profile obtained from co-located MPL observations during winter and pre-monsoon seasons. In these examples, we have considered mean aerosol extinction profiles from MPL measurements obtained during 19:00 to 21:00 IST, which is the closest time of radiosonde launching, to understand the role of ABL on the vertical distribution of aerosols. A strong negative gradient in the refractivity (which is considered as ABL altitude) is found at ~2.5 km and ~2.4 km on 14 January 2016 and 13 January 2016, respectively, where the higher extinction values in the aerosols distributed up to these altitudes (Fig. 5(a) & (b)). However, on 21 December 2015, though the ABL altitude is found at ~1.5 km, higher extinction values persist up to ~2.8 km and get trapped completely, where the stronger temperature inversion exists. Similarly, during pre-monsoon days (29 March 2011 and 23 May 2016), the vertical distribution of aerosols is not completely trapped at the ABL altitude (Fig. 5(d) & (f)). Only in few cases, aerosols are completely trapped at the ABL altitude (~1.8 km) as observed on 07 March 2016 (Fig. 5(e)). From these examples, it is again confirmed that ABL altitude alone is not a dictating factor for controlling the vertical distribution of aerosols. It is interesting to note that the higher extinction in the aerosols is distributed up to strongest gradient in the temperature inversion (inversion height) rather than the ABL altitude (Fig. 5(c)–(f)). It has been already emphasized that any small perturbation in the temperature profile within or beyond the ABL in association with a higher concentration of (BC) aerosol can significantly affect the positive buoyancy required for rising air parcel and hence the stability (Talukdar et al., 2019). High BC concentrations enhance the atmospheric stability and suppress the atmospheric turbulence. The reduced turbulence mixing affects the evolution of boundary layer which lowers the boundary layer height (Liu et al., 2019). On the other hand, the reduced turbulent mixing also leads to higher relative humidity (RH) because of weakened mixing of moist air up through the interface and weakened mixing of dry tropospheric air down into the mixed layer (Wilcox et al., 2016). Thus, the increased BC concentration was attributed to the presence of high RH and lowering of the ABL (Liu et al., 2019). Talukdar et al. (2019) also showed high concentration of surface BC decrease the convection and increases the stability conditions. However, conditions will be completely different if elevated layer of BC is present. In this case, additional warming at the elevated layer may occur which will affect the ABL strength and depth. We have investigated the effect of surface BC on the ABL altitude over Gadanki. For this purpose, we have taken ABL height from the radiosonde launched during 17:00 IST and mean surface BC during concentration during 16:00 to 18:00 IST from the Aethalometer observations over Gadanki. Fig. S2 shows that as BC concentrations increase ABL height decrease. In next sub-section, we will investigate how much strength and depth in the temperature inversion is required to fully trap the aerosol concentrations.

4.4. Role of temperature inversions on the vertical distribution of aerosols

On 14 January 2016, a single inversion is observed around ~2.5 km where both the refractivity and temperature gradients coincide exactly and the aerosols also distributed up to the inversion altitude and got trapped completely here (Fig. 5(a)). Double inversions are observed in the temperature profile one at ~2.5 km and other at ~5.2 km on 13 January 2016 (Fig. 5(b)). However, aerosols are completely trapped at the lower inversion itself which is the strongest. In case of multiple inversions which was observed (at ~1.5 km, ~2.8 km, ~4.2 km and ~4.8 km) on 21 December 2015 (Fig. 5(c)), a strongest inversion exists at ~2.8 km where most of the aerosols are trapped. Though several weak inversions are noticed on this day, but the trapping occurred at the strongest inversion only as expected. During pre-monsoon, similar features are observed in trapping of aerosols (Fig. 5(d)–(f)). Most of the aerosols are confined below the inversion (~4 km) on 29 March 2011 (Fig. 5(d)). Two inversions (at ~1.8 km and ~4.5 km) are noticed on 07 March 2016 (Fig. 5(e)) but concentrations of aerosols are mainly restricted at the lower inversion which is the strongest. Although multiple inversions (at ~1.6 km, ~2.3 km and ~5.4 km) are found on 23 May 2016 (Fig. 5(f)), but the aerosols are vertically distributed and got trapped at the strongest inversion at ~5.4 km altitude. Note that ABL altitude is at 2.2 km but the aerosols are trapped at the strongest inversion. This strongest temperature inversion does not coincide with the ABL altitude at all the times. Moreover, one cannot expect ABL altitude extending up to 4–5 km and above in the normal conditions. These results clearly provide evidence that complete trapping of aerosols occurs only when the strongest inversion exists irrespective of the altitude, time and season. In few cases, we have observed that ABL altitude coincide exactly with the inversion altitude but not in most of the cases. We have estimated the how much percentage of the ABL altitude match with the inversion altitude during the entire study period. It is found that only ~36.5% times of the ABL altitude match with the inversion altitude with highest match in winter season (48.4%) followed by pre-monsoon (42.2%) and post-monsoon (28.5%) and lowest in monsoon season (25.1%). When the inversion strength of greater than 1 K (double the uncertainty of the temperature measurement) is only considered, about 51.7% of the time ABL altitude match with the strongest inversion altitude with 58.6%, 58.5%, 42.0% and 32.5% during winter, pre-monsoon, post-monsoon and monsoon seasons, respectively. Rest of the time the strongest inversion occurred always well above the ABL altitude (Table 1). Thus, it is clear that for about 50% of the time only, ABL altitude matches with the strongest inversion altitude. It is expected that around similar percentage of aerosol only will be trapped at the ABL altitude and rest of it at the strongest inversion altitude. In order to check this aspect, AOD is estimated from the aerosol extinction profiles from MPL observations averaged during 19:00 IST to 21:00 IST (closest time of radiosonde launch). From Fig. 3(g), it is clear that the AOD doesn’t show any significant diurnal variation. AOD is estimated from surface to the ABL altitude and from ABL altitude to 8 km separately and is shown in Fig. 6(a). In Fig. 6(b), we also show AOD estimated from the surface to the strongest inversion altitude and from this inversion altitude to 8 km. The number of days of the data used during different seasons is also depicted in this figure. From this figure it is clear that the significant amount of aerosol still exists above the ABL altitude in all seasons except during winter, where much of aerosols are being trapped at ABL altitude itself. However, significant amount of aerosol concentrations is reduced above the strongest inversion altitude but a noticeable amount is still persisting above particularly during monsoon and post-monsoon seasons. This could be due to aerosol present at higher altitudes (above ABL and strongest inversion altitudes) that is being transported directly through long-range transport (Prasad et al., 2019; Ratnam et al., 2018). A typical example for such

<table>
<thead>
<tr>
<th>Season</th>
<th>Inversion strength (K)</th>
<th>Inversion depth (m)</th>
<th>Inversion base height (km)</th>
<th>Inversion top height (km)</th>
<th>ABL (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Winter</td>
<td>2.74 ± 1.51</td>
<td>199.19 ± 99</td>
<td>2.37 ± 0.60</td>
<td>2.57 ± 0.61</td>
<td>2.23 ± 0.71</td>
</tr>
<tr>
<td>Pre-monsoon</td>
<td>1.87 ± 0.80</td>
<td>142.98 ± 61</td>
<td>3.07 ± 1.06</td>
<td>3.22 ± 1.08</td>
<td>2.68 ± 0.87</td>
</tr>
<tr>
<td>Monsoon</td>
<td>1.68 ± 0.87</td>
<td>147.07 ± 69</td>
<td>3.71 ± 1.21</td>
<td>3.86 ± 1.22</td>
<td>2.75 ± 1.02</td>
</tr>
<tr>
<td>Post monsoon</td>
<td>1.99 ± 0.97</td>
<td>175.91 ± 102</td>
<td>3.17 ± 1.17</td>
<td>3.35 ± 1.16</td>
<td>2.46 ± 1.01</td>
</tr>
</tbody>
</table>
case is shown in Fig. S3 where significant amount of aerosol persists above the inversion height and equal amount is being trapped below the strongest inversion altitude. In such cases, neither the ABL altitude nor the strongest inversion will influence the aerosol vertical distribution. Thus, our results clearly suggest that the ABL is a good indicator to control the vertical distribution of aerosols during winter, whereas TI plays a significant role in dictating the vertical extent of aerosols during pre-monsoon. However, the effect of ABL/TI on the vertical extent of aerosols is not much noticed during monsoon and post-monsoon season.

From the abovementioned examples, it is clear that each temperature profile shows different inversion strength and depth and trapping occurs mainly at the strongest temperature inversion only (Figs. 4 and 5). In next sub-section, we discuss the inversion statistics and at what inversion strength and depth, the complete trapping of aerosols takes place.

4.5. Statistical characteristics of temperature inversions

In the previous section, we have noticed that the strong temperature inversions formed in the lower troposphere have more impact on the vertical distribution of aerosols. The IS indicate the stability of the atmosphere and ID represents the vertical extent of the inversion (Pavelsky et al., 2011). The IS and ID will have different ranges and mainly depend on the seasons and geographic dependencies (Guo et al., 2020). Fig. 7 depicts the monthly and seasonal variation in the IS and ID over Gadanki during 2010–2018. The maximum and minimum in IS are observed during January (~3.01 ± 1.60 K) and July (~1.56 ± 0.59 K), respectively (Fig. 7(a)). Deep and shallow ID is observed during January (~205.42 ± 98.4 m) and August (~142.72 ± 55.6 m), respectively (Fig. 7(b)). The mean inversion top (base) height is found to be maximum of about 4.24 ± 1.41 km (~4.08 ± 1.39 km) during June and minimum of about ~2.43 ± 0.69 km (~2.22 ± 0.69 km) in December (Fig. 7(c)). Higher values of ITH and IBH during pre-monsoon and monsoon are mainly attributed to the strong surface heating prevailing during these seasons causing strong vertical mixing which pushes the inversion to higher altitudes (Li et al., 2019). In addition, strong seasonal variation is observed in all the inversion parameters (Table 1). The maxima in IS is noticed during winter (~2.74 ± 1.51 K) followed by post-monsoon (~1.99 ± 0.97 K), pre-monsoon (1.87 ± 0.80 K) and monsoon (1.68 ± 0.87 K) (Fig. 7(d)). Large variability in ID is noticed in winter (~199.19 ± 99 m) followed by post-monsoon (~175.91 ± 102 m), monsoon (~147.07 ± 69 m) and pre-monsoon (~142.98 ± 61 m) (Fig. 7(e)). During monsoon months, inversion top (base) height varies from ~0.61 to ~5.92 km (~0.46 to ~5.74 km) whereas in winter it varies from ~1.15 to ~4.87 km (~1.06 to ~4.69 km) (Fig. 7(f)).
The percentage occurrence of inversion parameters (i.e., strength, depth, base and top) for different ranges are provided in Table 2. The occurrence of different IS (ID) magnitudes are distributed over a wide range from 1 K to >5 K (60 m to >400 m) during winter season (Table 2). Low IS values (<3 K) are noticed during monsoon (~94%) and pre-monsoon (~88%) seasons. The ID values are found to be low (~200 m) during monsoon (~79%) and pre-monsoon (~81%) compared to winter (~56%) and post-monsoon (~66%). The percentage occurrence of ITH (~61%) and IBH (~57%) values are found to be particularly high in the range of 2–3 km during winter. During other seasons, the ITH and IBH value ranges from 1–2 km to 5–6 km. Large seasonal variability has been observed in these parameters over Gadanki region with stronger and deeper inversions during winter compared to other seasons. The variability in IS and ID mainly due to warm air advection, aerosols, greenhouse gases etc. (Abdul-Wahab, 2003, Abdul-Wahab et al., 2004; Bradley et al., 1993). There are several other potential factors which can also contribute to the variability in IS and ID based on the geographical and meteorological conditions (Guo et al., 2020). Iyer and Nagar (2011) have investigated IS and ID over 20 stations in the Indian region during 1971–2000 and reported that wind speed influences the sustaining of the inversion and magnitude of strength over different stations in India. They have also mentioned that urbanization, industrialization and increase in urban population are major causes for the observed changes in the IS and ID during study period. In next sub-section, the role of inversion parameters (IS and ID) on the vertical distribution of aerosols has been discussed.

### 4.6. The effect of inversion strength and depth on the vertical distribution of aerosols

We have estimated AOD by integrating the aerosol extinction profile below and above the inversion top height for different IS and ID during winter and pre-monsoon and is shown in Fig. 8. The number of observations considered for different ranges is also provided in the respective panels. It is very interesting to note that the AOD below the inversion increases with increasing strength up to 3–4 K and then decrease at 4–5 K (Fig. 8(a)). The AOD below the inversion increases as ID increases up to 300–400 m and then decreased at 400–500 m (Fig. 8(b)). The AOD above the inversion is very low irrespective of different ISs and IDs (Fig. 8(a) & (b)). During pre-monsoon, similar feature is noticed i.e., as IS (ID) increases, AOD below the inversion increases up to 2–3 K (100–200 m) and then decreases at 3–4 K (200–300 m). The AOD above the inversion height shows slightly higher values during pre-monsoon (Fig. 8(c) & (d)) consistent with that shown Fig. 6 (without discriminating according to the strength). These results clearly indicate that aerosols are being trapped at the inversion altitude up to certain...
inversion strengths and depths in a given season and after that it does not matter even further stronger inversion exists or not.

Note that the occurrence of strong inversions during monsoon and post-monsoon seasons are very low compared to other seasons (Fig. S1(b)). Further, these two seasons are mainly influenced by southwest and northeast monsoon over the observational site. Due to this reason, we have very less number of (clear sky) MPL observations during these two seasons. Nevertheless, the AOD above and below the inversion for different ISs and IDs in these seasons is shown in Fig. S4. During monsoon season, AOD doesn’t increase below inversion as IS and ID increases (Fig. S4(a) & (b)) whereas, in post-monsoon AOD decreases as IS and ID increases (Fig. S4(c) & (d)). In both the seasons, irrespective of IS and ID values, higher AODs are observed above the inversion (Fig. S4(b) & (d)). This is mainly due to long-range transport and wet-scavenging which influence the AOD particularly during the monsoon season (Prasad et al., 2019; Ratnam et al., 2018) irrespective of inversion strength and depth. Further, we already showed with an example in Fig. S3 to support the above statement. Thus, it is very clear that IS and ID has more influence on the vertical distribution of aerosols especially during winter and pre-monsoon over Gadanki.

The influence of inversion top temperature on the vertical distribution of aerosols is also investigated and is shown in Fig. S5. Interestingly, we noticed that inversion top temperature also modulates the AOD below and above inversion. The AOD below (above) the inversion decreases (increases) with increase in the inversion top temperature and stronger during pre-monsoon compared to the winter season (Fig. S5). Higher temperature (irrespective of altitude) leads to dispersion of the pollutants.

Further, we have also investigated how much AOD below the inversion is contributing to the total AOD for different inversion strengths using sky-radiometer measurements (co-located instrument) at Gadanki. For this purpose, we considered the sky-radiometer AOD between 15:00 to 17:00 IST on all the MPL data available days. Before that, we have compared MPL derived AOD during 19:00 to 21:00 IST and sky-radiometer AOD during 15:00 to 17:00 IST and is shown in Fig. S6. A very good correlation of about 0.79 between these two independent measurements is observed though different time periods have been chosen. This is expected as no significant diurnal variation in the AOD is observed (Fig. 3(g)). The percentage contribution of AOD from the below and the above inversion to the total AOD obtained for different inversion strengths during winter and pre-monsoon seasons is shown in Fig. S7. A maximum of ~64% (~58%) below the inversion AOD is contributing to the total AOD is observed at 3–4 K during the winter (pre-monsoon) season. These results provided further evidence that as strength increases, large fraction of aerosols is trapped below the inversion, which is contributing more to the total AOD.

The effect of inversion parameters (IS and ID) on the aerosol vertical distribution are also tested over other parts of India using the IMD radiosonde and CALIPSO measurements from 2006 to 2020 (Table S1). As mentioned earlier, the radiosonde stations over India were grouped into 7 different geographical regions based on the spatial heterogeneity of AOD and topography as shown in Fig. S1. The seven different geographical regions are denoted as Indo-Gangetic Plain (IGP), Central India (CI), North East (NE), South India (SI), East Coast (EC) and West Coast (WC). Fig. S8 shows the AOD below and above the inversion for different inversion strength and depth ranges during winter and pre-monsoon. The IGP and SI stations show a consistent increase in AOD below the inversion as IS increases up to maximum strength of about 2–3 K (Fig. S8(a)). During pre-monsoon, IGP, CI and NE show an increase in below inversion AOD as IS increases up to 2–3 K (Fig. S8(b)). The EC and WC show low and similar AODs for all the strength ranges during winter and pre-monsoon (Fig. S8(a) & (c)). The AOD above the inversion during winter is low and in pre-monsoon it is high in all the regions irrespective of the strength and depth ranges (Fig. S8(b) & (d)). The IGP region and NE shows the similar pattern in the below inversion AOD for different ID ranges during winter (Fig. S8(e)). Irrespective of ID range, large (small) variability in the AOD above the inversion is noticed during pre-monsoon (winter) (Fig. S8(f) & (h)). These results confirm that the inversion parameters (IS and ID) have large influence on the vertical distribution of aerosols and are in good parity with that observed over Gadanki.

5. Summary and conclusions

Our study examined the influence of atmospheric boundary layer (ABL) altitude and the strong temperature inversions on the vertical
distribution of aerosol using 9 years (2010–2018) of co-located ground-based Micro Pulse Lidar (MPL) and radiosonde observations over a tropical station Gadanki, India. We also make use of other co-located instrument measurements like Sky Radiometer, Aethalometer, nephelometer, Aerodynamic Particle Sizer, CRDS analyzer etc, from Gadanki to support the observed features. This study is extended covering complete India having different geographical locations using Cloud Aerosol Lidar Infrared Pathfinder Satellite Observations (CALIPSO) and India Meteorological Department (IMD) radiosonde network.

As expected, a good relation between ABL altitude and aerosol (and trace gases) at the surface is found with higher (lower) concentrations during night (day) times. This relation also holds good even at seasonal scales with lower (higher) concentrations during pre-monsoon to monsoon season (post-monsoon to winter). However, such relation is not seen in aerosol columnar properties like Aerosol Optical Depth (AOD) at both sub-daily and seasonal scales. No significant diurnal variation in AOD is found that obtained while combining MPL (night time) and Sky Radiometer (day time). The higher AOD is noticed during pre-monsoon to monsoon season though ABL altitude is also found high leaving behind the ABL altitude influence on the aerosol columnar properties. On several occasions, the observations show that aerosols are being trapped at the ABL (or SBL) altitude with sudden dip in aerosol extinction but a significant amount of aerosol still persists above the ABL altitude at both sub-daily and seasonal scales. Interestingly, we found that strong temperature inversions present in the lower troposphere act as capping inversion, where most of the aerosols are being trapped but not at the ABL altitude irrespective of the time and season. The strongest inversion coincides with the ABL altitude for about 51.7% of the time with highest percentage in winter (58.6%) followed by pre-monsoon (58.5%), post-monsoon (41.0%) and minimum in monsoon (32.5%) seasons. Rest of the time the strongest inversion occurred always well above the ABL altitude. Thus, more than 50% of the aerosol still persists above the ABL altitude except during winter and pre-monsoon. Though this percentage reduced drastically above the strongest inversion altitude but traceable quantity of aerosol still persists above except in winter. Surprisingly, trapping of aerosols increases with the increase in the inversion strength (and depth) but only up to a certain magnitude (3–4 K (300–400 m) during winter and 2–3 K (100–200 m) in pre-monsoon). Thereafter no influence is seen. Similar features are also seen spatially across India covering different geographical locations but with slight differences over the coastal stations.

Thus, main conclusions drawn from this study is that the strongest temperature inversion present in the lower troposphere is the deciding factor for the aerosol vertical distribution irrespective of the season rather than the ABL altitude which was usually thought as the capping inversion. ‘Aerosol concentrations below the inversion increase with the increasing inversion strength and depth but only up to a certain magnitude’ thereafter no influence is seen. The quantitative assessments of inversion statistics, the impact of inversion on vertical distribution of aerosols are important for modeling and predicting the diffusion of aerosols near the ground under different synoptic conditions. Moreover, accurate relationship between aerosols and inversions can provide reliable management strategies to control the pollution levels especially in highly polluted areas. A recent study, statistically showed that the temperature inversions are correlated with concentration of air pollutants and respiratory diseases (Trinh et al., 2019). Hence, these results can greatly benefit in understanding the aerosol/trace gases effects on the health, climate, atmospheric dynamics and air pollution control.

CRediT authorship contribution statement

P. Prasad: Methodology, Visualization, Validation, Data curation, Writing - Original draft preparation. Ghouse Basha: Methodology, Writing - Original draft preparation. M. Venkat Ratnam: Conceptualization, Visualization, Investigation, Supervision, Writing - Reviewing and editing.

Declaration of competing interest

Authors declare no financial interest.

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Appendix A. Supplementary data

Supplementary data to this article can be found online at https://doi.org/10.1016/j.scitotenv.2021.149758.

References


