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Influence of South Asian Dust Storm on Aerosol Radiative Forcing at High-Altitude Station in Central Himalayas

A. K. Srivastava1*, P. Pant1, P. Hegde1, Sachchidanand Singh2, U. C. Dumka1, M. Naja1, Narendra Singh1 and Y. Bhavanikumar3

1Aryabhatta Research Institute of Observational Sciences (ARIES), Nainital, India
2Radio and Atmospheric Sciences Division, National Physical Laboratory (NPL), New Delhi, India
3National Atmospheric Research Laboratory (NARL), Gadanki (India)

(*Corresponding author: atul@tropmet.res.in)

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**Keywords:** Aerosol; Dust storm; Back-scattering ratio; Aerosol optical depth; Radiative forcing

# Present Affiliation of A. K. Srivastava:

2
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Indian Institute of Tropical Meteorology, Pune (New Delhi Branch), New Rajendra Nagar, New Delhi, India

1. Introduction

The impact of dust storms on immediate environment and climate has received a considerable importance in recent past (Liu et al., 2004; IPCC, 2007 and references therein). Dust in the atmosphere have terrestrial sources and represent an important process of land-atmosphere interaction (Tegen et al., 1996). In the northern hemisphere, dust particles play a key role in modulating climate through the indirect aerosol effects that alter the cloud properties (Sassen et al., 2003). As far as the dust storms in south Asia are concerned, they are common in northwest India, especially in the western Rajasthan region, which is mainly covered by the Thar and Margo Deserts (Dey et al., 2004; Prasad et al., 2007; Pandithurai et al., 2008) and are the single largest contributor to the mineral dust over the northwest Indian regions (Todd et al., 2007). Although the dust particles directly and indirectly affect the climate by scattering and absorbing solar radiations, nevertheless their role in climate forcing is poorly understood (Kaufman et al., 2002; Satheesh and Moorthy, 2005). However, radiative properties of dust aerosols are essential to access their impact on climate.

Satellite and ground based observations have shown that the global sources of atmospheric dusts are arid and semi-arid desert regions contributing to the long-range transport of dust particles during dust events (Husar et al., 1997; Washington et al., 2003). Over the years, numerous studies have been made on the characterization and impact assessment of transported dust aerosols during dust events at different low-altitude stations (Dey et al., 2004; Park et al., 2005; Singh et al., 2005; Nee et al., 2007; Pandithurai et al., 2008). However, most of these studies focused mainly on urban and/or semi-urban regions. There are very few studies related to the characteristics and impacts of dust aerosols on radiative forcing at the high-altitude stations, which generally has pristine environment and
away from these dust impacts. Moreover, these investigations are considerably important in the context of recent theory of the “Elevated Heat Pump” as proposed by Lau et al. (2006). Recently, Hegde et al. (2007) have found clear evidences of the dust storm impact on the measured surface aerosol parameters at a high-altitude station, Nainital.

The present study is focused on a dust storm that primarily originated in the Thar Desert area during second week of June 2006 and reached over the high-altitude station of Manora Peak, Nainital at an altitude of about 1958 m above mean sea level (msl). An elevated dust aerosol layer, associated with an enhanced backscatter signal, was observed for the first time over the station using ground based micro-pulse lidar system. Apart from the lidar observation of aerosol dust layer, focus of the present study is also on the changes in aerosol radiative forcing due to enhanced dust concentration over this station in central Himalayas.

2. Site Description and Measurement Techniques

Observations were carried out from Manora Peak (29.4°N, 79.5°E, ~1958 m above msl), Nainital, located in the central Himalayas. The north and northeast of Manora Peak has sharply undulating topography of Himalayan mountain ranges; however, the southwest side has low elevated plain land merging into the Ganga Basin. Owing to its large elevation associated with a very low atmospheric boundary layer height condition, this site may be considered to be as a free tropospheric site. As this high altitude site is devoid of any major pollution (Sagar et al., 2004; Pant et al., 2006), the investigation from such a remote, sparsely inhabited regions have the importance of providing a sort of background level against which the impact of aerosol loading during the dust episodes can be assessed. Since the long range transport of dust make significant contribution to the atmospheric aerosols even at locations far from their sources, the impact of the enhanced level of aerosol concentration during the dust event over this experimental site can be distinctly marked in the vertical profiles of aerosol backscatter intensity of the lidar measurements.
A portable lidar system was installed at Manora Peak, Nainital in May 2006 for the study of vertical profiles of aerosols and clouds. This system is designed and developed by National Atmospheric Research Laboratory (NARL) Gadanki (Bhavanikumar, 2006), which is based on micro-pulsed lidar (MPL) technology (Spinhirne, 1993). The lidar employs a diode pumped Nd:YAG laser with second harmonic output at 532 nm and operated at a pulse repetition frequency of 2500 Hz. The emitter beam is co-axial to receiver field-of-view (FOV) and operated in the zenith direction. The lidar receiver employs a 150 mm Cassegrain telescope. The geometrical form factor for such a co-axial lidar system having no apertures (other than the objective lens or mirror of the telescope) or obstruction, is unity provided the divergence angle of the laser beam is less than the opening angle of the telescope (Measures, 1984). An iris (pinhole) of diameter 0.5 mm was used to obtain the receiver FOV of about 400 µ rad. A narrow band interference filter (IF) was positioned in front of the photon multiplier tube (PMT). A high gain PMT (Hamamatsu R3234), operating in photon counting mode, has been used at the receiver end to detect the backscattered photon counts. A computer based multi-channel analyzer (EG&G Ortec model MCA-pci) has been employed for recording these photon returns in to the PC. The backscattered signals are measured with a bin width of 200 ns, which corresponds to an altitude resolution of 30 m. Usually 300000 laser shots are integrated for a single raw data profile, which corresponds to a time resolution of 120 sec.

The lidar profiles are based on the solution of the lidar equation calculated for the case of single scattering (Measures, 1984) as:

\[ P_r(z) = P_t K \left( \frac{\lambda}{z^2} \right) c \tau z \beta(z) \exp\left[ -2 \int \sigma(z) dz \right] \]

where \( P_r(z) \) is the power received by lidar system from a scattering volume at an altitude \( z \). \( P_t \) is the power transmitted from lidar system and \( K \) is the system constant (includes...
transmitting and receiving optics losses and effective receiver aperture). $\frac{A}{z^2}$ is the solid angle subtended by the primary mirror at the range $z$ where $A$ is the telescope receiving area. $\frac{c\tau}{2}$ is the range resolution of the system where $c$ is the velocity of light and $\tau$ is the pulse duration of the laser beam. $\beta(z)$ is the volume back-scattering coefficient. It gives the fractional amount of the incident energy scattered per steradian (sr) in the backward direction per unit atmospheric path length and has the dimension of $m^{-1}\text{sr}^{-1}$. $\sigma(z)$ is the volume extinction coefficient of the atmosphere that has the unit of $m^{-1}$, defined as twice the integral between the transmitter and the scattering volume to obtain the net transmission ($T$).

The terms volume back-scattering and extinction coefficients, $\beta(z)$ and $\sigma(z)$, contain the contribution from both air molecules and aerosols, expressed as:

$$\beta(z) = \beta_m(z) + \beta_a(z) \quad \text{and} \quad \sigma(z) = \sigma_m(z) + \sigma_a(z)$$

(2)

where subscripts $m$ and $a$ represent air molecules and aerosols, respectively. The molecular (or Rayleigh) contribution to the signal is taken from the CIRA-1986 standard atmospheric model. The problem related with the lidar equation is that it contains two unknowns, $\beta(z)$ and $\sigma(z)$, which make it difficult to obtain the general solution. Appropriate inversion methods (Klett, 1985; Fernald, 1984) have been developed to solve the above equation (1). For the inversion of lidar signal, a reference altitude is to be assumed where aerosol contribution is insignificant and the total volume back-scattering coefficient will be equal to the molecular backscattering coefficient (i.e. $\beta \approx \beta_m$). The reference altitude is taken as 6 km in the present study, as most of the aerosols in the tropics are located below 6 km (Gadhavi and Jayaraman, 2006). The received backscatter signal can be expressed in terms of back-scattering ratio (BSR), which is the relative magnitude between the molecular (or Rayleigh) and aerosol (Mie) scattering (Nee et al., 2007) and given as
Further, aerosol extinction coefficient ($\sigma_a(z)$) was estimated by assuming appropriate value of lidar ratio (LR), which is the ratio of extinction-to-backscattering coefficient. LR value is considered to be 35 sr for the environment like Nainital on the basis of literature review (Muller et al., 2001; He et al., 2006 and references therein). Based on the above parameters, optical thickness of aerosol or dust particles (i.e. AOD) can be derived by integrating aerosol extinction coefficient within the altitude $z_1$ and $z_2$, expressed as

$$\text{Lidar}_{(AOD)} = \int_{z_1}^{z_2} \sigma_a(z) \, dz = \int_{z_1}^{z_2} \beta_a(z) \, LR \, dz$$

Besides the lidar observations in the nighttime, direct measurements of columnar AOD during the daytime, at different wavelengths from UV to near-IR region were also carried out at the station using Microtops-II Sunphotometer and Ozonometer (Solar Light Co., USA). The details of the instrument are available elsewhere (Morys et al., 2001). Measurements of surface BC mass concentrations were carried out using Aethalometer (Model AE-42, Magee Scientific Company, USA). More details about the instrument and the principle of measurements are available elsewhere (Hansen, 2003). Some of the observations of the optical and physical properties of aerosols over the station using these instruments have been reported by Hegde et al. (2007) during dust episode.

3. Results and Discussion

3.1. Identification of Dust Episode

The Moderate Resolution Imaging Spectroradiometer (MODIS) flying onboard Aqua satellite captured a large dust storm along the border between India and Pakistan on 12 June 2006 (http://earthobservatory.nasa.gov/NaturalHazards/Archive) originated over Thar Desert and blew through the Indus valley, as shown in Figure 1. The experimental site, Nainital has
been indicated by a star in the figure. It shows the spatial distribution of dust storm, which reveals that the plumes of dust aerosols engulf the entire northern part of India. It is also obvious from the figure that the dust plumes are headed toward the Himalayan Mountains and reached up to the high-altitude experimental station (top right corner of the image). Whereas, in the lower left corner, sprays of clouds also appear to blow in the northeast direction as that of the dust. Once the dust reaches the Himalayan Mountains, it changes its direction and blows along its southern edge. The manifestation of dust events can be discernible as an enhancement in the aerosol index as well as spectral aerosol optical depths.

3.2. Transport of Dust at Nainital

The prevailing meteorology during pre-monsoon season at Nainital is generally characterized by northwesterly winds, which passes through arid regions of the western India and mostly brings dry airmasses from the southwest Asia. Since the impact of ambient dust on radiation budget of the atmosphere changes with source regions and transport pathways, hence it is prerequisite to get the above information for their impact assessment. In order to know the transport pathways of dust aerosols due to dust activities primarily originated over the Thar Desert region, analysis of backward airmass trajectories and Aerosol Index (AI) were carried out over the experimental station from 08-15 June 2006. AI images obtained from Ozone Monitoring Instrument (OMI) aboard on Aura satellite, show regional distribution of aerosols, which indicates dispersion of dust aerosols over the Thar region, prominently on 12 June 2006 (Figure 2), having maximum AI ~4.0. From these index images, it is quite discernible that dust activity started on 11 June 2006 and engulfed most part of the northwest India in the following days. The maximum value of AI (~2.5) was observed over the experimental site on 12 June 2006. Recently, Papayannis et al. (2007) have found AI values over China ranged between 2.0 and 3.5 during the dust events, which exceeded up to 4.5 during the intense dust events. The high values of AI attribute to the
presence of absorbing aerosol particles over the region. Further, to substantiate the pathways for dust transport, 5-day airmass back-trajectories, employing the Hybrid Single Particle Lagrangian Integrated Trajectory (HYSPLIT) model, were plotted and superimposed on AI images (Figure 2). The backward trajectories were analyzed for three different time intervals, at 06:00 GMT (cyan color line), 12:00 GMT (red color line) and 18:00 GMT (green color line) for an altitude of 1500 m above ground level (AGL). The back-trajectories show different pathways for the transport of airmasses from the source region to the experimental site during different time periods.

3.3. Lidar Measurements

Lidar measurements were carried out on regular basis at Manora Peak, Nainital during nighttime (to avoid more background noise in the daytime). The impact of aerosol loading during dust episode (primarily originated over the Thar Desert) over the experimental site was clearly observed in lidar profiles which show a significant enhancement in backscattered signals on 12 June 2006 in terms of back-scattering ratio (BSR) as shown in Figure 3 (open circle). The figure shows vertical profiles of BSR derived from lidar system using equation (3), which depends on the variations in aerosol layer strength, its height and number concentration of aerosol particles. To distinguish the existence of dust layer from the clear-sky conditions, vertical profile of BSR on 08 June 2006 (i.e. pre-dust day, which is considered as clear-sky day) has also been plotted in Figure 3 (solid circle). The horizontal lines indicate the standard deviation of BSR at each altitude level. It is quite obvious from the figure that BSR values at each altitude are significantly larger on 12 June 2006 (dust day) than that on 08 June 2006 (pre-dust day), particularly in the altitude range of 1000 to 3000 m AGL. This is attributed mainly due to the presence of elevated layer of dust aerosols, transported from far-off Thar Desert region to the experimental site. The peak BSR value on dust day was found to be ~27 at slightly higher altitude at ~1300 m AGL, which extended up
to the altitude of ~3000 m AGL whereas the peak BSR on pre-dust day was found to be only ~18 at an altitude ~1200 m AGL. The value of BSR = 1 above and below the aerosol layer, fairly suggests that it will never be zero even in the case when there are no aerosols in the atmosphere as per the equation 3, which represents the molecular atmosphere. Recently, Hegde et al. (2009) have studied vertical profiles of aerosol over Nainital by analyzing aerosol backscatter profiles derived from micro-pulse lidar system and found significantly large BSR values, peaking at an altitude around 1500 m AGL, during pre-monsoon months, which is highly associated with the prevailing strong convection conditions.

Temporal evolution of airmasses before 120 hours from the date of dust observation over the station ending at the altitude of 1500 m AGL (approximate altitude of observed peak BSR) have been analyzed from HYSPLIT model and shown in Figure 4. Significant advancements of airmasses were observed near to the altitude of peak BSR at different time intervals (as mentioned in Figure 2) on 12 and 13 June 2006. It shows constancy in the vertically upward trend in the transport of airmasses, thereby facilitates the upward transport of dust from the source region to the experimental site.

Vertical profiles of aerosol in terms of backscattered signals, having temporal resolution of 120 sec were obtained by lidar system every day to study the height versus time variations in BSR. The temporal variation of BSR with height on 12 June 2006 (Figure 5) shows vertical distribution of dust aerosol layer with its strength, height and the number concentrations. The dust layer peaks at an altitude ~1300 m AGL having maximum BSR value ~27, which extends up to an altitude ~3000 m AGL. The corresponding aerosol extinction profiles have been integrated within the altitude region from ~200 m AGL (lowest altitude detected by lidar system) up to ~4500 m AGL to get the AOD (at 532 nm) using equation (4). The temporal variation of lidar-derived AOD is plotted over the same plot (continuous solid line, Figure 5). By and large BSR and AOD seems to be perfectly
complimentary to each other as far as their temporal variation is concerned. The lidar-derived AOD (at 532 nm) values vary from ~0.6 to 1.2 with an average value ~0.83 (±0.12). Recently, Nee et al. (2007) have investigated long-range transport of Asian dust storm using a polarization lidar system at Chungli (24.6°N, 121.1°E). They have observed an average dust layer height ~1.6 km having an average BSR value ~12 (±16.75) and AOD ~0.54 (±0.31). However, Chen et al. (2007) have reported Raman and depolarization lidar measurements of optical properties of Asian dust in the free troposphere during dust seasons over Taipei, Taiwan (25.14°N, 121.54°E). They have observed dust layers within the altitude ranging from 1-6 km, having optical thickness ranged from 0.01 to 0.55.

3.4. Effect of Dust on Surface Aerosol Properties

The effect of dust on the surface aerosol properties at Nainital for the whole month of June 2006 has been described by Hegde et al. (2007). They have found five to ten times increase in the coarse and giant mode particles during dust day as compared to their respective monthly mean using optical particle counter and two to four times increase in the spectral AOD values to its normal day-mean. Further, they have also found large enhancement in total suspended particulate matter concentration (~280 µg m⁻³) during dust episode. For the purpose of radiative forcing estimations in the present analysis, the spectral variations of AOD, measured by Microtops-II, on dust day (13 June 2006) and pre-dust day (08 June 2006) as shown by vertical columns in Figure 6a and 6b, respectively, have been used. More than two folds increase in AOD values were observed at all the wavelengths on dust day as compared to that on pre-dust day. Mean AOD at 500 nm was observed to be as high as ~0.63 during dust day and ~0.23 on pre-dust day. The value of AOD of the order of 0.63 is very high for a pristine site like Nainital though it may be significantly low as compared to the values observed at other low-altitude stations in India during dust events (Dey et al., 2004; Singh et al., 2005; Pandithurai et al., 2008).
An increasing trend of AOD with respect to wavelengths observed on dust day, clearly indicates the presence of coarser dust particles over the station. Similar trend in AOD has also been reported by Singh et al. (2005) during a dust event over New Delhi. On the other hand, the AOD observed on pre-dust day follow the normal behavior, i.e., AOD decreases with the increase in wavelength as per Angstrom relation given below. Ångström exponent ($\alpha$) is a good indicator of the size of particles, which was computed from the spectral variation of AOD in the wavelength range from 380 to 1020 nm using the Ångström (1964) relationship:

$$\tau_a = \beta \lambda^{-\alpha}$$

where $\tau_a$ is the aerosol optical depth, $\beta$ is the turbidity coefficient, indicating the aerosol loading and $\lambda$ is the wavelength.

During dust day, the value of $\alpha$ was observed to be ~0.04 with corresponding value of $\beta$ ~0.65; however, significantly large value of $\alpha$ (~0.42) was observed during pre-dust day whereas the value of $\beta$ was observed to be significantly low (~0.18). Higher values of $\alpha$ are typically observed for accumulation or fine mode particles (Reid et al., 1999; Eck et al., 1999) whereas the lower values are observed for coarse mode particles such as Saharan dusts and Asian dusts (Eck et al., 1999; Sakai et al., 2002). Moreover, the high $\beta$ value during dust day compared to that on the pre dust day (which is more than 3.6 times) clearly indicates a heavy aerosol loading on dust day. Unavailability of chemical data over Nainital makes it difficult to comment on the source of dust aerosols in terms of minerology. Nevertheless, the analysis of back-trajectories and AI images during the dust storms, strongly support the concept of the transport of dust aerosols from the western Thar Desert to the investigating site at Nainital (Figure 2). Moreover, the absence of enhancement in black carbon and accumulation mode particles over the site suggests a negligible change as far as the influences of the anthropogenic activities on the aerosol properties during the study period are concerned (Hegde et al, 2007).
3.5. Radiative Effect of Dust Aerosols

It is already known that the dust aerosols scatter as well as absorb solar radiation in the short-wave and also absorb and emit outgoing long-wave radiation (Kaufman, et al., 2001). Some of these dust aerosols are highly absorbing component of the atmosphere, which are significant contributors to the radiative warming in lower atmosphere due to short-wave absorption (Alpert et al., 1998). The forcing at top-of-atmosphere (TOA) and at the surface is defined as the difference in net fluxes (down minus up) with and without aerosols at their respective levels (Moorthy et al., 2005). In the present analysis, net flux has been estimated in the short-wave region (0.25-4.0 µm) with and without aerosols, separately at TOA and at the surface using the Santa Barbara DISORT Atmospheric Radiative Transfer (SBDART) model (Ricchiazzi et al., 1998). SBDART is a plane-parallel radiative transfer model based on discrete ordinate approach to estimate aerosol radiative forcing in clear and cloudy conditions. The radiative transfer equations are numerically integrated with Discrete Ordinate Radiative Transfer (DISORT) module (Stamnes et al., 1988) in the SBDART. This method uses numerically stable algorithm to solve the equations of plane-parallel radiative transfer in vertically inhomogeneous atmosphere (Ricchiazzi et al., 1998). All important processes that affect the ultraviolet, visible and infrared radiation fields are incorporated in the SBDART.

The main input parameters required in the SBDART for estimating aerosol forcing are aerosol optical depth (AOD), single scattering albedo (SSA) and asymmetry parameter. Among these input parameters, SSA plays a key role in forcing estimations (Satheesh and Moorthy, 2005), which indicates relative contribution of scattering and absorption to the total extinction by aerosol particles. The scattering phase function required for radiative transfer computations is generated from the asymmetry parameter using the Heyney-Greenstein approximation (Dey and Tripathi, 2007). Besides the above optical parameters, the other important input parameters in SBDART model are solar geometry, a model atmosphere and
surface albedo. Since the surface albedo or reflection is one of the key parameter in forcing estimation, Satheesh (2002) has incorporated different kind of surface reflection in his detailed analysis of aerosol forcing. He has suggested that due to the high surface albedo (e.g. soil); aerosol forcing at TOA will change its sign from negative to positive. Recently, Pandithurai et al. (2008) have reported positive TOA forcing at New Delhi during pre-monsoon month, which is mainly attributed to the lower SSA and higher surface albedo values. For estimating the radiative forcing over Nainital, surface albedo has been considered to be as vegetation type among all the existing surface types in SBDART.

As there are no direct measurements of SSA and asymmetry parameter over the station, measured spectral AOD (using Microtops-II) and BC mass concentration (using Aethalometer) values were used to constrain the modeled SSA and asymmetry parameter using optical properties of aerosol and cloud (OPAC) model given by Hess et al. (1998). To distinguish aerosol characteristics during dust and clear-sky days, measured spectral AOD and BC mass concentration values were chosen on dust (13 June 2006) and clear-sky (i.e. pre-dust day on 08 June 2006) days and followed the approach given by Moorthy et al. (2005) of anchoring point by defining new mixtures in OPAC model. The model is then fine-tuned to match the measured AOD at different wavelengths. The components of the mixture used in the OPAC model for dust day are comprised of water-insoluble, water-soluble, transported mineral dust and soot. For clear-sky day the components used were water-insoluble, water-soluble, mineral (nuclei mode), mineral (accumulation mode) and soot. The value of soot has been taken from the actual measurements of BC during these two days. The number density of each of these species was adjusted while maintaining the mass fraction of BC to the observed one so that the modeled spectral AOD values (assuming spherical and externally mixed type aerosols) agreed well within 5% to those of observed mean AOD values. Because of this closure (with spectral AOD) and anchoring the BC mass fraction, the
The initial assumption of the model will not have any significant impact on the estimated forcing, as has already been pointed out by Satheesh and Srinivasan (2005).

The estimated AOD values from aerosol model are found to agree well with the measured AOD values at all the wavelengths as shown by dashed line with open circles in Figure 6a and 6b on dust (13 June 2006) and pre-dust (08 June 2006) days, respectively. The model estimated AOD values are well within the standard deviations of measured AOD values at each wavelength. The corresponding values of SSA and asymmetry parameter, estimated using OPAC model, were then used in SBDART model for forcing calculations. Significantly low value of SSA was observed on both pre-dust and dust days. The SSA (500 nm) was found to be \( \sim 0.69 \) during pre-dust day, which indicates significant aerosol absorption. On the other hand, a small increase in SSA \( \sim 0.1 \) was observed during dust day, which is due to the abundant dust loading in the region as clearly indicated by AOD values (Figure 6), attributing to the scattering state of the atmosphere with the presence of absorbing nature of dust aerosols. The observed low SSA on pre-dust day could be due to the presence of large BC mass concentrations \( \sim 1.95 \, \mu g \, m^{-3} \) over the station, which was found to be more than twice the value observed on dust day \( \sim 0.8 \, \mu g \, m^{-3} \). However, high SSA on dust day could be due to non-sphericity of dust mix with anthropogenic aerosol species during long-range transport over the polluted regions, which suggest their radiative importance over the station. Results are found to be in good agreement with the values variously observed by Chinnam et al. (2006); Costa et al. (2006); Pandithurai et al. (2008) during the dust events at different locations. Ramana et al. (2004) have observed the SSA value (500 nm) in the range of 0.70–0.90 in the Himalayan region (Kathmandu, Nepal) by combining the actual measurements of scattering and absorption coefficient using a nephelometer and a particle soot absorption photometer (PSAP). On the other hand, Pant et al. (2006) have estimated the
SSA value at Nainital in the range of 0.87 to 0.94 with a mean value of 0.90 during winter month.

In order to obtain the diurnally averaged forcing at TOA and at the surface, radiative forcing were estimated for the conditions of solar zenith angle (SZA) at every $5^\circ$ interval, following the procedure given by Dey and Tripathi (2007). The difference between the TOA and surface forcing is considered as atmospheric forcing ($\Delta F$), which represents the amount of energy trapped by aerosols within the atmosphere and gets transformed into heat. The atmospheric heating rate due to aerosol absorption is given by

$$\frac{\partial T}{\partial t} = \frac{g}{C_p} \frac{\Delta F}{\Delta P}$$  \hspace{1cm} (6)

where $\partial T/\partial t$ is the heating rate (K day$^{-1}$), $C_p$ is the specific heat capacity of air at constant pressure, $g$ is the acceleration due to gravity and $\Delta P$ is the atmospheric pressure difference between top and bottom surfaces of each layer.

Based on the measured parameters and the prevailing atmospheric conditions over the station, the mid-latitude summer model atmospheric profile has been considered in SBDART model. The measured values of columnar water vapor and total column ozone, retrieved by Microtops-II have also been used for model estimation, which was found to be ~0.68 cm and ~358 DU on dust day (13 June 2006) and ~0.83 cm and ~344 DU on pre-dust day (08 June 2006). The estimated values of aerosol forcing at TOA, surface and in the atmosphere on dust and pre-dust days are shown in Figure 7. The heating rates on the corresponding days are shown in the parentheses. During dust day, TOA and surface forcing values were found to be as high as about -30 and -45 W m$^{-2}$, respectively implying a net cooling effect. However, the resultant atmospheric forcing about +15 W m$^{-2}$ was absorbed into the atmosphere due to presence of mineral dust and soot aerosols. Consequently, the short-wave atmospheric absorption causes heating of the lower atmosphere by ~0.4 K day$^{-1}$. Forcing due to mineral
dust aerosols was observed to be higher than the pre-dust day by about 2, 3 and 5 times at the TOA, surface and in the atmosphere, respectively. The corresponding atmospheric heating rate was observed to be ~0.1 K day$^{-1}$ on pre-dust day, which is one forth as compared to that on dust day. Since the aerosol optical depth is considered to be an index for total columnar burden of aerosols in the atmosphere, therefore the results obtained on two different days are highly associated with the spectral behavior of the AODs. In the earlier studies, Ramana et al. (2004) have observed a surface radiative forcing of about -25 W m$^{-2}$ in the Himalayan region at Kathmandu, Nepal. However, Pant et al. (2006) have estimated the aerosol forcing at Nainital during winter month (December 2004) as -4.2, +0.7 and +4.9 W m$^{-2}$ at TOA, surface and in the atmosphere, respectively. Recently, Prasad et al. (2007) have found a change in average surface and TOA forcing by about -23 and -11 W m$^{-2}$, respectively during dust event over Kanpur in the IG plains. However, Pandithurai et al. (2008) estimated the values of surface and TOA forcing ranging from -39 to -106 and -12 to +24 W m$^{-2}$, respectively at New Delhi during the pre-monsoon season, which exerts large heating rate ranging from 0.6 to 2.5 K day$^{-1}$.

The observation of such an enhanced dust layer over the high-altitude region of central Himalayas during pre-monsoon or late spring can play a crucial role in heating the atmosphere, as suggested by Lau et al. (2006). The effect can be more pronounced due to the enhanced concentration of BC getting mixed with the Desert dust aerosols causing more heating at elevated heights. In the present analysis the atmospheric heating of the order of 0.4 K day$^{-1}$ is only due to the short-wave (0.25-4.0 µm) absorption by the dust aerosol particles. Nonetheless there may be a significant contribution in the atmospheric heating due to the long-wave absorption by these dust particles as well. Such an elevated heating may lead to an advance of the rainy periods and subsequently an intensification of the Indian summer monsoon as suggested by Lau et al. (2006).
4. Conclusions

Impact of long-range transport of south Asian dust storms on optical and radiative properties of dust aerosols has been investigated, for the first time, over the high-altitude station in central Himalayas. The salient features of the present study are:

- The BSR values were observed to be significantly high at each altitude level on dust day as compared to that on pre-dust day, having a peak enhancement of ~27 at an altitude ~1300 m AGL and extends up to the altitude ~3000 m AGL, which is mainly attributed to the presence of large amount of dust aerosols over the Himalayan region, transported from the associated intense dust episode in the Thar Desert.

- Both Aerosol Index images and airmass back-trajectories strongly support the origin and transport of dust aerosols from the western Thar Desert region towards the experimental site.

- Microtops-II measured mean AOD (500 nm) was observed to be as high as ~0.63 and 0.23 on dust and pre-dust days, respectively. Inferred Angstrom exponent (α), using the observed spectral AODs, is found to be ~0.04 and 0.42, on dust and pre-dust days, respectively. These values of α clearly indicate the dominance of coarse mode particles on dust day and accumulation or fine mode particles on pre-dust day.

- The presence of mineral dust over the station decreases the short-wave radiations reaching to the Earth’s surface. Thus, a net negative forcing of ~30 and 45 W m\(^{-2}\) was observed at the TOA and surface, respectively which implies a net cooling effect. However, a net positive forcing of ~15 W m\(^{-2}\) was observed in the atmosphere, which implies a net warming effect and causes heating of the lower atmosphere by ~0.4 K day\(^{-1}\).

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Figure Captions

Figure 1. Spatial distribution of dust storm observed by MODIS on 12 June 2006.

Figure 2. OMI Aerosol Index (AI) images showing the source and the progressive movement of the dust aerosols before and after the dust storm activity in the month of June 2006 (5-day airmass back-trajectories at 06 GMT (cyan colour line), 12 GMT (red colour line) and 18 GMT (green colour line) at an altitude of 1500 m AGL are superimposed on respective days of AI image).

Figure 3. Vertical profiles of BSR on dust (12 June 2006) and pre-dust (08 June 2006) days.

Figure 4. Temporal evolution of airmasses at 1500 m AGL for three different time intervals on 12 and 13 June 2006.

Figure 5. Height versus time variations in BSR on dust day (12 June 2006) along with the variations in lidar-derived AOD (532 nm).

Figure 6. Spectral variations of Microtops-II measured AOD values on (a) dust day (13 June 2006) and (b) pre-dust day (08 June 2006) shown by vertical columns. Model estimated AOD values are shown by dashed line with open circles.

Figure 7. Short-wave radiative forcing (in W m$^{-2}$) at the TOA, surface and in the atmosphere on dust (13 June 2006) and pre-dust (08 June 2006) days. Atmospheric Heating rates (in K day$^{-1}$) on corresponding days are given in the parentheses.
Figure 5

Average AOD = 0.831 (std=0.12)

12 June, 2006

BSR

24
21
18
15
12
9
6
3
0

Height (m, AGL)

20:00  21:00  22:00  23:30

Local Time (Hr)