Investigation on Seasonal Variations of Aerosol Properties and its Influence on Radiative Forcing over an Urban Location in Central India

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Abstract:

Aerosol plays an important role in modulating solar radiation, which are of great concern in perspective of regional climate change. The study analysed the physical and optical properties of aerosols over an urban area and estimated radiative effect using three years in-situ data from sunphotometer, aethalometer and nephelometer as input to radiative transfer model. Aerosols properties indicate the dominance of fine mode aerosols over the study area. However presence of coarse mode aerosols is also found during pre-monsoon [March-April-May]. Daily mean aerosol optical depth showed a minimum during winter [Dec-Jan-Feb] (0.45-0.52) and a maximum during pre-monsoon(0.6-0.7), while single scattering albedo (ω) attains its maximum (0.78±0.05) in winter and minimum (0.67±0.06)during pre-monsoon and asymmetry factor varied in the range between 0.48±0.02 to 0.53±0.04. Episodic events of dust storm and biomass burning are identified by analyzing intrinsic aerosol optical properties like scattering Ångström exponent (SAE) and absorption Ångström exponent (AAE) during the study periods and it has been observed that during dust storm events ω is lower (~0.77) than that of during biomass burning(~0.81). The aerosol direct radiative effect at top of the atmosphere during winter is -11.72 ±3.5 Wm^{-2}, while during pre-monsoon; it is -5.5 ±2.5 Wm^{-2}, which can be due to observed lower values of ω during pre-monsoon. A large positive enhancement of atmospheric effect of

~50.53Wm^{-2} is observed during pre-monsoon compared to winter. Due to high aerosol loading in pre-monsoon, a twofold negative surface forcing is also observed in comparison to winter.

Keywords: Aerosol Optical Depth, Single Scattering Albedo and aerosol direct radiative effect.

1. Introduction

Atmospheric aerosols and their impact on climate have gained considerable attention from scientific community and policy makers during past recent years [IPCC, 2013]. Aerosols effects on climate are mainly classified into two viz. aerosol direct radiative effect (ADRE) and indirect radiative effect. ADRE [Charlson et al., 1992] is associated with scattering or absorbing of incoming solar radiation by aerosols thereby, producinga negative or positive radiative effect at Top of the Atmosphere (TOA). Scattering aerosols like sea salt, sulphate etc. induce climate cooling, while absorbing aerosols like black carbon (BC) induce a warming effect at TOA [Bond et al., 2013]. Indirect radiative effect is the mechanism by which aerosols alter microphysical and radiative properties of clouds, thereby influencing albedo, lifetime and precipitation of clouds [Ramanathan et al., 2001]. Jacobson [2001] has pointed that radiative properties of aerosols depends also on its mixing state, that is the degree to which chemical components occur as independent particles (external mixing) as compared to a component mixture in each individual particle (internal mixing). Mixing state of aerosols will be extremely important while considering impact due to long range transported aerosols [Chinnam et al., 2006]. Global estimation of ADRE is found to be varied from -0.85 to +0.15 Wm^{-2}with an uncertainty of 1 Wm^{-2} [IPCC 2013].Uncertainty involved in ADRE estimation is due to ambiguities associated with sources, lifetime, distribution etc. [Anderson et al., 2003].

Uncertainty associated with ADRE can be minimized to some extend by their accurate characterization [Quinn et al., 1998]. Towards this, many international and national efforts have initiated like Aerosol Robotic Network (AERONET) [Holben et al., 1998], SKYNET [Takamura and Nakajima, 2004], Aerosol Radiative Forcing Over India Network (ARFI Net) [Moorthy et al., 2009] etc. to list a few. In addition to this, past decade have viewed a number a campaigns for regional aerosol characterization over water bodies like INDOEX [Ramanathan et al., 2001]; ACE-1 [Bates et al., 1998] and also over distinct land masses like ACE-Asia [Huebert et al., 2003]; SAFARI-2000 [King et al., 2003] etc. These efforts have brought out diverse physical and optical properties of varied aerosol species, which in turn has helped to minimize their uncertainty in determining their radiative effect.

Because of their high spatio-temporal heterogeneity it is speculated that aerosols may be more capable of altering atmospheric and oceanic circulation, especially on regional scale, than greenhouse gases [Ming and Ramaswamy., 2011]. Present study aims at better estimation of ADRE over an urban location, Hyderabad in Central India using in-situ data of 2010-2012 as input to radiative transfer model. Novelty of this study is that critical aerosol optical parameters (viz. Aerosol Optical Depth (τ or AOD), Single Scattering Albedo (ω or SSA) and asymmetry factor (g) etc.) required for estimation of ADRE are either measured or derived from *in-situ* observations.

Section 2 of this paper describe about study area, data sets and methodology. Sensitivity analysis of estimated ADRE is discussed in section 3. In Section 4, seasonal variation of aerosol radiative properties and associated radaitive effect are discussed. Further aerosol episodic events (viz. dust storm and biomass burning) which had impact on local aerosol system are identified and analyzed. Conclusions are presented in Section 5.

2. Study area, data set and methodology

2.1 Study area

Hyderabad is the capital of newly formed state Telangana in Central India, with a population of more than 4 million (http://censusindia.gov.in). Altitude of the study site is ~557 m above mean sea level. It has a hot semi-arid steppe climate with four dominant seasons; winter (Dec-Jan-Feb), pre-monsoon (Mar-Apr-May), monsoon (Jun-Jul-Aug-Sep) and post-monsoon (Oct-Nov). Meteorological parameters recorded at India Meteorological Department (IMD) (www.imd.gov.in), Begumpet, Hyderabad over a period of 1997 – 2012 is analysed. Study revealed that long term daily mean of maximum (T_{max}=36.1^oC) and minimum temperatures (T_{min}=17.54^oC) are similar to that was observed during the study period of 2010-2012 (T_{max}=36.0^oC and T_{min} 17.33^oC). However, analysis of diurnal variation of temperature recorded during study period shows that T_{max} and T_{min} are ~42^oC and ~16^oC, respectively. Long term (1997-2012) mean relative humidity varies between 40 – 80% with lowest observed during pre-monsoon and maximum during monsoon. Local wind speed varies typically from 5 – 15ms^{-1}, with a maximum observed during monsoon. Long term (1957-2012) annual mean of rainfall over Hyderabad is ~827mm, while annual rainfall during study period 2010–2012 is found to be 1192 mm, 625 mm and778 mm, respectively. Aerosol measurements are carried out at the campus of National Remote Sensing Centre (NRSC) (17.47^0 N, 78.43^0 E) located in heart of the city. The sampling site is near to state high way and main sources of aerosols are from vehicular emission, local/long range transport of dust aerosols, biomass burning and industrial emissions [Jose et al., 2015b].

2.2 Data set and methodology

2.2.1 Aerosol optical depth

Microtops II (Solar lights, USA) is a hand held sun photometer that measures spectral aerosol optical depth τ at five narrow spectral bands centered at 380, 440,500,675 and 870nm with a wavelength precession of $\pm$1.5nm and a full width at half maximum (FWHM) band pass of 10nm. Details about design, calibration and performance of Microtops are detailed in Morys et al.[2001]. It makes use of Beer-Lamberts-Bouguer law for estimation of τ. In general uncertainties associated with 'τ' measurements by Microtops is <0.02 for lower wavelength bands and <0.01 for higher bands [Porter et al., 2001]. In the present study, data were collected only during clear days with a sampling frequency of 30 min.

Spectral dependence of τ can be expressed using well-known Ångströms empirical relation [Ångström, 1964,] where, τ_λ is the aerosol optical depth at λ, 'β' is the Ångström turbidity coefficient which equals τ at λ=1μm and 'α' is the Ångström exponent. 'α' provides information about aerosol size distribution and Ångström turbidity coefficient (β) gives an idea about atmospheric condition [Eck et al., 1999]. Daily observed τ and derived α is then averaged to obtain a daily mean, which forms basic data set for further analysis.

2.2.2 Aerosol scattering coefficient

A calibrated integrating Nephelometer (TSI- 3563, USA) is used to measure scattering coefficient of aerosols during the study period. It measures aerosol scattering and back scattering coefficient at three different wavelengths viz. 450, 550 and 700nm with high sensitivity and proven accuracy [Anderson et al., 1996]. Its principle of operation is well documented in literatures [Anderson et al., 1996]. It is operated at a flow rate of 20 LPM with a data averaging time of 300 sec during study period. We followed Anderson and Orgen, [1998] to correct measurement biases due to non-angular idealities.

Parameter g, which is the mean value of Cosine of scattering angle over the total solid angle weighted by the phase function is calculated using an empirical relation relating backscattering fraction (b) provided by Andrews et al. [2006]. Uncertainty associated with g calculated using the above method is generally less than 10% [Niranjan et al., 2011].

2.2.3 Estimation of aerosol absorption coefficient.

Aerosol absorption measurements are carried out using a seven channel Aethalometer (AE-31, Magee Scientific, USA) for the same study period. It measures light attenuation at seven wavelengths (370, 470, 520, 590, 660, 880 and 950 nm) through a quartz filter matrix in which particles get deposited. It is operated at a flow rate of 3 LPM with a data averaging time of 5min. Differences in light transmission through particle laden spot and blank portion of filter are attributed to attenuation, which is directly proportional to raw absorption coefficient [Hansen et al., 1984]. There are several literatures explaining methodologies for rectifying the above raw absorption coefficient [Arnott et al., 2005; Weingartner et al., 2003]. In this study we adopted methodology proposed by Arnott et al. [2005]. The expected uncertainties in the estimation of σ_{abs} using this methodology are generally less than 10%.

Simultaneous measurements on absorption and scattering are then used to obtain aerosol single scattering albedo at 550 nm (ω_{550}), which is defined as the ratio of scattering to extinction at 550 nm.

2.3 Radiative transfer model

In this study we used Santa Barbara DISORT Atmospheric Radiative Transfer (SBDART) code [Ricchiazzi et al., 1998] for analyzing radiative transfer in short wave (SW) range (0.25-4.0

μm). It is one of the widely used models to compute radiative transfer in both clear and cloudy atmospheres. Radiative transfer equations for a vertically inhomogeneous, non -isothermal, plane-parallel atmosphereare numerically integrated in SBDART using DISORT (Discrete Ordinate Radiative Transfer) radiative transfer module developed by Stamnes et al. [1988]. Present SBDART code allows up to 65 atmospheric layers and 40 radiation streams (40 zenith angles and 40 azimuthal modes).

The main input parameters like spectral values of τ, ω and angular phase function of scattered radiation for SBDART are taken from in-situ observations. We have used MODIS derived land surface products (MCD43) on black-sky albedo and white-sky albedo to calculate actual albedo over Hyderabad required for radiative transfer calculation [Satheesh et al., 2010]. We also used tropical atmospheric profiles for running the model. With these inputs, we run SBDART with and without aerosols for all clear sky days during the study period with a frequency of 1hour and for 24 hours period. ADRE at TOA and surface, are then estimated by taking difference between instantaneous changes in net solar radiation for with and without aerosols skies [Russell et al., 1999]. Atmospheric effect which implies the amount of energy trapped within the atmosphere due to aerosols is calculated by taking difference between TOA and surface effect. Negative values of ADRE indicate cooling of Earth-Atmosphere system and positive values indicates warming. Overall uncertainty in estimated effect is due to small deviations in simulations and uncertainties in surface parameters used as input to SBDART. Ricchiazzi et al. [1998] reported the accuracy of SBDART is within 3% when compared with precise ground observations in SW range.

In this study, we analysed ARDF for two seasons viz. pre-monsoon and winter as all the concurrent essential *in-situ* input data for RT model run were available for these seasons.

3. Sensitivity analysis of ADRE

Uncertainties involved in estimation of ADRE utilizing in-situ data depends greatly on the quality of data fed in to the radiative transfer model. Uncertainty in ADRE estimation can be achieved following McComiskey et al.[2008]

$$\Delta(ADRE) = \sqrt{\left(\frac{\partial ADRE}{\partial \tau} * \Delta\tau\right)^2 + \left(\frac{\partial ADRE}{\partial \omega} * \Delta\omega\right)^2 + \left(\frac{\partial ADRE}{\partial g} * \Delta g\right)^2} \quad (1)$$

where, $\partial ADRE/\partial\tau$, $\partial ADRE/\partial\omega$ and $\partial ADRE/\partial g$ represents sensitivity of τ, ω and g respectively. Mean value of τ, ω and g observed during study period are independently constrained while performing sensitivity analysis of each. $\Delta\tau$, $\Delta\omega$ and Δg are the uncertainties associated with measurements of τ, ω and g. Calculated uncertainty values for each parameter are ±0.02, ±0.03, ±0.05, respectively and estimated ΔADRE using Equation (1) is 2.4 Wm^{-2}.

4. Results and Discussions

4.1 Temporal variation of aerosol radiative properties over Hyderabad.

Key optical parameters that are required to understand aerosol complex interaction with radiation are τ, ω and Phase function [Haywood and Shine, 1995]. The τ is vertical integral of aerosol extinction coefficient from Earth's surface to TOA. The ω is a function of particle size, shape and its refractive index. Higher magnitude of ω is considered as an index for relative dominance of scattering properties of aerosols with respect to its absorption. Angular distribution of scattering radiation is represented by phase function. Since, estimation of phase function is mathematically complex; Henyey-Greenstein (HG) phase function [Henyey and Greenstein, 1941] is an often-used approximation for phase functions to speed up calculations in some radiative transfer codes.

Monthly and seasonal variation of aerosol radiative properties (τ, ω, α and g) at 550 nm over Hyderabad is shown in Figure-1. Vertical bar represents standard deviation from mean. Figure showed a significant seasonal variation of aerosol radiative properties during study period. Mean τ_{550} showed a minimum (0.45-0.52) during winter and maximum (0.6-0.7) in pre-monsoon. Observed high values during pre-monsoon can be associated with high surface temperature leading to aerosol lofting by convection coupled with lack of removal mechanism. Babu et al. [2011] reported two elevated BC layers one at ~4.5 km, and another above 8 km over this region using balloon experiment. Also dust storms and biomass burning are very frequent over Indian region during this period [Gharai et al., 2013], and elevated layers of these aerosols can be observed in locations away from source regions [Jose et al., 2015a]. Local re-suspended soil dust also play a significant role in columnar aerosol loading during this period. One of the recent studies [Babu et al., 2016] on vertical distribution of atmospheric aerosols over Indian mainland have reported a clear enhancement in aerosol loading and its absorbing nature at lower free troposphere (FT) levels (above the planetary boundary layer) during pre-monsoon compared to winter. Model simulations of BC over Indian region [Kumar et al., 2015] also reported that BC showed distinct but opposite seasonality in the lower troposphere (LT) and FT with BC showing winter maximum and summer minimum in the LT and vice versa in the FT. Ångström parameter, α (380-870nm), which represents spectral dependencies of AOD showed lowest values ~0.56 during May-Jun (pre-monsoon) and maximum values >1.25 in Nov-Dec (post-monsoon to winter) during the study period. This large seasonality in α could be due to dominance of fine mode aerosols in winter and mixed with coarse mode aerosols in pre-monsoon months. Eck. et al. [2010] also reported similar seasonal variations of α(440-870nm) over Kanpur, India.

The ω of aerosols during study period showed a maximum in winter and minimum in pre-monsoon. Observed mean values of ω during pre-monsoon, and winter are 0.67±0.06, and 0.78±0.05, respectively. This seasonal difference in ω significantly affects TOA effect over this region as a small difference in ω can change the sign of TOA forcing [Hansen et al, 1997]. The ω observed over Hyderabad is comparable to those reported from other Indian urban cities, like Ahmedabad [0.62-0.84], Delhi [0.63-0.72] and Vishakhapatnam [0.65-0.9] [Srivastava et al., 2011; Soni et al., 2010; Niranjan et al., 2011]. However, observed ω over Hyderabad is lower than those reported from rural locations [Montilla et al., 2011], which can be due to enhancement in absorbing particles like BC over urban sites. Reported BC mass fraction to total suspended particles (TSP) over Hyderabad varies from 5-15% [Sinha et al., 2013] which is quite high compared to other sites of the country [Safai et al., 2013]. Pre-monsoon minimum in ω can also be associated with presence of more long range transported aerosols (biomass burning and dust) in addition to local emissions, which are more absorbing in nature. Pandithurai et al. [2008] reported a reduction in ω from 0.84 to 0.74 over New Delhi from March to June, when dust transports are more prominent.

Our analysis shows that during study period g varied in the range 0.53±0.04 in winter and 0.48±0.02 in pre-monsoon. The column averaged g values during pre-monsoon over other locations of the country are found to be 0.44 ± 0.05 (Bhubaneshwar), 0.48 ±0.03 (Chennai), 0.43 ± 0.07 (Trivandrum) and 0.38 ± 0.06(Goa), respectively [Ramachandran and Rajesh., 2006]. Sensitivity analysis of g with respect to TOA effect suggests that a 10% decrease in g corresponds to a 19% reduction in radiative effect at TOA and a 13 % reduction at the surface as reported by Niranjan et al. [2011].

4.2 Aerosol direct radiative effect (ADRE).

Daily ADRE is computed using SBDART with observed aerosol optical parameters as primary inputs. Figure 2a shows averaged ADRE at TOA, Surface and Atmosphere during winter and pre-monsoon seasons of the study period. Vertical lines on each bar represent the respective standard deviation. It is observed from figure that aerosol radiative effect at TOA during winter is -11.72 ±3.5 Wm^{-2}, while during pre-monsoon it becomes -5.5 ±2.5 Wm^{-2}. This is largely due to seasonal variation of ω, as it varies from 0.67±0.06 to 0.78±0.05 from pre-monsoon to winter. Figure 2b shows percentage occurrence of ADRE at TOA during winter and pre-monsoon. Analysis reveals that during winter about 96% of the days $ADRE_{TOA}$ are having negative values indicating a cooling effect, while during pre-monsoon ~41% of the days $ADRE_{TOA}$ are having positive values. Corresponding effect at surface during winter and pre-monsoon are found to be -46.25±3.4 Wm^{-2} and -90.56.4±2.4 Wm^{-2}, respectively. Atmospheric effect which represents energy trapped within the atmosphere is calculated as difference between TOA and surface effect. Estimated atmospheric effect during winter and pre-monsoon over the study period is found to be 34.52±6 Wm^{-2} and 85.05±10 Wm^{-2}, respectively. In the present study, heating rate is calculated following Liou et al.[2002]. Reported boundary layer height (BLH) over this region is ~3.2 and 1.7 Km, respectively during pre-monsoon and winter [Mahalakshmi et al., 2011]. Owing to maximum BLH of ~3 km, we considered pressure difference (ΔP =300hPa) between the surface and 3 km height for calculating heating rate. Heating rate during winter and pre-monsoon are 1.12 and 2.75 $Kday^{-1}$, respectively. However, realistic value of heating rate may deviate as much as 30 to 50% depending on aerosol mixing height compared to that estimated using a mean ΔP of 300 hPa [Babu et al., 2007].

Aerosol radiative effect over any location is very intricately dependent on several parameters such as total column burden of aerosols, their vertical distribution in the atmosphere, single scattering albedo, their size distribution, scattering phase functions, reflectance of underlying surface, relative humidity in the atmosphere, solar insolation and many more. A comparison study of ADRE reported from other sites of the country including oceanic regions will be of great interest and also provides an insight into radiative impact caused by aerosols in this part of the world. Figure 3 shows spatial distribution of reported atmospheric effect (Wm^{-2}) due to aerosol over Indian subcontinent and adjacent oceanic regions. It can be observed that during per-monsoon significantly high atmospheric effect (>30 Wm^{-2}) is observed over Indian land mass compared to winter. Oceanic regions surrounding Indian landmass especially Bay of Bengal showed high atmospheric effect (~11Wm^{-2}) in comparison to the Arabian Sea (~3Wm^{-2}) [Moorthy et al., 2009]. High atmospheric effect observed during pre-monsoon can have larger implications on regional-scale dynamical process such as wind flow, convective activity and even precipitation patterns [Menon et al., 2002]. Ramanathan et al. [2001] have pointed out that high atmospheric effect over Indian Ocean will reduce evaporation from ocean surface; thereby reducing moisture inflow which in turn weakens the monsoon rainfall. On the contrary, Lau et al. [2006] and Lau and Kim [2006] have proposed the Elevated Heat Pump (EHP) hypothesis, suggesting that desert dust, mixed with soot aerosols over northern India and the foothills of the Himalayas, may cause enhanced heating in the middle/upper troposphere over southern slopes of the Tibetan Plateau [Gautam et al., 2009] and in turn strengthened the land-sea gradient, thus resulting in the advancement of the monsoon rainfall in early summer. From above discussions it's obvious that aerosols undoubtedly have a role in climate change, whether it enhances or offsets this change is still a matter of scientific debate and have to be assessed carefully with more observational studies.

4.3 Influence of long range transport.

In this section we tried to analyse the influence of long range transported aerosols over study area. Previous studies [Badrinath et al., 2009; Jose et al., 2015a] have reported the presence of long range transported biomass burning and dust aerosols during pre-monsoon over study area under favorable metrological condition. These episodic events during study days are identified by analyzing intrinsic aerosol optical properties like scattering Ångström exponent (SAE) and absorption Ångström exponent (AAE). Spectral dependence of aerosol scattering (absorption) coefficient is quantified in terms of SAE (AAE) and is estimated by negative slope of scattering (absorption) vs. wavelength in log-log plot. SAE gives an understanding of size distribution of aerosol particles, while AAE reveals the source characteristics of absorbing aerosols. AAE ~1 indicates the dominance of fossil fuel aerosols, AAE ≥2 indicate the presence of dust and biomass burning aerosols and AAE between 1 and 2 can be considered as a mixture of both [Bergstorm et al., 2002]. SAE depends primarily on the particles' size and ranges from 4 (Rayleigh atmosphere) to 0 (large particles) [Valenzuela et al., 2014]. Analysis shows that mean SAE and AAE values of aerosol particles over the study area during pre-monsoon are 1.56±0.27 and 1.19±0.16, respectively. These values represent a typical urban distribution where local emission dominates other sources. Figure 4 shows the temporal variation of SAE and AAE during pre-monsoon days during study the period, episodic events are identified and highlighted in red circles for dust storm events and blue circles for biomass burning days. Each episodic event is further confirmed with satellite observations. In order to demonstrate,satellite observations during 22-23March, 2012 are investigated, where the regional atmosphere is influenced by both long range transported dust and aerosols due to biomass burning. True color image composite of event day (22-23 March 2012) as seen by MODIS is shown in Figure 5(a) and the overall aerosol loading as observed by Terra MODIS is shown in figure 5(b). An intense dust loading can be seen on the Persian Gulf region (North western side of study area). Jose et al. [2015c] analysed the impact of this intense dust storm on SW radiation at TOA using satellite measurements over the Arabian Sea (AS) and reported, a TOA SW aerosol radiative force efficiency of -39 Wm^{-2} per unit AOD. MODIS derived active fire location [Gigilo et al., 2003] are overlaid on the image, which clearly shows the presence of biomass burning over Central India. Five days backward trajectory obtained from Hysplit model [Draxler et al., 2003] ending over study area on the same day at different altitude (1, 2 and 4Km) overlaid on true color image also indicates the possible transport of dust and biomass aerosols towards study site. During this event an enhancement in τ (0.5-0.7) and ω (0.7-0.8) is observed over study site. An enhancement in AAE during these days indicates more loading of biomass aerosols as the source region is in close proximity to the study site. However a decrement in SAE from 1.5 to ~1 during the event days also reveals the possible presence of coarser aerosols which could be due to presence of dust aerosols. Estimated effect on 22March, 2012 at TOA, surface and atmosphere are -8.1, -91.51 and 83.4 Wm^{-2}, respectively. Variations of aerosol optical properties and estimated effect over the study area during similar episodic events occurred are given in Table1. Important observations that can be drawn from these events are; during dust storm events α values are found lesser than those compared with biomass events. The ω during majority of dust storm days (except on 23March 2012 and 3May2012) are lower (~0.77) than biomass burning days (0.81). Low value of ω during dust storm prominent days over study area can be possibly due to deposition of black carbon on dust particles [Satheesh et al., 2006]. Chemical analysis of dust samples collected over Kanpur, India during dust events also reveals possible mixing of anthropogenic pollution and estimated ω is found to be low (0.74) due to presence of black carbon [Chinnam et al., 2006]. Our analysis revealed an enhancement of heating rate between

1.55 to 3.6 Kday^{-1} occurred during dust dominant days, which can be attributed as enhanced absorption by possible mixing of dust with local atmosphere. In contrast to dust storm days, high ω are observed during biomass burning days suggests an increase in organic carbon which is light scattering part in carbonaceous aerosols and thereby induce a cooling effect at TOA over the study area. Net atmospheric effect during biomass burning events varied from ~50-80 Wm^{-2}; which corresponds to a heating rate of 1.5 to 2.7 Kday^{-1}.

5. Conclusions

Present study estimated ADRE over an urban location (Hyderabad) in Central India during the study period 2010-2012. Critical aerosol optical parameters required for estimation of ADRE are either measured or derived from *in-situ* observations. The salient findings are the following:

- Aerosol radiative properties showed significant seasonal variation during the study period. Seasonal variation of microphysical aerosol optical properties (α, SAE) indicate~~s~~ the dominance of fine mode aerosols over study area; also presence of coarse mode aerosols are identified during pre-monsoon months.
- Mean 'τ_{550}' showed a minimum (0.45-0.52) during winter and maximum (0.6-0.7) in pre-monsoon. While, ω of aerosols showed a maximum in winter (0.78±0.05) and minimum in pre-monsoon (0.67±0.06). The g varied in the range 0.48±0.02 to 0.53±0.04 during the study period.
- Estimated radiative effect at TOA during winter is -11.72 ±3.5 Wm^{-2}, while during pre-monsoon, it is -5.5 ±2.5 Wm^{-2}. Corresponding values at surface during winter and pre-monsoon are -46.25±3.4 Wm^{-2} and -90.56.4±2.4 Wm^{-2}, respectively.

- Atmospheric radiative effect during winter and pre-monsoon are 34.52±6 Wm^{-2} and 85.05±10 Wm^{-2}, respectively. This corresponds to a heating rate of about 1.12 and 2.75 $Kday^{-1}$, respectively.
- Investigation on identified dust storm and biomass burning events reveals that α values are lesser during dust storm than those compared with biomass events. Also during majority of dust storm days, ω are lower (~0.77) than biomass burning days (0.81).
- Estimated heating rate due to biomass burning and dust storm aerosols on local atmosphere are about 2.26±0.51and 2.08±0.6 $Kday^{-1}$, respectively.

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