Aerosol Radiative Forcing

The global climate system is balanced by interactions between its various subcomponents. The main processes that determine the overall equilibrium state of the climate system are heating by the incoming solar shortwave radiation and cooling by the outgoing longwave (infrared) terrestrial radiation. Any process that can disturb the energy balance can cause climate change. A process that alters the radiation balance of the climate system is known as radiative forcing (Coakley et al., 1983; Coakley and Cess, 1985; Ramanathan et al., 1989; Charlson et al., 1991; Bates, 1999). Radiative forcing can be external or internal. External forcing operates from outside the Earth’s climate system and includes orbital variations and changes in incident solar flux at the top of the atmosphere. Internal forcing is caused due to changes in atmospheric compositions and the best examples are the Greenhouse gases and aerosols in the atmosphere (Haywood and Ramaswamy, 1998; Myhre et al., 1998; Haywood et al., 2003). While Greenhouse gases are known to decrease the outgoing long wave radiation and contribute to global warming (positive forcing), aerosols can either increase or decrease the outgoing solar radiation by scattering...
or absorption (direct effect) (Jayaraman et al., 1998) or by changing the microphysical properties of the clouds (indirect effect) (Lohmann and Feichter, 2005).

The change in radiative fluxes due to aerosols is mainly by scattering and absorption of the incoming shortwave solar radiation and outgoing longwave terrestrial radiation. The interaction depends on the physical, optical and chemical properties of aerosols. Aerosols have different properties according to their natural and anthropogenic origin. Besides, the abundance of natural aerosols on the global scale is several times greater than that of the major anthropogenic aerosols. For example, sea-salt radiative forcing is in the range of -0.6 to -2 Wm\(^{-2}\) at low wind speeds and at higher wind speeds this can be as high as -1.5 to -4 Wm\(^{-2}\) (Winter and Chylek, 1997), whereas the change (between now and the early 20th century) in the reflected solar flux (the direct effect) due to anthropogenic sulfate averaged over the northern hemisphere is estimated as -1.1 Wm\(^{-2}\) (Charlson et al., 1991). On the global scale, anthropogenic sources contribute a small fraction of 10-15% of aerosol to the total production from various sources (Kiehl and Rodhe, 1995; Andreae, 1995). But on a regional scale, it may be much greater than the global average due to large production rate and short residence time. Aerosols, with varying optical properties are non-uniformly distributed around the globe, and their radiative forcing also varies accordingly.

5.1 Models for Radiative Forcing Calculation

5.1.1 Optical Properties of Aerosols and Clouds

To compute the regional aerosol radiative forcing the spectra of Aerosol Optical Depth (AOD), Single Scattering Albedo (SSA) and asymmetry parameter are required for that particular location. The OPAC (Optical Properties of Aerosols and Clouds by Hess et al. (1998)) model is used to infer indirectly the chemical compositions of aerosols and derive these parameters, due to the unavailability of chemical data of the various aerosol components. OPAC mainly has 10 aerosol components.
which are insolubles (mostly soil particles), water soluble aerosols (mainly sulfate and nitrate aerosols of anthropogenic origin), soot (of anthropogenic origin), sea salts (naturally produced on the oceanic surface by wind and also available in the atmosphere of coastal regions) in accumulation and coarse mode, mineral dust (generally coming into atmosphere from the arid surface by wind) in three modes, mineral-transported and sulfate droplets (mainly found at stratospheric altitude). This model is used to derive the AOD spectrum using a combination of these aerosol components and in the present study the sulfate droplets are not considered. OPAC model is useful to derive different microphysical and optical properties for suitable mixture of aerosol components at the measurement site to best fit the observed aerosol optical depth. Some of the aerosol components which are hygroscopic in nature, may change their optical properties, and hence OPAC outputs are available for eight different relative humidity (0%, 50%, 70%, 80%, 90%, 95%, 98% and 99%) conditions.

Optical properties for different aerosols are different. SSA, one of the important parameters for radiative forcing calculation, determines the sign of radiative forcing. Fig. 5.1 shows the computed SSA for different aerosol species using the data from OPAC. Water soluble (sulfate, nitrate, etc.) and sea salt do not absorb significantly in the visible range (SSA ≥ 0.9 at 0.5 μm) but they absorb significantly in the infrared region (SSA ≤ 0.4 at 10.0 μm). Major aerosol components are scattering type in the shortwave range (0.25-4.0 μm) whereas in the longwave range (4.0-40.0 μm) they can be totally absorbing such as soot. The SSA of soot in the shortwave is ~ 0.22 (at 0.5μm), whereas in the longwave range it is totally absorbing. Dust is mainly scattering in nature in the shortwave range and in the longwave range it exhibits significant absorption. On one hand in the longwave range absorption decreases the outgoing radiation while on the other hand the energy re-emitted consequent to this absorption increases the surface reaching infrared radiation. The net SSA over a particular location is weighted average of the SSA of all the aerosol components.
Aerosol Radiative Forcing

5.1.2 Santa Barbara Discrete Ordinate Radiative Transfer

Aerosol radiative forcing (ARF) is computed using Santa Barbara DISORT Atmospheric Radiative Transfer (SBDART) (Ricchiazzi et al., 1998) code, a well established code for estimation of radiation flux in the shortwave (0.25-4.0 μm) as well as longwave (4.0-40.0 μm) range. SBDART is a radiative transfer code that computes plane-parallel radiative transfer in clear and cloudy conditions within the earth’s atmosphere and at the surface. In the present study only clear sky conditions are considered. All the important processes that affect the ultraviolet, visible, and infrared radiation, are included in this code. For molecular absorption SBDART uses the low-resolution band models of LOWTRAN-7 atmospheric transmission code (Pierluissi and Peng, 1985). LOWTRAN-7 codes can take into account
the effects of all radiatively active molecular species found in the earth's atmosphere with wavelength resolution of about 5 nm in the visible and about 200 nm in the thermal infrared. In SBDART, the radiative transfer equations are numerically integrated with DISORT (Discreet Ordinate Radiative Transfer) code (Stamnes et al., 1988). This discrete ordinate method provides a numerically stable algorithm to solve the equations of plane-parallel radiative transfer in a vertically inhomogeneous atmosphere. The intensity of both scattered and thermally emitted radiation can be computed at different heights and directions. Presently, SBDART is configured to allow up to 50 atmospheric layers and 20 radiation streams (20 zenith angles and 20 azimuthal modes).

The ground surface cover is an important determinant of the overall radiation environment because spectral albedo of the surface which defines the ratio of upwelling to downwelling spectral irradiance at the surface, determines upwelling irradiance from the surface. In SBDART there are five basic surface types, namely (1) ocean water (Tianre et al., 1990), (2) lake water (Kondratyev, 1969), (3) vegetation (Reeves et al., 1975), (4) snow (Wiscombe and Warren, 1980) and (5) sand (Staetter and Schroeder, 1978). The spectral albedo describing a given surface is often well approximated by combinations of these basic surface types. Input parameters in SBDART allow the user to specify a mixed surface consisting of weighted combinations of water, snow, vegetation and sand. SBDART can compute the radiative effects of several lower and upper atmosphere aerosol types. In the lower atmosphere, typical rural, urban, or maritime conditions can be simulated using the standard aerosol models of Shettle and Fenn (1975). SBDART gives the opportunity to calculate radiative effects up to five aerosol layers specified (i.e., at five different altitudes) that model fresh or aged volcanic, meteoric, and upper-tropospheric background aerosols. Fig. 5.2 shows the solar irradiance at top of atmosphere and at the surface.

The major inputs required to estimate the aerosol radiative forcing for DISORT module in SBDART include spectral values of solar radiation incident on the atmosphere, spectral values of columnar AOD, SSA and angular phase function of the
scattered radiation or asymmetry factor ($g$). The asymmetry factor is used to generate a scattering phase function through the Henyey-Greenstein approximation. The Henyey-Greenstein parameterization provides good accuracy when applied to radiative flux calculations (van de Hulst, 1968; Hansen, 1969). Spectral values of AOD, SSA and asymmetry parameter are obtained from OPAC. OPAC model derived aerosol optical parameters are obtained by varying the number concentration of individual components in small steps until all the following criteria are satisfied. (1) Sum of root mean square difference between the model estimated and the observed AOD values at all the six wavelength channels is minimum (for the present study, we have constrained this sum to within 0.1 AOD). (2) Angstrom parameters for the observed and the model estimated AOD spectra are comparable. (3) Total mass concentration for the model estimated aerosol mixture are comparable with aerosol mass concentrations measured using QCM (Quartz Crystal Microbalance) cascade impactor from stage 2 to stage 10 (4) Black carbon (BC) mass used in the model is comparable to the measured value using Aethalometer. (5) Mass fraction of BC in the total aerosol mass are also constrained close to their actual values derived from simultaneous and co-located measurements using Aethalometer and QCM impactor. (6) Model derived values of SSA at 0.5 μm closely matches with the SSA at 0.525 μm estimated using simultaneous measurements of absorption.

Figure 5.2: Solar irradiance (W m⁻² μm⁻¹) computed at 0° solar zenith angle at top of atmosphere (TOA) and Earth’s surface for no aerosol condition (left) and at Earth’s surface for different atmospheric conditions (right) obtained from SBDART code.
Aerosol Radiative Forcing

and scattering coefficient of aerosols.

5.2 Radiative Forcing for Different Environmental Conditions

5.2.1 Radiative Forcing in Urban Environment

In this study, aerosol radiative forcing is calculated over Ahmedabad, an urban region in the western part of India using Santa Barbara DISORT Atmospheric Radiative Transfer (SBDART) code (Ricchiazzi et al., 1998) in the Short Wave (SW) region (0.25-4.0 μm). Radiative forcing calculations are performed only for clear sky days. The spectral values of SSA, asymmetry parameter etc. required for the SBDART are obtained from OPAC by best fitting of model obtained AOD spectral values and measured AOD values at six wavelength channels (0.38, 0.44, 0.5, 0.675, 0.87 and 1.02 mm) using Microtops Sunphotometer, as discussed earlier. The measured values from Microtops are used for the columnar water vapor for individual days and ozone concentration data is obtained from TOMS (Total Ozone Mapping Spectrometer) satellite. One of the important parameters which can introduce large errors in radiative forcing calculation over land regions like Ahmedabad is the surface albedo of the location (Wielicki et al., 2005). In the present study, MODIS (Moderate Resolution Imaging Spectroradiometer) derived surface reflectance data is used over Ahmedabad to estimate radiative forcing calculation. For this the surface reflectance data is obtained from Nadir BRDF-Adjusted Reflectance 16-Day L3 Global 0.5 km SIN Grid product which is derived at the mean solar zenith angle of Terra overpasses for every successive 16-day period, calculating surface reflectance as if every pixel in the grid are viewed from nadir direction. Surface reflectance data available in seven wavelength bands of MODIS centered around 0.47, 0.56, 0.65, 0.86, 1.24, 1.64 and 2.13 μm are used to reproduce the spectral dependence of surface albedo for the entire SW range using a combination of three
different surface types, namely, vegetation, sand and water. The monthly variation of surface reflectance data available at seven wavelength bands during 2007 is shown in Fig. 5.3. Vertical lines in this figure represents ±1 σ variation about the monthly mean value of surface reflectance measured. On an average surface reflectance values are found to be high during premonsoon (Apr-May) and low during postmonsoon (Sep-Nov) and winter (Dec-Feb). Another very important aerosol parameter required for radiative transfer calculations, is the vertical profile of aerosols in the atmosphere. The major uncertainties introduced in the estimation of radiative forcing calculation are due to the uncertainty in the vertical distribution of aerosols and the single scattering albedo (Haywood and Ramaswamy, 1998; Chung et al., 2005). Fig. 5.4 shows the seasonal variation of SSA at Ahmedabad obtained from OPAC model. During monsoon SSA is maximum at 0.86±0.04 whereas during postmonsoon SSA is minimum at 0.80±0.09.

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**Figure 5.3**: Monthly variation of surface reflectance obtained from MODIS satellite at different wavelengths over Ahmedabad.
The different kind of aerosols (scattering or absorbing) present at different altitude can cause cooling or heating in the atmospheric layer and influence the temperature profile. This can even further change the precipitation pattern in the tropical regions which is mainly caused by the convection processes in the atmosphere over those regions (Chung and Zhang, 2004). For the present radiative transfer calculations aerosol extinction profiles obtained from Micro Pulse Lidar measurements which are normalized for the AOD value at the MPL wavelength (0.523 μm) have been used. Fig. 5.5 shows the monthly variation of aerosol profile obtained from Micro Pulse Lidar over Ahmedabad during 2002-2004.

The aerosol radiative forcing estimation on clear sky days is obtained by running the radiative transfer model for the calculated radiant fluxes with and without

Figure 5.4: Seasonal variation of Single Scattering Albedo at Ahmedabad obtained from OPAC model.

![Graph showing seasonal variation of Single Scattering Albedo](image-url)
Aerosol conditions in the atmosphere at every hour of the day over the entire SW region. Fig. 5.6 shows the seasonal variation of the averaged values of entire SW region of aerosol direct radiative forcing at TOA, surface and in the atmosphere over Ahmedabad for two years 2006 and 2007. Vertical lines on top of each bar represent ±1 σ variation about the mean value of radiative forcing for a particular season. The important results found from this study are as follows. During winter both the radiative forcing at surface and TOA are negative while during premonsoon surface forcing is negative and TOA forcing is positive. During monsoon and postmonsoon TOA forcing changes its sign from positive to negative. The radiative forcing values at surface level during premonsoon and postmonsoon are very similar and are -46.0±3.6 Wm⁻² and -41.0±11.6 Wm⁻², respectively. During postmonsoon there is large variation about the mean. During monsoon surface
radiative forcing is minimum at \(-28.1\pm13.3\) Wm\(^{-2}\) and during winter the value is slightly higher at \(-36.6\pm3.9\) Wm\(^{-2}\). In case of TOA, aerosol radiative forcings are found at \(-1.8\pm1.8\) Wm\(^{-2}\) during winter, \(3.1\pm1.1\) Wm\(^{-2}\) during premonsoon, \(-0.5\pm1.8\) Wm\(^{-2}\) during monsoon and \(-1.4\pm3.5\) Wm\(^{-2}\) during postmonsoon. The atmospheric aerosol radiative forcing is the difference between the TOA and surface radiative forcing. The atmospheric aerosol radiative forcing represents the amount of energy trapped in the atmosphere by aerosols and is a measure of the atmospheric heating (Ramanathan et al., 2001a). This forcing generally increases as the TOA forcing becomes positive and decreases as TOA forcing becomes negative. During the study period, the atmospheric forcings are \(33.8\pm5.3\) Wm\(^{-2}\),
49.1±4.3 Wm\(^{-2}\), 27.5±14.7 Wm\(^{-2}\) and 39.5±15.1 Wm\(^{-2}\) during winter, premonsoon, monsoon and postmonsoon, respectively. The aerosol radiative forcing over a location is highly influenced by several parameters like total aerosol load in the atmosphere, their vertical and size distribution, single scattering albedo, scattering phase function, reflectance of the Earth’s surface, solar insolation, meteorological parameters mainly wind, temperature and relative humidity (Haywood and Boucher, 2000). The surface reflectance mainly changes by the area covered by vegetation. In Fig. 5.7 shows the surface reflectance of western part of India observed by the MODIS satellite during April and October, which has a large influence on the TOA radiative forcing. In this case, during premonsoon and monsoon forcings affect the regional dynamical systems such as wind, thermal convection and also the precipitation patterns (Menon et al., 2002). Therefore, Earth-Atmospheric system can be perturbed on a large scale for this kind of regional cooling at the surface and warming of the lower troposphere. Normally the radiation flux reaching the surface is balanced by the evaporation (latent heat flux) and sensible heat flux from the surface. From the observations over the Indian Ocean region it is seen that there is greater probability of solar flux reduction at surface being balanced by
reduction in evaporation and therefore, this reduction of evaporation slows down the hydrological cycle (Ramanathan et al., 2001b,a).

Large seasonal variation of the aerosol properties is observed in the present study and hence in the aerosol radiative forcings. Radiative forcing is useful for better understanding of how the various aerosol components in the atmosphere over a region can perturb the Earth-Atmospheric system. At Ahmedabad, during winter, the aerosols are mainly concentrated in the low level atmosphere and the BC mass concentration is significantly high at 3.3 μg.m⁻³. So the surface solar reduction is high due to these absorbing BC aerosols. Similarly, radiative forcing is also negative at TOA. Therefore, energy is trapped significantly in the atmosphere. During premonsoon, aerosols are distributed in the higher altitudes also because the boundary layer height increases and BC mass concentration also reduced to 2.5 μg.m⁻³. In addition, the abundance of dust aerosols is enhanced. This dust aerosols are mainly scattering particles and hence the forcing becomes positive at the TOA and large energy is trapped into the atmosphere. The atmospheric heating by absorbing aerosols can evaporate some of the low-level clouds, resulting in a decrease of cloud cover and planetary albedo (Ackerman et al., 2000). This kind of cloud burnings can influence the cloud coverage and hence the precipitation amount over the Indian region where the maximum annual rainfall is received during monsoon season. In another work related to climate effects of BC aerosols over China and India Menon et al. (2002) have shown that large amounts of BC (soot) particles and other pollutants are responsible for causing the observed changes in precipitation patterns and temperature trends over China in particular. During 2006 and 2007 the rainfall during monsoon over Ahmedabad is relatively higher in comparison to previous years. During monsoon due to heavy rain aerosols are washed out and reduces the forcing value at surface. However, during the end of monsoon and the beginning of postmonsoon a stable aerosol layer is observed in the the vertical aerosol distribution (Ganguly et al., 2006a) and therefore large amount of solar flux is reflected by this layer and the TOA atmosphere forcing becomes positive. In addition to that, during postmonsoon, BC mass concentration
is maximum at 5.7 μg.m⁻³ and these BC aerosols absorb large amount of solar radiation and further reduce the radiation at the surface. Therefore, a combination of all these effects results in large amount of energy being trapped in the atmosphere.

### 5.2.2 Radiative Forcing in Hilltop Environment

Radiative forcing is calculated over Mt. Abu using the same procedure of that at Ahmedabad (Sec. 5.2.1). Fig. 5.8 shows the seasonal variation of SSA at Mt. Abu obtained from OPAC model. It shows a large value of 0.90±0.03 and 0.90±0.02 during winter and postmonsoon, respectively, due to less abundance of absorbing aerosols and a minimum of 0.83±0.01 during postmonsoon due to enhancement of anthropogenic activities.

![Figure 5.8: Seasonal variation of Single Scattering Albedo at Mt. Abu computed using the OPAC model.](image_url)
Among the other inputs, ozone and water vapor values are used from the TOMS data and Microtops measurements, respectively. Another important parameter that can introduce large error in the TOA radiative forcing over land is surface reflectance. In the present study, 8-days MODIS derived Level 3 surface reflectance data with the grid of 0.5 km for seven wavelengths centered at 0.47, 0.56, 0.65, 0.86, 1.24, 1.64 and 2.13 \( \mu m \) have been used. Fig. 5.9 shows the monthly variation of surface reflectance values. The spectral surface reflectance values of Mt. Abu are considered as the combination of three different surfaces, viz., water, sand and vegetation, and are used as input to the SBDART code for Mt. Abu radiative forcing calculation. In comparison with Ahmedabad, as shown in Fig. 5.10 Mt. Abu has higher surface reflectance values during premonsoon (Apr-May) while lower values during postmonsoon (Sep-Dec). This implies that during premonsoon, Mt. Abu has less vegetation and bare sandy surface enhances the surface reflectivity while during postmonsoon the surface is totally covered by vegetation which reduces surface reflectivity at 1.64 \( \mu m \) wavelength. In comparison to
Ahmedabad the vegetation covered by the surface is more uniform and therefore, the reflectivity is higher at Mt. Abu.

Fig. 5.11 shows the seasonal variation of SW aerosol radiative forcing at the surface and TOA. The variation of surface forcing is very similar to variation of AOD. At the surface large variation of forcing has been found in the range -23.4 to -8.8 Wm$^{-2}$ while at TOA forcing varied from -3.2 to +0.2 Wm$^{-2}$. During winter the estimated forcing is minimum at -8.8 Wm$^{-2}$ at the surface and -2.7 Wm$^{-2}$ at TOA. During premonsoon radiative forcing becomes maximum at -23.4 Wm$^{-2}$ while at TOA forcing changes its sign and becomes +0.2 Wm$^{-2}$. After winter the forcing gradually decreases at surface and becomes negative at TOA. One of the main reason for the surface radiative forcing variation is due to the boundary layer height variation at hill top region. During winter the boundary layer height becomes less than the station’s altitude and as a result the observatory region becomes clearer and therefore at the surface the solar radiation reduction has been reduced due
to less aerosol present in the atmosphere. But surface forcing becomes more negative as the boundary layer height gradually increases and aerosols increases at the observation site. During April the surface forcing was enhanced by a factor of four not only due to the enhancement of boundary layer height but also due to an increase of anthropogenic activities by local tourists. The BC measurements also show an enhancement during April and these absorbing anthropogenic BC cause large reduction of solar radiation at the surface. During the rest of the season surface radiative forcing reduces. The absorption of BC reduces during that period due to transportation of the BC aerosols by strong wind. In addition enhancement of boundary layer height results in a larger room to dilute the BC concentration. Therefore, the reduction of solar radiation by the absorbing BC particles decreases
and hence surface forcing reduces. On the other hand, during premonsoon season dust storms occur frequently in the desert areas and due to these dust storms there is a large transportation of dust particles to the hill top region. This transported dust aerosol scatter more solar radiation and result in positive forcing at TOA. And during monsoon forcing reduces as heavy rainfall washes out the dust aerosols from the atmosphere. During monsoon there is also large transportation of seasalt aerosols from Arabian sea (Rastogi and Sarin, 2005a). These aerosols enhance the total aerosol burden in the atmosphere and forcing becomes -13.8±3.3 Wm$^{-2}$ at surface. During postmonsoon the transportation of seasalt gradually decreases. Therefore, at TOA scattered solar radiation by aerosols reduces and TOA forcing becomes more negative at -3.2±1.2 Wm$^{-2}$. This also increases the surface reaching solar radiation and reduces the surfaces forcing at -10.6±2.9 Wm$^{-2}$. During winter the boundary layer height goes below the measurement site and the site is in free-tropospheric region. Therefore surface forcing reduces to minimum.

5.2.3 Fog Induced Aerosol Radiative Forcing

Estimation of aerosol components

To estimate chemical compositions of aerosol, OPAC (Optical Properties of Aerosols and Clouds) model is used to best fit the observed aerosol optical depth spectra. This will help in understanding the changes in aerosol physical and optical properties such as aerosol scattering, absorption and extinction coefficients and single scattering albedo ($\omega_0$) during different RH conditions. When RH increases water soluble particles such as sea salt and sulfates will grow (Hess et al., 1998) and when RH exceeds 80%, the aerosol components can be modified by both particle growth and transformation to an aqueous solution which in turn can affect the refractive index of the components (Lubin et al., 2002). For all days, the AOD observed during the afternoon hours, when the RH value is less than 50% are compared with OPAC produced AOD (Fig. 5.12) and the relative abundances of aerosol components such as water soluble (WS), insoluble (IS), soot (SO) and mineral dust (MD)
are estimated. The most dominating components are found to be WS and SO. The soot particle concentration is found between the continental average and the urban aerosol types as defined by Hess et al. (1998). In Table (5.1) aerosol components for three different days representing, prior to foggy days, during foggy days and after the foggy days are given and the corresponding angstrom parameter, are compared. During foggy days both soot and WS particle concentrations have increased and after foggy days though their number concentrations have decreased and they have not reached the values that were observed prior to the foggy days.

**Table 5.1: Aerosol components viz., Soot (SO), Water Soluble (WS), Water Insoluble (IS) and Mineral Dust (MD) are estimated from OPAC model fitting of the observed aerosol optical depth (AOD) spectra. The Angstrom Parameter (a) estimated for the observed as well as the fitted AOD spectra are compared**

<table>
<thead>
<tr>
<th>Date</th>
<th>Aerosol Components</th>
<th>Angstrom Parameter (a)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>SO (par/cc)</td>
<td>WS (par/cc)</td>
</tr>
<tr>
<td>8 Dec 2004</td>
<td>24000</td>
<td>18000</td>
</tr>
<tr>
<td>13 Dec 2004</td>
<td>60000</td>
<td>27000</td>
</tr>
<tr>
<td>27 Dec 2004</td>
<td>55000</td>
<td>12000</td>
</tr>
</tbody>
</table>
Aerosol mass concentrations measured by QCM and HVS are compared with the OPAC estimated aerosol mass concentration (Fig. 5.13). While HVS gives the total mass concentration of particles collected on quartz filter paper for a whole day (24 hours) and used further for chemical analysis (Ramachandran et al., 2006) the QCM measurements were limited to only daytime and data corresponding to 50% RH or less are only considered. With increasing RH the WS component will undergo a major change in their size distribution. Also, the aerosol components can mix internally and externally with other components and hence the relative contribution of different components to the total aerosol loading can change after the foggy period. Therefore, on foggy days the AOD spectra obtained during the afternoon hours are only considered (after the fog has dissipated) and fitted with OPAC model. Prior to foggy days, the aerosol single scattering albedo (SSA) at 500


\( \mu m \), determined by OPAC fitting the observed AOD spectra SSA was 0.83±0.03 at 50% RH while during foggy days (in the afternoon hours) the SSA value was 0.72±0.08 at RH 50% which however increases to 0.82 at 70%. The very low SSA value obtained after the foggy period was mainly due to the scavenging of scattering particles (which are also hygroscopic) such as sulfate and dust while the submicron soot particles, which are hygrophobic and contribute to the absorption, were less scavenged.

**Aerosol radiative forcing**

Aerosol radiative forcing (ARF) is calculated for the wavelength range 0.28-40 \( \mu m \) using the Santa Barbara DISORT Aerosol Radiative Transfer (SBDART) model (Ricchiazzi et al., 1998) at the top of the atmosphere (TOA), within the atmosphere and at the surface. The ARF values are estimated by running the radiative transfer code for an assumed aerosol free tropical atmosphere and with the specified aerosol components derived by model fitting the observed AOD spectra. Fig. 5.14 shows the diurnally averaged ARF values for 50% RH computed for the campaign days when AOD values are available. Prior to the foggy days the ARF values are comparatively small with surface forcing in the range of -10 Wm\(^{-2}\), and the TOA forcing in the range of 5 Wm\(^{-2}\) leading to an atmospheric forcing (difference between surface and the TOA forcings) of about 15 Wm\(^{-2}\). However during the foggy days (12, 13, 14 Dec) there is a large increase in ARF values with surface forcing in the range of, -20 to -25 Wm\(^{-2}\), TOA forcing in the range of 5 to 15 Wm\(^{-2}\) which give large atmospheric forcing in the range of 25 to 40 Wm\(^{-2}\). The large forcing values are mainly due to the increase in the aerosol number concentration and size during the foggy days as well as due to the relative increase in the abundance of absorbing particles due to poor ventilation. After the foggy days there is a reduction in the ARF values (Fig. 5.13) though they remained higher than the values obtained prior to the foggy days. In order to study effect of fog on the radiative forcing, the ARF values are computed on hourly basis, taking into account the solar zenith angle
Figure 5.14: Daily averaged aerosol radiative forcing over Hissar for those days when aerosol optical depth data are available.

variation, independently for the aerosol components obtained for the ‘before fog’ days, foggy days and after the foggy days. The individual aerosol components estimated for the 50% RH values are allowed to undergo changes for the RH values of 80% and 99% according to the OPAC model, and the new optical properties are used further to estimate the diurnal variation in the ARF values. While for low RH values the dependence on aerosol optical properties on the change in RH values can be neglected (more true for shorter wavelengths) at higher RH values even small changes in RH can cause large variation in the aerosol optical properties.

It is known that radiative forcing depends on solar zenith angle. On foggy
days however the RH value also undergo large diurnal variation from a morning high of more than 90% to an afternoon low of about 50% or less. Fig. 5.15a-c show the results of changes in the ARF values for both solar zenith angle and RH variations for the aerosol components estimated independently for the 'before fog' days, foggy days and after the foggy days. One of the main findings is that as the RH values become very high (shown for 99%) the TOA forcing becomes negative due to large backscattering of radiation to space as a consequence of changing SSA. Thus there is a corresponding decrease in the atmospheric forcing. Also the nature
of the diurnal variation depends on the relative abundances of different aerosol components, and in case of ‘foggy days’ the diurnal variation (Fig. 5.15b) is found larger as it is also evident from the low SSA values and high AOD values. Also, on foggy days, as the day progresses the decreasing RH value reduces the forcing value while the decreasing solar zenith angle increases the forcing values. Thus the effect of decreasing RH and the solar zenith angle have the opposite effect on the ARF, while the effect of latter is however found larger.

5.2.4 Aerosol Radiative Forcing during Land and Sea Breeze Conditions

Aerosol radiative forcing is calculated over Kalpakkam during different breeze conditions using the procedure described in Sec. 5.2.1. In the present study, the model derived AOD values are compared with the daily averages of the observed AOD values obtained from the Microtops during land breeze and sea breeze for the entire campaign. Using the observed AOD and BC mass concentration $\omega_0$ is also obtained from OPAC (Optical Properties of Aerosols and Clouds by Hess et al. (1998)) model. Model derived $\omega_0$ (0.5 $\mu$m) is 0.91 and the observed mean $\omega_0$ is 0.91±0.05 (Table 5.2). The variation is because of the the change in BC concentration at the measurement site caused due to change in wind direction. During sea breeze model derived $\omega_0$ is 0.93 which is same as observed. However, during 20-29 March when the wind is coming from the Indo-Gangetic basin (IGB) model derived $\omega_0$ decreased to 0.89, whereas the observed $\omega_0$ is 0.90. In Urban regions like Delhi and Kanpur in IGB $\omega_0$ is even smaller. Ganguly et al. (2006b) reported that $\omega_0$ is in between 0.60-0.80 with an average of 0.68 during December 2004 at Delhi. Singh et al. (2005) estimated SSA over Delhi to be 0.67 using the OPAC model. At Kanpur, another industrial region in IGB, $\omega_0$ was estimated to be 0.76 (Tripathi et al., 2005) and at Bangalore, another Urban region in southern India, $\omega_0$ was estimated to be 0.73 (Babu et al., 2002).

Fig. 5.16 shows the diurnal averaged SW ARF, calculated for individual days
and then grouped into three wind regimes coming from 1) Central BoB (16-19 March) 2) Indo-Gangetic Basin via BoB (IGB) (20-29 March) and 3) Arabian sea via North Indian Ocean (NIO) (30-31 March). During the study period a large negative forcing variation at the surface has been observed from -18.59 to -24.60 Wm$^{-2}$ whereas at TOA it varies from -5.76 to -7.37 Wm$^{-2}$. The surface forcing is -19.48 Wm$^{-2}$ for central BoB wind but the maximum forcing is estimated at -24.60 Wm$^{-2}$ for winds from IGB and minimum at -18.59 Wm$^{-2}$ for NIO. At TOA daily averaged forcing for central BoB is more negative (-6.95 Wm$^{-2}$) followed by IGB (-5.76 Wm$^{-2}$) and NIO (-7.37 Wm$^{-2}$). There is a small but significant difference in TOA forcing for IGB and NIO region. The maximum atmospheric forcing is estimated to be 17.23 Wm$^{-2}$ for IGB. At surface the aerosol radiative forcing intricately depends on many aerosol parameters such as columnar AOD, $\omega_0$ etc. During the NIO periods $\omega_0$ was large and columnar AOD was minimum, so the surface ARF is minimum. But during 20-29 March surface ARF is showing maximum of -24.60
Table 5.2: Aerosol parameters for Radiative forcing computation for three different wind regimes and OPAC derived modelled (Mod) parameters are compared with observed (obs), satellite (sat).

<table>
<thead>
<tr>
<th>Winds from</th>
<th>Period</th>
<th>AOD (0.5μm)</th>
<th>α_{0.38-1.02}</th>
<th>BC</th>
<th>ω₀</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Obs Sat Mod</td>
<td>Obs Mod</td>
<td>Obs Mod</td>
<td>Obs Mod</td>
</tr>
<tr>
<td>Central BoB</td>
<td>16-19 Mar</td>
<td>0.27 0.25 0.29</td>
<td>1.2 1.1</td>
<td>1.73 1.75</td>
<td>0.90 0.91</td>
</tr>
<tr>
<td>IGB</td>
<td>20-29 Mar</td>
<td>0.32 0.33 0.34</td>
<td>1.2 1.0</td>
<td>2.38 2.40</td>
<td>0.89 0.89</td>
</tr>
<tr>
<td>NIO</td>
<td>30-31 Mar</td>
<td>0.20 0.23 0.24</td>
<td>1.1 1.0</td>
<td>1.69 1.65</td>
<td>0.93 0.91</td>
</tr>
</tbody>
</table>

Wm⁻² due to large amount of BC that has been carried from IGB to the measurement site and hence at the surface more radiation is reduced due to enhancement of absorbing particles.

Satheesh et al. (2002) estimated the clear-sky short wave aerosol radiative forcing to be -38 Wm⁻² at surface, -7 Wm⁻² at TOA and 31 Wm⁻² in the atmosphere over BoB from the cruise experiment during March 2001. Ramachandran (2005) reported the forcing to be -57, 9 and 62 Wm⁻² respectively over Chennai during the same period. At the east coastal urban region the surface solar reduction is enhanced by a factor of 1.5 than the BoB and in the atmosphere the factor becomes 2. In the present study at Kalpakkam BC was mainly transported either from Chennai and its surrounding suburban region (during land breeze) or from Indo-Gangetic Basin via BoB and hence large variation in aerosol forcings is observed. Hence, depending upon wind direction at Kalpakkam the aerosol radiative forcing is different during different breeze conditions.

Sensitivity Test

Among all anthropogenic aerosols, BC plays a very major role in climate change by absorption of solar radiation, warming of the atmosphere and cooling of the surface (Andreae et al., 2005; Bond, 2001). The aerosol radiative forcings are estimated over Kalpakkam during March 2006 for different measured BC concentration to investigate the sensitivity of forcing by changing the BC concentration. In the present study there is a large variation in the diurnal BC variation. During the land breeze BC concentration is maximum at 3.0 μg.m⁻³ and during sea breeze it
Aerosol Radiative Forcing

Figure 5.17: Sensitivity tests for aerosol radiative forcing by varying the BC concentration when compared to the observed BC concentration under different conditions at Kalpakkam during March 2006. See text for more details. Top of atmosphere, surface and atmosphere aerosol radiative forcing are shown for BC equivalent to Continental Average (Test 5), Continental Polluted (Test 6), Urban (Test 7) and Marine Polluted (Test 8) for comparison.

is 1.2 $\mu g.m^{-3}$. However, during 20-29 March when wind was mainly coming from IGB, BC concentration increased to 2.0 $\mu g.m^{-3}$ during sea breeze and to 4.0 $\mu g.m^{-3}$ during land breeze. So as an input to SBDART code BC concentration was varied according to breeze condition. For the different conditions sensitivity test has been performed to the mean BC during land breeze (test 1), mean BC during land breeze when wind was mainly coming from IGB (test 2), mean BC during sea breeze (test 3) and mean BC during sea breeze when wind was mainly coming from IGB (test 4). Top of atmosphere, surface and atmosphere aerosol radiative forcings are also being studied for BC equivalent to Continental Average (Test 5), Continental Polluted (Test 6), Urban (Test 7) and Marine Polluted (Test 8) for comparison. The results of these tests are shown in Fig. 5.17.

In Fig. 5.17 the forcings vary over a wide range from -1.71 to -5.6 Wm$^{-2}$ at TOA. At the surface the forcing is always estimated to be negative and there is
Table 5.3: Aerosol radiative forcing by varying the BC concentration as observed BC concentration under different wind conditions at Kalpakkam. Test 5 to 8 are for OPAC model values

<table>
<thead>
<tr>
<th>Test</th>
<th>BC (μg/m³)</th>
<th>ω₀</th>
<th>Radiative Forcing</th>
<th>Condition</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>3.0</td>
<td>0.88</td>
<td>-6.2 -27.3 -21.2</td>
<td>BC equivalent to Mean LB</td>
</tr>
<tr>
<td>2</td>
<td>4.0</td>
<td>0.85</td>
<td>-5.1 -31.7 -26.6</td>
<td>BC equivalent to Mean LB during IGB</td>
</tr>
<tr>
<td>3</td>
<td>1.2</td>
<td>0.98</td>
<td>-9.2 -14.7 5.4</td>
<td>BC equivalent to Mean SB</td>
</tr>
<tr>
<td>4</td>
<td>2.0</td>
<td>0.97</td>
<td>-9.2 -15.1 5.9</td>
<td>BC equivalent to Mean SB during IGB</td>
</tr>
<tr>
<td>5</td>
<td>0.5</td>
<td>0.96</td>
<td>-8.8 -16.4 7.6</td>
<td>BC equivalent to Continental Averaged</td>
</tr>
<tr>
<td>6</td>
<td>2.0</td>
<td>0.91</td>
<td>-7.1 -23.3 16.2</td>
<td>BC equivalent to Continental Polluted</td>
</tr>
<tr>
<td>7</td>
<td>7.8</td>
<td>0.76</td>
<td>-1.7 -46.4 44.6</td>
<td>BC equivalent to Urban</td>
</tr>
<tr>
<td>8</td>
<td>3.1</td>
<td>0.97</td>
<td>-9.0 -15.6 6.5</td>
<td>BC equivalent to Marine Polluted</td>
</tr>
</tbody>
</table>

(LB: Land Breeze, SB: Sea Breeze, IGB: Wind from Indo-Gangetic Basin)

also a large variation depending on the BC concentration under different conditions because of large variation in ω₀. In test 1 and 2, TOA atmosphere forcings are negative and are -6.2 and -5.1 Wm⁻² and the surface forcings are found to be -27.3 and -31.7 Wm⁻², respectively. This implies that there is a reduction in surface solar radiation during land breeze due to enhancement of large amount of absorbing BC concentration. In test 2 wind carried large amount of BC from Indo-Gangetic Basin and surface aerosol radiative forcing became maximum during the campaign period. In test 3 and 4 also the TOA forcings estimated to be negative and are -9.2 and -9.2 Wm⁻², respectively. At surface they become -14.7 and -15.1 Wm⁻². During the sea breeze periods wind carried more scattering sea salt particles and less BC concentration, so at the surface there is less solar radiation reduction. During sea breeze conditions though background BC concentration was enhanced by a factor of 1.6 and ω₀ decreased very little from 0.97 to 0.98 the effect on radiative forcing at surface is significantly low. Therefore, the surface forcings are -14.7 and -15.1 Wm⁻², respectively, in tests 3 and 4. From the models comparison, surface forcings for test 5 (-16.4 Wm⁻²) and test 8 (-15.6 Wm⁻²) are similar to the sea breeze conditions (test 3 and 4). For test 6 surface forcing is -23.3 Wm⁻² and at TOA forcing is -7.1 Wm⁻². For test 7 BC concentration is assumed to be maximum at 7.8 μg/m³ equivalent to "urban" model. In this case SSA is a minimum of 0.76. At TOA forcing is found to be -1.7 Wm⁻² and at surface forcing is -46.4 Wm⁻².
In test 7 forcing is found maximum. It is clearly understood that enhancement of BC at Kalpakkam has great effect on $\omega_0$ and hence on the reduction of surface solar radiation and at TOA forcing becomes more positive as observed during land breeze.

5.2.5 Aerosol Radiative Forcing during Dust Storm at Mt. Abu

The impact of changes of background aerosol characteristics due to the dust storm at the hill top region has been quantified as their capability to perturb the radiation budget of the Earth's atmospheric system and is shown in Fig. 5.18 during (a) stormy days and (b) normal days. Aerosol radiative forcing has been calculated using the same procedure described in Sec. 5.2.1. The clear sky aerosol radiative forcing has been determined from difference of the solar radiation in no aerosol and with aerosol conditions in the short wave (0.25-3.9 $\mu$m), long wave (4.0-40.0 $\mu$m) and net (0.25-40.0 $\mu$m) regions. In the present study, 8-days MODIS derived Level 3 surface reflectance data during the observation period in May 2006 with the grid of 0.5 km for seven wavelengths centered at 0.47, 0.56, 0.65, 0.86, 1.24, 1.64 and 2.13 $\mu$m have been used for the estimation of radiative forcing over Mt. Abu. Spectral surface reflectance values are considered to be the combination of three different surfaces, viz., water, sand and vegetation, and are used as input to the SBDART model (Ganguly et al., 2006b).

Fig. 5.18 shows the short wave (SW), long wave (LW) and net radiation budgets at the background condition during (a) stormy days and (b) normal days. During the stormy days SW radiative forcing is 4.9 Wm$^{-2}$ at top of atmosphere (TOA) and -28.8 Wm$^{-2}$ at the Earth's surface. As a result the energy trapped in the atmosphere is 33.7 Wm$^{-2}$. The LW radiative forcing shows 4.2, 14.7 and -10.5 Wm$^{-2}$ at TOA, surface and in the atmosphere respectively. As a result the net forcing is 9.1 Wm$^{-2}$ at TOA, -14.1 Wm$^{-2}$ at the surface and hence 23.2 Wm$^{-2}$ in the atmosphere during the stormy days. During the normal days at TOA SW and LW forcing is 3.0 and 3.2 Wm$^{-2}$, respectively whereas at the surface they are -21.4, 12.2 Wm$^{-2}$
respectively. So in the atmosphere SW and LW forcings are 24.4 and -9.0 Wm\(^{-2}\), respectively. Therefore net forcings are 6.2 Wm\(^{-2}\) at TOA, -9.2 Wm\(^{-2}\) at surface and as a result in the net atmosphere forcing becomes 15.4 Wm\(^{-2}\). Pant et al. (2006) has reported the surface radiative forcing to be -4.2 Wm\(^{-2}\) and background atmospheric forcing was 4.9 Wm\(^{-2}\) during winter at Nainital, a high altitude station in central Himalayas. However, in the Himalayan region at Nepal the surface radiative forcing efficiency has been reported to be -73 Wm\(^{-2}\) with single scattering albedo in the range 0.7 to 0.9 due to large abundance of BC particles (Ramana et al., 2004). In case of western part of India during premonsoon the background surface radiative forcing is observed to be slightly higher than the earlier reported background forcing in Himalayan region. The reason is that the adjacent ‘Thar desert’ helps to enhance the background dust particles and urban regions of western part of India add BC particles in the background region. Due to more dust...
and BC aerosols, more radiation is trapped in the atmosphere in SW region during the stormy days whereas the steady background fine dust aerosols trap almost the same LW terrestrial radiation in the atmosphere during the observation period. As a result the net forcing at TOA is slightly enhanced in comparison to normal days due to more reduction of radiation at the surface.

5.2.6 Aerosol Radiative Forcing over Arid Region

**OPAC estimation of AOD and Aerosol Mass Comparison**

Aerosol optical properties strongly depend on the chemical properties of aerosols. Hence it is necessary to have the knowledge of chemical properties of the various types of aerosols. However, in the absence of chemical composition information, AOD spectrum (0.25-40.0 μm) can be derived from the Optical Properties of Aerosols and Clouds (OPAC) (Hess et al., 1998) using the known chemical compositions of aerosol over western part of India (Ganguly et al., 2006a; Rastogi and Sarin, 2005a). The modelled AOD spectrum should match the observed AOD values. Detail procedure is discussed by the earlier researchers (Ramachandran et al., 2006; Ganguly and Jayaraman, 2006) and hence not discussed here. The OPAC estimated AODs and observed AODs and the angstrom parameters match quite well with each other and are given in Table 5.4.

<table>
<thead>
<tr>
<th>Station</th>
<th>AOD</th>
<th>$\alpha_{0.38/1.02}$</th>
<th>Total Mass</th>
<th>BC</th>
<th>Model SSA</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Obs</td>
<td>Sat</td>
<td>Mod</td>
<td>Obs</td>
<td>Mod</td>
</tr>
<tr>
<td>Ahmedabad</td>
<td>0.31</td>
<td>0.34</td>
<td>0.31</td>
<td>0.50</td>
<td>0.51</td>
</tr>
<tr>
<td>Udaipur</td>
<td>0.30</td>
<td>0.28</td>
<td>0.29</td>
<td>0.46</td>
<td>0.51</td>
</tr>
<tr>
<td>Mt. Abu</td>
<td>0.28</td>
<td>0.27</td>
<td>0.28</td>
<td>0.75</td>
<td>0.73</td>
</tr>
</tbody>
</table>
Single Scattering Albedo

Single scattering albedo (SSA) is an important parameter in the radiative forcing calculation. During April, dust storm is a common phenomena in the western part of India (Deepshikha et al., 2006a). Due to the frequent dust storm occurrence the dust abundance in the atmosphere increases. Using chemical transport model, "Goddard Global Ozone Chemistry Aerosol Radiation and Transport" (GOCART) model (e.g., Chin et al., 2003) and AERONET (e.g., Holben et al., 1998) at 0.550 µm, SSA was found to be 0.87-0.93 over western part of India during April (Chung et al., 2005). But the SSA of dust strongly depends on the presence of BC concentration in the atmosphere. SSA of dust varies form 0.9-0.98 depending on the amount of BC present in the atmosphere and it decreases with increase in BC concentration (Chung et al., 2005). Another satellite observation shows that infrared SSA over this region varies in the range 0.88-0.94 due to the moderate absorbing dust aerosols in the atmosphere (Deepshikha et al., 2005, 2006a,b; Moorthy et al., 2007).

In the present study SSA is estimated using OPAC model for Ahmedabad, Udaipur and Mt. Abu. During April, SSA is 0.83 over Ahmedabad and 0.87 over Udaipur, whereas over Mt. Abu SSA is 0.88. From simultaneous measurements of single scattering coefficient using Nephelometer and calculated absorption coefficient from Aethalometer observations, SSA has been calculated and reported to be 0.84±0.04 over Ahmedabad during premonsoon (Ganguly et al., 2006a). Over Ahmedabad SSA becomes minimum compared to other stations due to large production of anthropogenic BC and the presence of dust particles. At the hilltop station Mt. Abu, SSA is high due to less concentration of BC. At Udaipur SSA is slightly less due to increase in dust particles concentrations.

Aerosol Radiative Forcing

The aerosol radiative forcing is computed using the same procedure as described in Sec. 5.2.1. A comparison of the observed and the model values are given in
Figure 5.19: Spectral Surface Reflectance over Ahmedabad, Mt. Abu and Udaipur during April 2007 that has been used in the present study to compute the aerosol radiative forcing.

Table 5.4. The clear sky aerosol radiative forcing has been determined from difference of the solar radiation in no aerosol and with aerosol conditions in the short wave (0.25-3.9 \( \mu \text{m} \)), long wave (4.0-40.0 \( \mu \text{m} \)) and net wave (0.25-40.0 \( \mu \text{m} \)) regions. The spectral surface reflectance values are considered to be a combination of three different surfaces, \textit{viz.}, water, sand and vegetation as shown in Fig. 5.19.

Fig. 5.20 shows the mean aerosol radiative forcing over Ahmedabad, Udaipur and Mt. Abu at the top of atmosphere (TOA), Earth’s surface and in the atmosphere computed for individual days. During April the SW radiative forcing obtained at the surface over Ahmedabad is maximum at -44.3 ± 5.2 Wm\(^{-2}\). The forcing is also maximum at TOA over Ahmedabad and is 9.0 ± 1.0 Wm\(^{-2}\). The surface forcing is found to be governed by the magnitude of the AODs, presence of BC absorbing particles and hence SSA. But the TOA forcing is very intricately dependent on several other aerosol parameters, like scattering and absorbing particles concentration, surface albedo etc., in addition to AOD, SSA. The difference between
Figure 5.20: Aerosol radiative forcing over Ahmedabad, Udaipur and Mt. Abu in the (a) shortwave (SW) and (b) longwave regions (LW) and (c) the Net forcing (SW+LW).
the surface and TOA forcing gives the atmosphere forcing which indicates the energy trapped in the atmosphere. Earlier workers have shown that the TOA and surface forcings are 8.0±2.0, -41.4±5.0 Wm⁻², respectively, at Ahmedabad during premonsoon (Ganguly et al., 2006a). The difference between surface and TOA forcing gives the atmospheric forcing which indicates the energy stored in the atmosphere. At Ahmedabad the SW atmospheric forcing is a maximum of 53.3±5.7 Wm⁻² which represents the large SW radiative heating in the atmosphere. At Udaipur the forcings are 5.1±1.4 Wm⁻², -32.2±6.3 Wm⁻² and 37.3±7.5 Wm⁻² at TOA, surface and in the atmosphere, respectively. At Udaipur the SW radiative heating in the atmosphere is moderate as less amount of SW radiation is trapped. This is due to the presence of large abundance of scattering particles like dust.

In the background environment at Mt. Abu, the forcings are 3.5±0.1 Wm⁻² and -27.3±0.8 Wm⁻² and 30.8±0.9 ± at TOA, surface and in the atmosphere, respectively, during April. The forcings are minimum due to the presence of less anthropogenic BC aerosols at this high altitude.

In the LW region soot is a perfect absorber and hence its single scattering albedo becomes zero whereas dust particles contributes to more scattering than the other aerosol components (Hess et al., 1998; Satheesh and Lubin, 2003). Dust is an efficient scatterer in the 4.0-6.0μm range and beyond that starts exhibiting absorbing properties also. It has an absorption peak at ~6μm and in the longer wavelength (>11.0μm) dust exhibits absorbing properties. This indicates that any variation in the amount of dust particles alone can affect the longwave terrestrial radiation and hence the LW aerosol radiative forcing. In addition to dust, the trace gases ozone and water vapor also have an impact on the LW radiative forcing. Ozone causes warming in the lower stratosphere whereas in upper stratosphere it helps to cool the atmosphere associated with water vapor and carbon dioxide in the LW radiation (Ramaswamy, 2002). LW radiative forcings are also shown in Fig. 5.20 for Ahmedabad, Udaipur and Mt. Abu. The forcings are comparable within ±1σ at Ahmedabad and Udaipur due to similar water vapor and ozone concentrations and dust aerosols. The variation of LW forcing at Udaipur is large due to large
Aerosol Radiative Forcing

variation in dust aerosols. The forcings at Mt. Abu are slightly low due to the low abundance of water vapor and transported dust particles at this high altitude. The net (SW+LW) forcings are maximum at -28.8 and 43.6 Wm$^{-2}$ at surface and in the atmosphere, respectively, at Ahmedabad followed by Udaipur at -17.5 and 27.8 Wm$^{-2}$, respectively. The minimum forcings are -14.6 and 22.3 Wm$^{-2}$, respectively, at Mt. Abu due to its background conditions. At Udaipur net forcing is moderate due to less anthropogenic SW forcing at the surface.