The Direct Effect of the Anthropogenic Aerosols on the Winter Climate of South Asia: Focused on the Model Produced Climate Features

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Abstract

The direct effects of anthropogenic aerosols on the winter climate of South Asia (SA) are investigated in this study. The study is based on two eight-year simulations, with and without the aerosols, by the Regional Climate Model (RegCM4) coupled with the aerosol model. There is generally a good agreement between the observed and simulated climatology for the surface air temperature and precipitation. The model captures very well the observed and regional patterns of both temperature and precipitation; in particular, the spatial gradients and the topographically induced structures. However, the model over-estimated the precipitation over the mountainous region. The Emission Database for Global Atmospheric Research (EDGAR) data shows that, SO_2 emission is up to 6.5×10^{-10} ($mg^{-1}.m^{-2}$.day) over the mega cities of India whereas the emission of black carbon (BC) and organic carbon (OC) is nearly equal to 1×10^{-12} ($mg^{-1}.m^{-2}$.day) over the northern parts of India. Furthermore, the column burden of SO_2 is up to 1.3 ($mg^{-1}.m^{-2}$.day) and the BC and OC is up to 0.0035 ($mg^{-1}.m^{-2}$.day) in the eastern parts of India. The low level wind flow both in NCEP and RegCM4 is towards the Indian Ocean in most part of the domain. The aerosols induced changes in the surface radiative forcing (SRF) is negative over the whole domain and is up to -8 ($W.m^{-2}$) in the east and south coast of India, whereas, the temperature near the surface is also decreased up to 0 ºC in the eastern parts of India. However, the relative humidity (RH) increased up to 0.5 % in the eastern and southwestern part of India. The aerosols indirectly affect the large scale circulation, near the surface the northerly winds strengthen with higher geo-potential which strengthens the winter monsoon over the Sri Lanka whereas the winds become weaker in the upper atmosphere.

Key Words: Anthropogenic aerosols, RegCM4, South Asia, Monsoon, Atmosphere.

Introduction

Aerosols are tiny solid particles or liquid droplets suspended in the atmosphere. They have the ability to absorb and reflect the solar radiations which otherwise would reach to the Earth surface, which can affect the Earth's radiation budget. Changes in the radiation budget can also influence the hydrological cycle of the Earth (Ramannathan et al, 2005; Ramannathan et al, 2001). There are numerous types of aerosol sources but the main categories are (i) Natural sources (ii) Anthropogenic Sources. Desert dust, sea spray, volcanic, biogenic and organic emission belongs to the natural sources, while burning of fossil fuel and bio fuel belongs to the anthropogenic sources. The chemical composition of the aerosols tells us that in South Asia (SA) roughly 75 % of the aerosols are from an anthropogenic origin (Ramanathan et al, 2001; Lelieveld et al, 2001). Among these aerosols, sulphate and nitrate has cooling effect while black carbon has warming effect (Kaleem, 2016).

The South Asian region is a highly populated and polluted region of the world. The main emissions of aerosols in the rural areas are from bio fuel burning, however, these emission are difficult to estimate since it is emitted from the large rural areas of Pakistan, India, Bangladesh and Nepal. Also, Lelieveld et al, (2001) mentioned that the burning phase is not well-defined due to the different fuel type and combustion processes (i.e. smoldering or flaming).

The energy consumption of the people living in the rural areas of the India, Pakistan and Bangladesh depends upon bio fuel; however, the urban population is using the soft coke, kerosene and other type of

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liquid fuels. In Asia as a whole, one quarter of the population is using bio fuel and in India the amount is close to 50 % (Araújo et al, 2017).

The aerosols effects on the atmosphere are classified into four type's i.e. direct effects, semi-direct effects, indirect effects of the 1st kind and the indirect effect of 2nd kind. In the direct effect, aerosols act to block the sunlight and cool the surface (i.e. solar dimming) but in the semi- direct effect these aerosols increase the stability and reduce convection. The 1st indirect effect involves aerosol particles acting as a Cloud Condensation Nuclei (CCN), increasing the amount of cloud droplets and thus enhancing the reflection of solar radiation by the clouds (i.e. cloud albedo effect). Aerosols, particularly sulphate aerosols, provide the necessary nuclei for cloud drops and ice crystals. Larger cloud droplets at the same cloud liquid water increases the cloud albedo and this will lead surface cooling (Kaleem, 2016). Smaller droplets results in the longer life time of the clouds; this is the so-called second indirect effect, which enhances the cooling due to the first indirect effect. This increases atmospheric solar heating in atmospheric thermal structure and change the regional circulation system, such as suppression of rainfall, less efficient removal of pollutants and also affects the monsoons season. The anthropogenic aerosols are dimming the earth surface, while making it brighter from the top of the atmosphere (TOA). Both the direct and indirect effect has a net radiative forcing (i.e. the alteration in the atmospheric radiative heating by the absorption and scattering of the solar radiation is called radiative forcing) at the TOA. Over the tropical Indian ocean is approximately -5(±2) W.m-2 whereas the major finding from Indian Ocean Experiment (INDOEX) showed the net radiative forcing at the surface was four times larger than the TOA forcing -20(±3) W.m-2, while the atmospheric radiative forcing was three times larger +15(±3) W.m-2 (Ramanathan et al, 2007).

The transport of the aerosols (e.g. sulfur dioxide, nitrogen dioxide and organic compounds) is important since these aerosols can travel thousands of kilometers away from their sources due to atmospheric circulation. The atmospheric aerosols, especially particulate matters with a diameter less than 2.5 µm, can be transported hundreds of the kilometers downwind from their sources and can contribute to regional and urban haze. Most of the pollutants build up over South Asia (SA) and then flow out over the Indian Ocean. This happens mostly in the dry winter which is a conducive period for the accumulation of air pollutants due to the cloud free situation (Ramanathan et al, 2007).

The SA is the region extending over Afghanistan, Pakistan, India, Nepal and Bangladesh. This region has particular complex morphology (Giorgi et al, 2001, Giorgi and Bi, 2005) due to the high terrain of the Himalayan region to the North. Therefore, high resolution modeling would be a good tool to investigate climatic changes due to aerosols over the region. It is, however, important to first test the ability of the Regional Climate Model (RCM) to regenerate the climatic features over the SA. A few studies have been conducted using RCM's for the aerosol climatic changes over the Asian region but not particularly over the SA. There are various scientific studies on the aerosols by using different kind of global and regional models. Daniel et al, (2004) used the Model for Ozone and related chemical Tracers Version 2 (MOZART-2), a global model developed at National Center for Atmospheric Research (NCAR), to study the seasonal variability of secondary organic aerosols (SOA) globally. They found that maximum SOA concentrations were found over the Northern Hemisphere and tropics in June, however, the SOA concentration remains higher in tropics until December.

Zanis, (2009) studied the direct effect of aerosols on near surface air temperature over Southeastern Europe during the summer 2000 based on regional climate modeling. This study showed that the aerosols induced changes in the air temperature from the lower to upper troposphere is not correlated with the surface radiative forcing due to the complex mechanism of the aerosol radiative forcing linked with the induced changes through the dynamical effects of aerosols on the atmospheric circulation.

Wu et al, (2008) simulated the direct effect of the black carbon aerosol on temperature and hydrological cycle in Asia with a regional climate model. The major finding from this experiment was that the effect on the surface radiative forcing was negative and its absolute value was larger than the TOA. In this study, the direct effect of anthropogenic aerosols on the winter climate of SA is studied while the indirect effect and natural aerosols sources are excluded. The change in the temperature, precipitation, relative humidity and large scale circulation over the South Asia due to the inclusion of anthropogenic aerosols in a regional climate model is studied.

Climate of the South Asia

The region of the SA comprising the countries Pakistan, India, Nepal, Bangladesh and Sri Lanka which have unique characteristics of the ge`ographical location, complex physiographic pattern, large mountain ranges, large rivers basins, plateaus and long coastlines extending over the Bay of Bengal, Indian ocean and Arabian sea. The domain of the SA as defined here is 6ºN to 38ºN and 60ºE to 95ºE with complex topography over the region, as shown in Figure (1).

The climate of the SA consists of four distinct seasons; winter (Dec-Feb), spring (Mar-Apr), summer (May-Aug) and autumn (Sep-Nov). However, there are various seasonal meteorological phenomena among these four seasons. In winter, the SA has a dominating influence from continental air from west with low humidity generally causing the dry weather over most of the subcontinent.

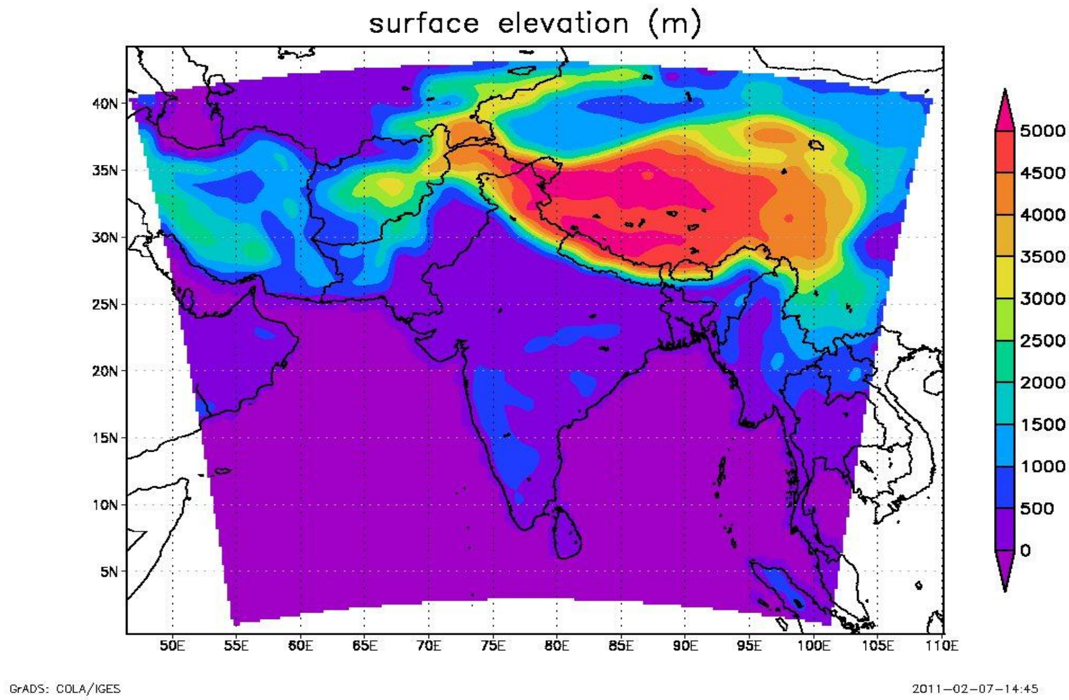


Figure 1: Model domain and topography (meters) over the South Asia.

Western disturbances from mid-latitude cyclones over the West Asia affect the South Asia north of 30ºN giving rise to cloudiness and precipitation. Weak cyclonic circulation often develops over the central parts of Pakistan and Rajasthan and move towards the east- northeast. These western disturbances mostly have occluded fronts, instead of clear warm or cold fronts. In winter, heavy precipitation occurs in the Northern Mountains of Afghanistan, Pakistan and India due to these western disturbances feeding of moisture from the Arabian Sea and Bay of the Bengal. SA is dominated by a tropical monsoon climate. The rainfall over the region is subject to a high degree of spatial and temporal variability with different climatic zones ranging from the arid to the tropical rainforest and occurrence of devastating floods and droughts. The distribution of the seasonal rainfall (i.e. summer monsoon and winter monsoon) has a primary significance in delineating the climatic regimes within the region. The climate of the region from North of Pakistan to Sri Lanka in the south traverses from the eastern edge of the Sahara desert to the equatorial tropical climates. The great mountains of the Himalayas effectively stop the mid -latitude climatic effect of the Central Asia and play an important role for the unique characteristics of the South Asia. For example there are fluctuating glacier margins from the Southern slope of the Himalayan ranges and shifting sand storms in the Thar Desert in Sindh. The dramatic change in the surface temperature from -50ºC in Himalaya during winter to 50ºC in the western part of SA in summer is most fascinating. Similarly, in winter snow storms frequently hit the western and central Himalaya extending from the area of Karakoram ranges in northwest of Pakistan to the world's highest mountain in northeast India and Nepal. However, in the summer northwest Thar Desert (i.e. extension of Saharan desert) experiences the frequent dust storms. These climatology facts justify the importance of the region in the world (Pant et al, 1997).

Model Description, Data and Experiment Design

Model Description

The regional climate model used for this research is RegCM version 4. The RegCM4 model is a primitive equations, hydrostatic model on a sigma level vertical coordinate system based on the NCAR-PSU Mesoscale Model version 5 (MM5) (Grell et al, 1994), using the mass flux scheme by (Grell et al, 1993).

Large-scale precipitation and clouds are calculated with the Sub-grid Explicit Moisture Scheme (SUBEX) (Pal et al, 2000). Ocean surface fluxes are calculated using the scheme proposed by the Zeng et al, (1998), while land surface fluxes are calculated using the BATS scheme (Dickinson et al, 1989). The radiation code is taken from the Community Climate Model version 3 (CCM3) (Kirkevåg et al,2002).

The radiative transfer scheme takes into account O_3, H_2O, CO_2 and O_2 for solar radiation linked with the 19 spectral intervals from 0.2 and 5 µm of the б-Eddington approximation of Kiehl et al, (1996). The surface radiation scheme of the Community Climate Model versions 1 (CCM1) is taken by Dickenson et al, (1986).The scattering and absorption parameter of the cloud follows Slingo et al, (1989). Coupled to RegCM is an aerosol model here used to investigate aerosol feedbacks on the climate (Solmon et al, 2006). This model also included the direct shortwave effect of aerosols.

Data

The Sea Surface Temperature (SST) data is taken from a 1.0 × 1.0 degree global gridded dataset of NCEP Climate Prediction Center, following the technique of Reynolds and Smith (1994). Land use and topography data is obtained from the 10-min United States Geological Survey (USGS) and Global Land Cover Characterization (Loveland et al, 2000) datasets, respectively. The initial and the 6-hourly lateral boundary conditions were from the NCEP reanalysis (Kalnay et al, 1996), at a 2.5×2.5 degree resolution.

For the model evaluation we use the TS3.0 datasets of climate observations from the CRU (Climatic Research Unit) for the period 1901-2006 taken by the British Atmospheric Data Centre (BADC). This data has a 0.5×0.5 degree resolution including monthly surface air temperature and precipitation. These data sets are revised and extended over the original data set of (New et al, 2002).

Limitations in this study include: First, the aerosol indirect effect in the cloud properties is not included in the simulation. While the (IPCC, 2007) predicted that the first indirect effect is larger than the direct effect over the globe, it is uncertain. The second indirect effect may be as large as first indirect effect but the uncertainties here are even larger (Lohmann and Feichter, 2004). Second, we do not consider natural aerosols emission, such as dust and sea-salt particles. This study is thus mainly focused on the effect of anthropogenic aerosols.

Experiment Design

The spatial resolution of the model is 50 Km with (112×95) grid points on 18 vertical levels from the surface to 100 hPa. The model is integrated from 1st January 1990 to 30th December 1998. The first year is taken as spin up and the results of the 8 winter seasons are averaged and analyzed here. Two experiments (i.e. Control run and Aerosol run) have been performed using the same set-up with the exception of including the aerosol in one of the experiments. The aerosol model has six prognostic species (i.e. SO2, SO4, black carbon hydrophobic and hydrophilic, organic carbon hydrophobic and hydrophilic) as described by the Qian et al, (2001). The SOx (SO2 +SO4) emissions were taken from Emission database for Global Atmospheric Research (EDGAR) on a 1.0×1.0 degree resolution. EDGAR accounts for the fossil fuel, biomass burning, industrial chemical production, bio fuel and waste treatment. The calculations for the emission of the BC and OC particles were taken from by the Cooke et al, (2002) also on a 1.0×1.0 degree resolution. The results of the two experiments are compared, particularly the direct radiative forcing of the aerosols and the consequences for the atmospheric circulation.

Results and Discussion

Model Evaluation

Before evaluation of the RegCM4, the accuracy of the model must be evaluated for regional climate characteristics. The evaluation of the model performance by using lateral boundary conditions should be established for determining the weakness and strength of the model over the SA. In this section, the consistency between the CRU observation and model output is evaluated in this section.

Temperature

There exist a good agreement between the observed and simulated run of the air surface temperature as shown in Figure 2(a) and 2(b). The topographically induced structure and spatial gradients are well captured and simulated values differ within few degrees of the observed values. The average winter temperature difference of CRU and Control run in the latitudinal bands (23°N to 36°N) is approximately up to 1°C.

In Afghanistan the CRU observational temperature is -3°C in north and 9°C in south. In Pakistan the CRU temperature is -3°C in north and between 6 to 15°C in south and southeast part, respectively. However, in the India the CRU observational temperature is between -3 to 12°C in the north part and 18 to 22°C in south part. In Nepal and Bangladesh the CRU temperature is -3°C in the north of Nepal and between 18°C to 20°C in the whole Bangladesh respectively.

The Control run differs from CRU in the southern part of India, where the differences for the winter temperature can be 1 to 2°C. Although the model performed well but some persistent biases such as the cold bias over the South part of Pakistan, India and Bangladesh can be seen.

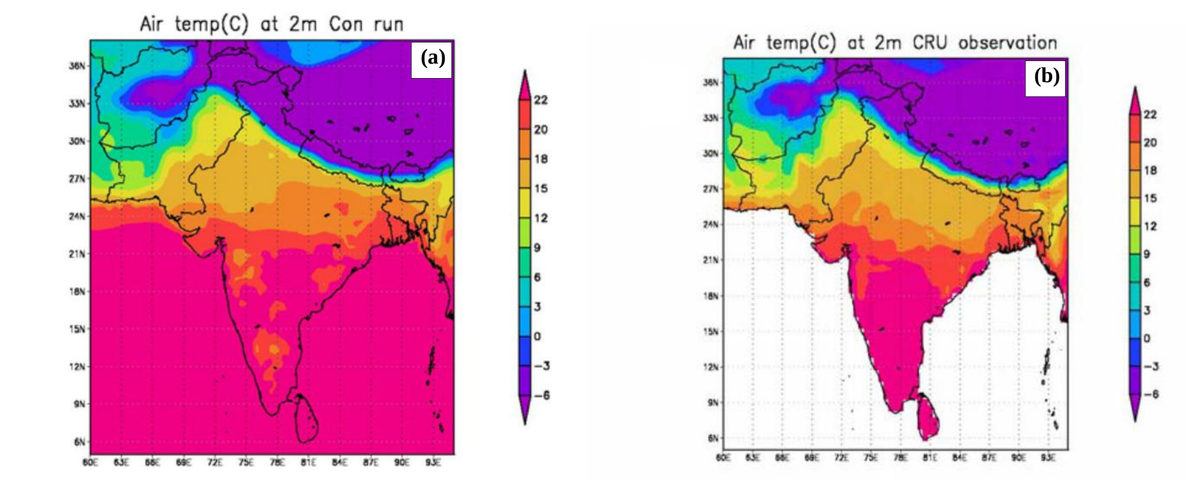


Figure 2: Mean (1991-1998) DJF 2 m air temperature °C **(a)** Con run **(b)** Cru

Precipitation

The RegCM4 Control run also captures the December-January-February (DJF) precipitation pattern well. The width of the rain belt and location of the maxima are well represented but the amount is over-estimated in the north part of Afghanistan, Pakistan and northern India. There are small differences while comparing the Figures 3(a) and 3 (b) of the Control and CRU observation.

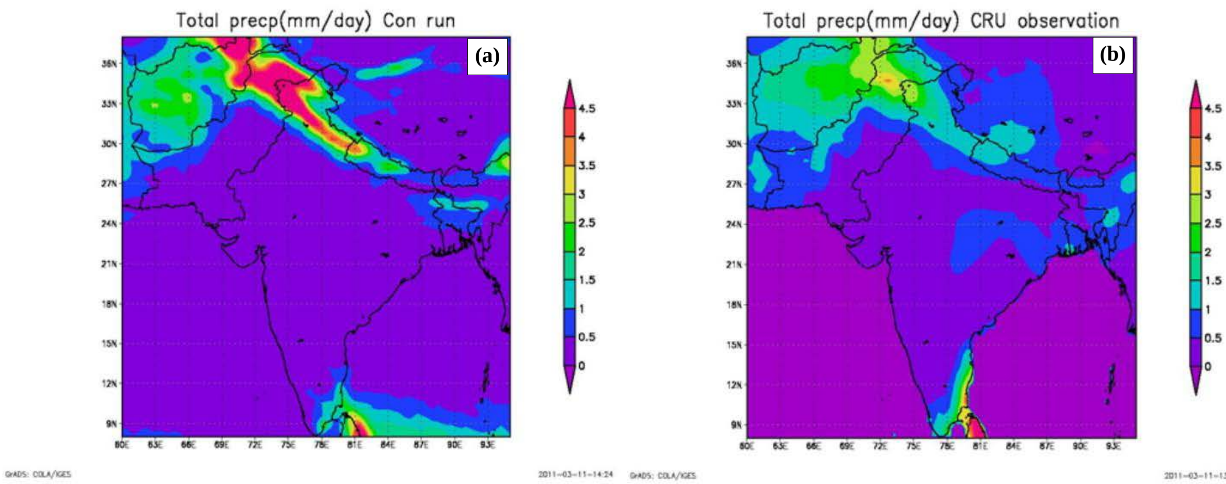


Figure 3: The mean (1990-1998) DJF precipitation (mm.d-1) **(a)** Con run **(b)** Cru observation

The Control run is showing more than 4.5 (mm.d^{-1}) precipitations over the north part of Afghanistan, Pakistan and India and over the Himalayan range, while the CRU observation shows 2 to 3 (mm.d^{-1}) precipitations over the same area. There is one additional area in the CRU observation near the East part of India (i.e. 18ºE to 24ºE latitudinal band) where precipitation is slightly over-estimated by up to 0.5 (mm.d^{-1}) in the winter season. The simulated values have higher precipitation than the observation values, however the CRU observation can also be underestimating the actual precipitation due to a low-elevation station bias and the gauge under-catch bias. Adam at al, (2003) explained that the precipitation can be underestimated by up to 30 % in the winter season. This discrepancy might therefore be due to both the lack of observation and a model over-estimation over the mountainous area of Afghanistan, Pakistan and India.

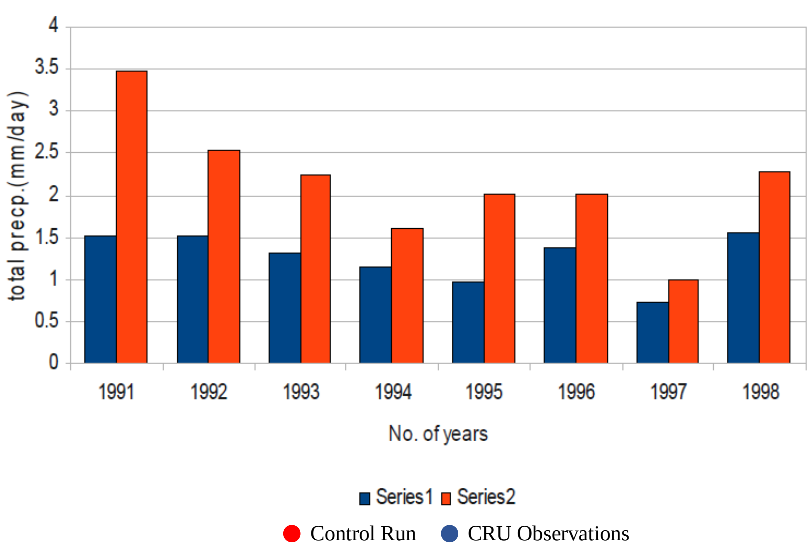


Figure 4: Seasonal averages (DJF) of total Precipitation (mm.d^{-1}) from 1990-1998,
Series1: CRU Observation, Series2: Control run

Figure (4) shows is a column chart for the Control run and CRU observation for total precipitation (mm.d^{-1}) for the winter season in individual years. There are certain differences between these two especially in 1991, 1992, 1993, 1995 and 1998 winter season. Figure (4) also shows that the variation is coming from years to years and has the lowest precipitation about 1 (mm.d^{-1}) in 1997. In that year, the observation data showed the lowest value of about 0.75 (mm.d^{-1}). While the model showed higher values of precipitation in 1991 about 3.5 (mm.d^{-1}), in 1992 about 2.5 (mm.d^{-1}), in 1993 about 2.2 (mm.d^{-1}), in 1995 about 2 (mm.d^{-1}) and in 1998 about 2.2 (mm.d^{-1}), and the observations are also higher for those years except for 1991. This variation in the model can be due to the higher elevation terrain of the region. However, this inter annual-variability can also be predicted by the interaction of the atmospheric circulation with this higher terrain region (Adam et al, 2003).

Some other studies have also showed that model is biased while dealing with the precipitation over the South Asia. Syed et al, (2009) simulated the winter climate over the Central-South Asia, with the emphasis on NAO and ENSO effect using RegCM3 and found larger biases in precipitation as compared to temperature. Kieu et al, (2004) also simulated the Southeast Asia rainfall using the RegCM3 model and found an underestimation of both precipitation and temperature in this model.

Mean flow

Since the observational datasets do not include wind speed, we evaluate RegCM4 against the NCEP reanalysis data. These results should be interpreted knowing that this is the same data as used for the later boundary condition. The mean flow at 1000 hPa in RegCM4 Figure 5(a) is very consistent with that in NCEP in Figure 5(b). In both, the winds are blowing from the Bay of Bengal (i.e. Northwest coast of India), which causes the winter monsoon over the Sri Lanka. Winds over land are lower than over the ocean and the flow is from northeast to the southwest. At 700 hPa, the winds are westerly with air flowing from west to east both in NCEP and in RegCM4, as shown in Figures 5(c) and 5(d).

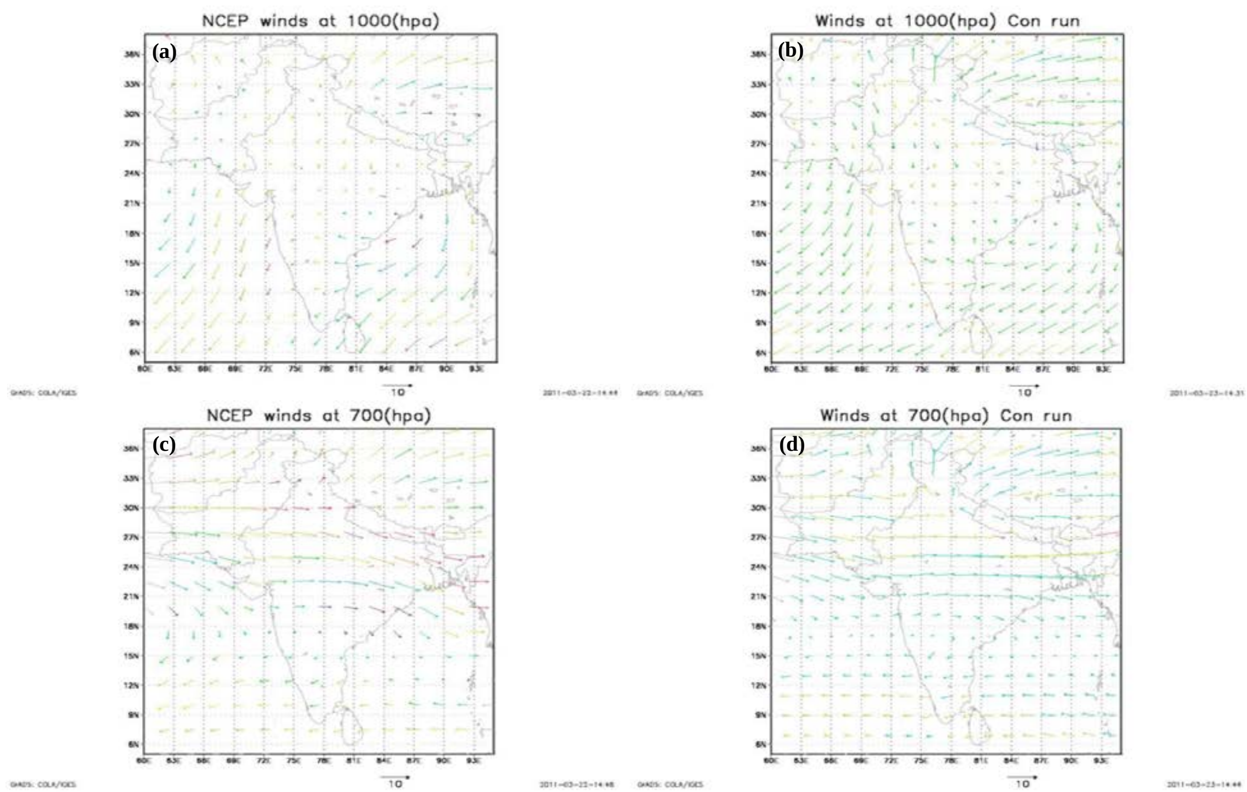


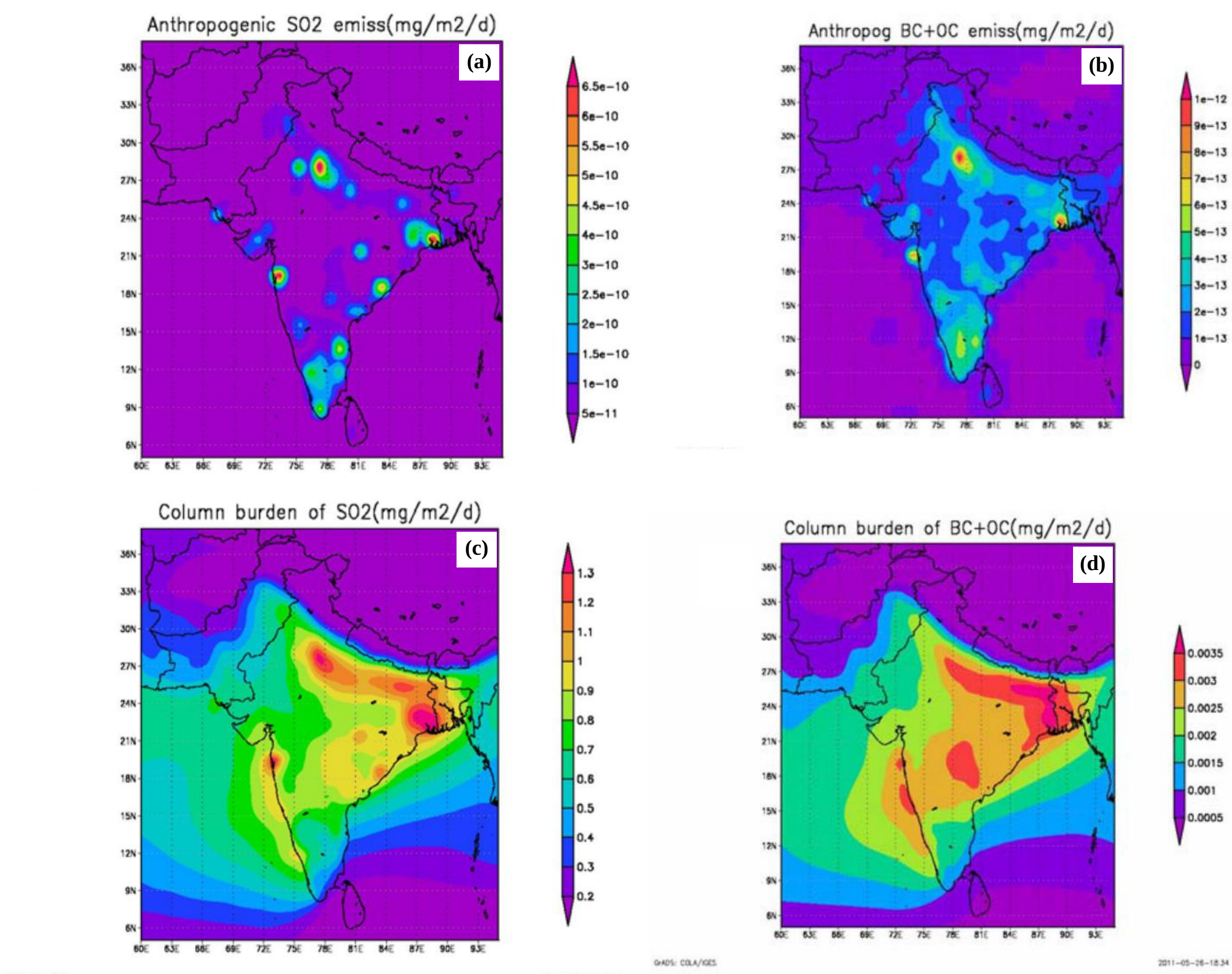
Figure 5: The mean winds (m/s) from 1990-1998 **(a)** at 1000 hPa NCEP **(b)** at 1000 hPa RegCM4 **(c)** at 700 hPa NCEP **(d)** at 700 hPa RegCM4

Certain meteorological patterns govern the transport of the aerosols over the South Asia. The DJF winter time lower-level wind over South Asia generally blows from the northeast to southwest. In this dry season there is little deep convection in the air flowing over the South Asia. So these air masses encounters the large clouds free Northern Indian Ocean down to the equator which helps the accumulation of pollutants. These pollutants build up due to photo-chemically generated ozone in the presence of strong solar radiations and in the absence of precipitation.

These offshore winds transport the aerosol plume over thousands kilometers down to the ITCZ. Most of the pollutant outflow is observed over the Indian Ocean in elevated layers between 1 and 3 Km. It has been proposed that the outflow is affected by the land-sea breeze circulation that is present over the west coast of India (Leon et al, 2001; Mohanty et al, 2001; Lelieveld et al, 2001; Raman et al, 2002).

Anthropogenic Aerosols Surface Emission and Surface Radiative Forcing

The Figures 6(a and b) below shows the spatial distribution of emissions of the anthropogenic aerosols from the EDGAR emission dataset of SO_2 , BC and OC over the South Asia (SA). Higher emission of SO_2 are found over the mega cities of India i.e. New Delhi, Pune, Bangalore and Kolkata, up to 6.5×10^{-10}($mg^{-1}.m^{-2}.day$). Some emissions are also found in Pakistan (i.e. Karachi and Lahore). Black carbon and organic carbon emissions are more equally distributed over the whole of India but higher values up to 1×10^{-12}($mg^{-1}.m^{-2}.day$) can be found over the New Delhi and Kolkata. Some concentration of BC and OC is also emitted by the Pakistani cities i.e. Karachi, Punjab and also in Bangladesh (Dhaka).



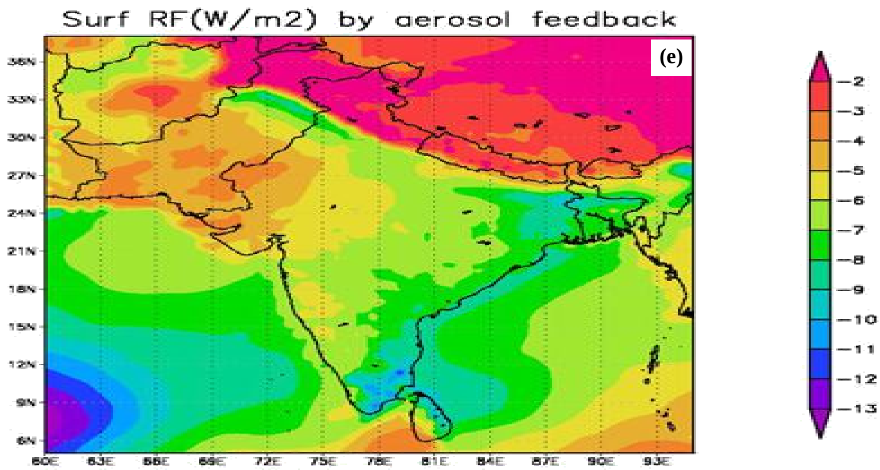


Figure 6: Anthropogenic emission of **(a)** SO_2 ($mg^{--1}.m^{-2}.d$) EDGAR emission **(b)** the sum of black and organic carbon ($mg^{-1}.m^{-2}.d$) by EDGAR emission **(c)** The SO_2 column burden ($mg^{-1}.m^{-2}.d$) RegCM4 (4) the black and organic carbon column burden ($mg^{-1}.m^{-2}.d$) RegCM4 **(e)** the mean (1990-1998) DJF winter surface radiative forcing ($W.m^{-2}$) due to the anthropogenic aerosols obtain from the RegCM4/aerosol simulations.

The aerosols column burden of SO_2, black carbon (BC) and organic carbon (OC) by RegCM4/aerosol mode, as shown in Figures 6(c) and 6(d). The SO_2 concentration up to 1.3 ($mg^{-1}.m^{-2}.day$) can be observed over the eastern side of India which can be due to the industrial area in the region. Similarly, the BC and OC column burden is up to 0.0035 ($mg^{-1}.m^{-2}.day$) in the same eastern side of India. Both Figures 6(c and d) showed higher concentration of aerosols over land as compared to oceans. These plots are consistent with the anthropogenic emission plots of SO_2, BC and OC.

Figure 6(e) shows the mean winter surface radiative forcing (SRF) obtained from the RegCM4/aerosol simulation. The SRF is negative as expected due to the solar dimming effect of the anthropogenic aerosols. These negative values can be seen throughout the SA domain with more negative value, down to -13W.m^{-2}, over the ocean. Northern of Pakistan, India and Nepal has SRF values up to -2 W.m^{-2} but the average value over the India is about -8 to -10 W.m^{-2}. This means that as the concentration of the aerosols increases in the atmosphere it will dim the surface of the Earth. Ramanathan et al, (2007) also has similarly explained the average accelerated trend of the solar dimming over the South Asia was -8W.m^{-2}, which is close to the results concluded in this study. Zanis et al, (2009) showed larger negative values as compared to this study, but that study was only concerned with one single season and was conducted over the southeastern Europe.

Changes in Air Temperature, Relative Humidity and Water Vapour Mixing Ratio due to Anthropogenic Aerosols

Figure 7(a) shows a vertical cross-section (VCS) of the difference in air temperature difference Aerosol-Control (i.e. Aer-Con) at 78ºE. Due to the presence of aerosols the temperature near the surface is decreased by -0.35^{0}C from the surface up till about 800 hPa from 6ºN to the foothills of Himalaya (i.e. 30ºN), while temperatures are increases slightly (about 0.05ºC) in the upper atmosphere. This is consistent with the Ramanthan et al, (2007), where microwave satellite data was used to show that the free troposphere temperature over the India increased as compared to the surface temperature from 1979 to 2003. That study showed that the free troposphere warmed compared to the surface by about 0.5ºC.

The main reason for this is black carbon aerosols in the atmosphere. The BC aerosols have two effects: First, they absorb direct solar radiation, which would otherwise reach to the earth surface. In this case, the air above the surface would be warmed but the surface would be cooled, which would affect the stability of the atmosphere and suppress convective clouds and precipitation. Second, by absorbing solar radiation it reduces the amount of solar radiation reflected to the space. The Figure 7(b) shows the difference (Aer-Con) in temperature at 1000 hPa. Both land and ocean areas have a negative temperature difference here due to the presence of aerosols. However, at 700 hPa, as shown in Figure 7(c), the temperature difference have negative values in some parts of the land area (i.e. North of Pakistan and India), by up to -0.06ºC in a latitudinal band (i.e. 24ºN to 35ºN) but have positive values by up to 0.06ºC over the ocean, i.e. in the Bay of Bengal, south of India and over the Arabian Sea in the (5ºN to 19ºN) latitudinal band. Similarly, at 500 hPa in Figure 7(d), the temperature difference is positive over the ocean in the latitudinal band (i.e. 5ºN to 15ºN) and is negative by up to -0.05ºC in the latitudinal band (i.e. 24ºN to 38ºN).

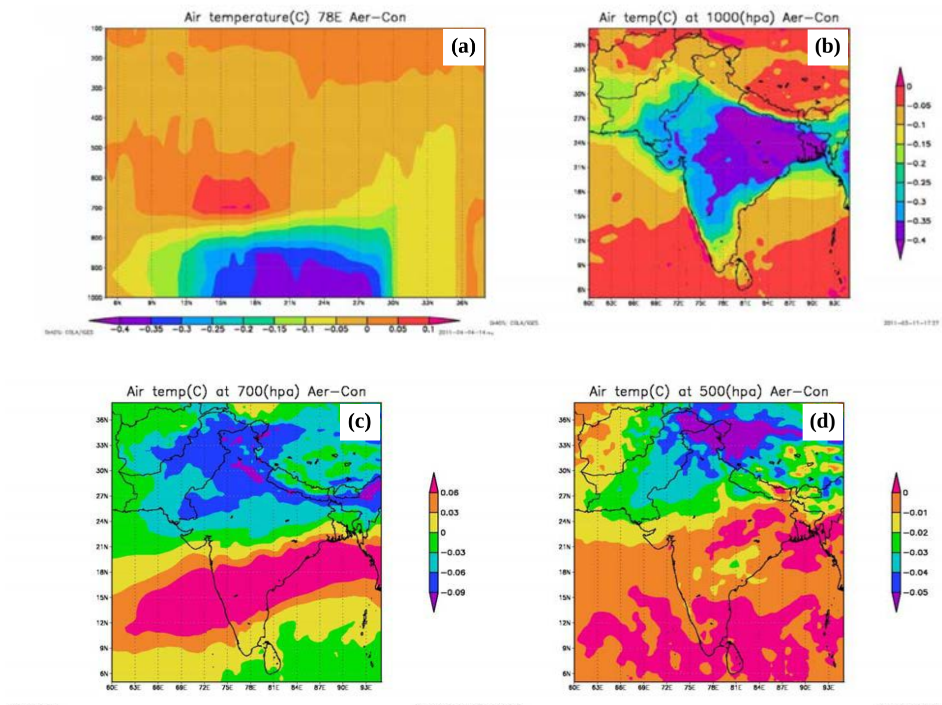
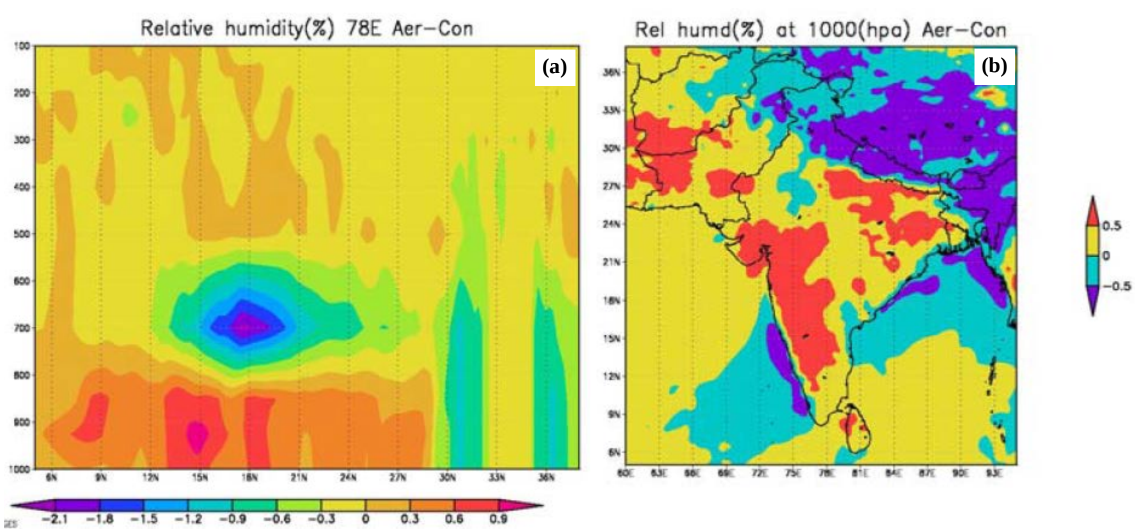


Figure 7: The winter mean (1990-1998) DJF temperature ºC **(a)** vertical cross-section at the 78 ºLongitude **(b)** at 1000 hPa **(c)** at 700 hPa **(d)** at 500 hPa.

The temperature difference thus has dipole structure in the free troposphere; it can be seen at 700 hPa and 500 hPa, which may be due to the transport of aerosols from land towards the ocean.



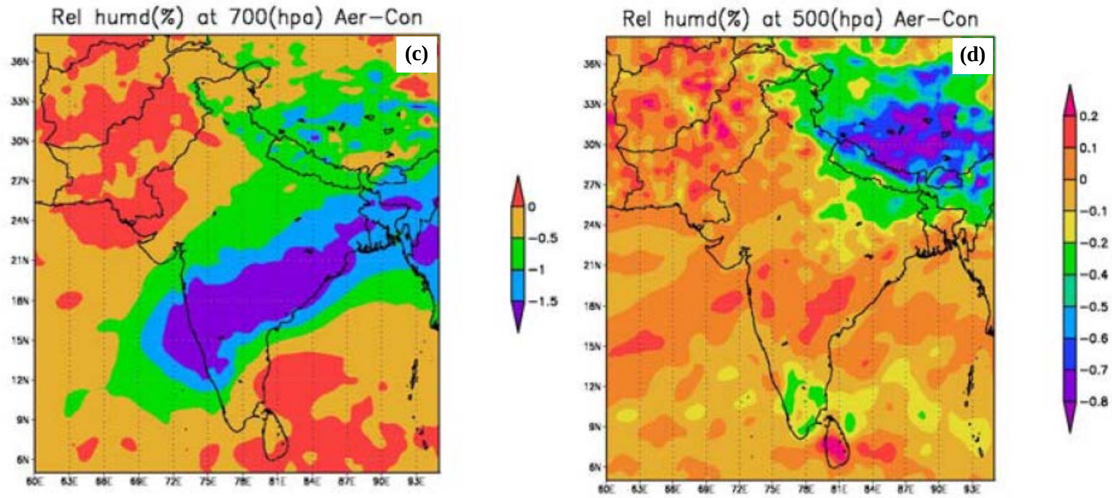


Figure 8: The mean winter (Aer-Con) DJF relative humidity (%) from 1990 to 1998 **(a)** vertical cross-section at 78º Longitude **(b)** at 1000 hPa **(c)** at 700 hPa and **(d)** at 500 hPa.

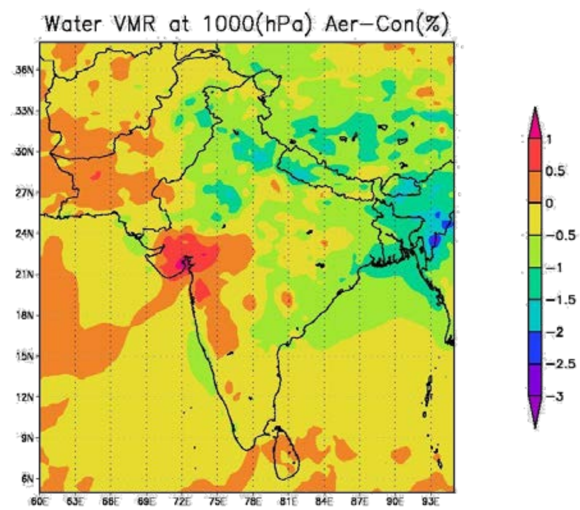


Figure 9: The mean Aer-Con DJF water vapour mixing ratio at 1000 hPa from 1990 to 1998.

The cross-sectional of the difference in relative humidity at 78ºE in Figure 8(a) is consistent with the Figure 7(a). In Figure 8(a), the relative humidity (RH) is increasing by about 0.6 % from the surface up till 800 hPa. This increase in RH is found from 5ºN to the foothills of the Himalaya (i.e. 29ºN) while it decreases by about -1.5 % in the upper atmosphere. However, by comparing Figure 8(b) with the Figure (9), the relative humidity at 1000 hPa is positive over the southwest and northeast of India, south of Afghanistan and southwest of Pakistan and is negative over the Arabian Sea, Bay of Bengal, and north of Pakistan whereas the water vapour mixing ratio difference, as shown in Figure 9. It has mostly negative values over the land except southwest of India. This shows that at 1000 hPa the difference in the moisture content (i.e. water vapour mixing ratio) is constant while RH changed (i.e. positive) as the temperature difference is negative.

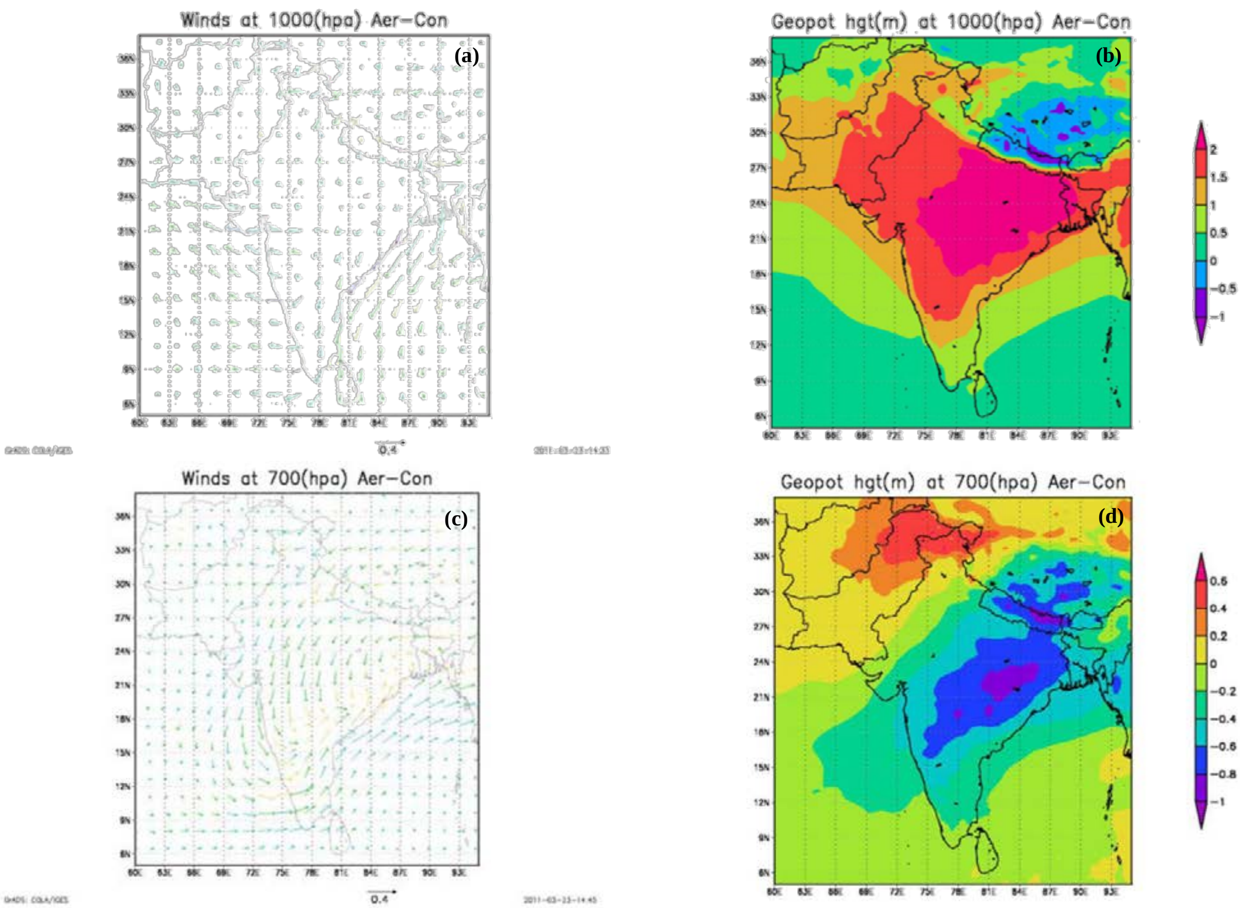
Figure 8(c) shows that the relative humidity at 700 hPa decreases by up to -1.5 % over the land and ocean (i.e. Bay of Bengal, south India and Arabian Sea) but is positive over the Indian Sea, south and west of Pakistan, east and south of Afghanistan, which is consistent with Figure 7(c). In Figure 8(d) at 500 hPa, the difference in relative humidity has positive values over the most part of Afghanistan and

Pakistan; however, it has some negative values over the Indian Ocean. The relative humidity has dipole structure at 700 hPa and 500 hPa and is consistent with the Figures 7(c) and 7(d).

Changes in the Large Scale Circulation

The difference in the air temperature, and consequently geopotential heights (GPH) through hydrostatic adjustment, implies change in the wind field. Figure 10(b) shows that a high pressure anomaly develops near the surface over whole Indian region and also over the east part of Pakistan. This pressure anomaly strengthens the northerly winds as shown in Figure 10(a), blowing from the Bay of Bengal and responsible for the winter monsoon over the Sri Lanka. The winds anomaly in Figure 10(a), due to aerosols, shows that wind is low over land and more perturbed over the ocean. Comparing, the difference in GPH in Figure 10(b) and to the Figure 7(b), the temperature at 1000 hPa in northeast India decreased about 0.4ºC while the difference in GPH increased in the same area about 2m. Figure 10(d) shows that at 700 hPa a low geopotential anomaly develops over the northeast India and a high geopotential anomaly develops over the north parts of Pakistan. The low geopotential anomaly strengthens the cyclonic circulation over the India, as shown in Figure 10(c), which weakens the northerly winds. In Figure 7(d) the difference in temperature increased about 0.06ºC due to the presence of aerosols in the 8ºN to 20ºN latitude band. However, in Figure 10(d) the difference in GPH decreased about 0.6 m in the 15ºN to 27ºN latitudinal band.

At 500 hPa, Figure 10(d) shows the shift of the low geopotential towards northern India and Nepal (i.e. over the Tibetan plateau) and the high geopotential shifts towards the ocean. This contributes to shift the cyclonic circulation anomaly of the winds over to the northern India. However, at 500 hPa the difference in temperature shown in Figures 7(d) is positive in the 60N to 150N and negative in the 26ºN to 38ºN latitude bands. Similarly, the difference in GPH in Figure 10(f) is positive in the 6ºN to 15ºN and is negative in 21ºN to 33ºN latitude bands.



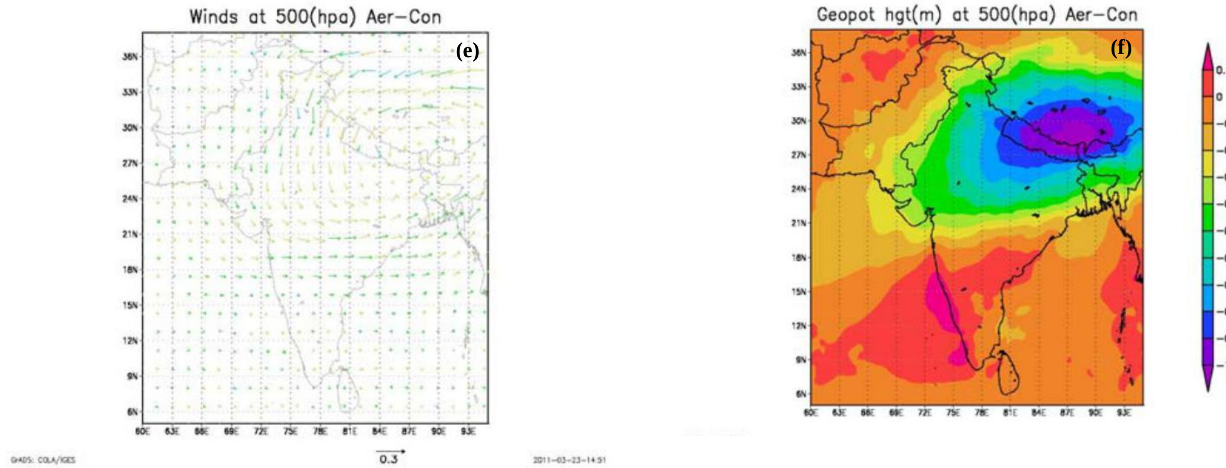


Figure 10: The mean Aer-Con DJF winds (m/s) and geopotential heights GPH (m) from 1990 to 1998 **(a)** Winds at 1000 hPa **(b)** GPH at 1000 hPa **(c)** Winds at 700 hPa **(d)** GPH at 700 hPa **(e)** Winds at 500 hPa **(f)** GPH at 500 hPa.

Conclusion and Summary

In this study the coupled RegCM4-aerosol regional model was used to study the climatic effects of the anthropogenic aerosols (i.e. SO_2, BC and OC) over the SA. The model performed quite well for the estimation of the winter season aerosol's impact. The main aim of the study is to evaluate the different scenario of the SA climate. The model reproduced realistic results for the temperature, precipitation and mean flow by comparing it with the CRU observation and NCEP data respectively. The model has the cold biased and it slightly over-estimated the precipitation over the SA.

By comparing the Control run and CRU observation, the temperature difference (i.e. Control and CRU) is approximately up to 1ºC in the latitudinal bands (23ºN to 36ºN) and difference in precipitation is up to 1 or 2 ($mm.d^{-1}$). However, for the mean flow the CRU observation did not include the wind speed; we then evaluate the RegCM4 against the NCEP reanalysis data. This showed the good consistency between these two datasets, at 1000 hPa the northerly winds are blowing from the Bay of Bengal (i.e. ocean) towards the Sri Lanka but on 700 hPa the westerly winds are blowing over the land in (18ºN-36ºN) latitudinal band.

The SO_2 emission inventory has been taken by the EDGAR while the BC and OC is taken by the Cooke et al, (2002) at 1.0×1.0 degree resolution. The emission of SO_2 is up to 6.5×10^{-10} ($mg^{-1}.m^{-2}.day$) whereas the emission of BC and OC is up to 1×10^{-12} ($mg^{-1}.m^{-2}.day$) from the mega cities of India. This anthropogenic emission directly effect: the SRF, temperature and relative humidity.

In the presence of aerosols, the SRF is negative in whole SA and is up to $-8W.m^{-2}$ due to the masking effect of aerosols, in the eastern and southern Indian coasts, while the difference in temperature is up to -0.4ºC near the surface in the eastern side of India. However, the difference in relative humidity is up to 0.5 % near the surface in the east and southwest part of India. This means that, in the presence of aerosols the SRF decreases, temperature decreases but the RH increases near the surface. Similarly, the vertical cross section (VCS) of temperature is showed that it decreased up to -0.35ºC in the lower atmosphere (i.e. 1000 hPa to 800 hPa) and is increased up to 0.05ºC in the upper atmosphere (700 hPa to 600 hPa). This increase of temperature in the upper atmosphere might be due to the absorption of the solar radiation by the BC. Whereas, the VCS of relative humidity showed that it is increased up to 0.6 % in the lower atmosphere (i.e. 1000 hPa to 800 hPa) and is decreased in between –1.5 to -0.3 in the upper atmosphere (i.e. 700 hPa to 600 hPa).

The aerosols indirectly effected the large scale circulation which strengthen the northerly winds due to the high geopotential near the surface, whereas; on high altitudes the presence of aerosols weakens the geopotential and also northerly winds which then shifts the circulation towards the northeast India.

It should be noted that the simulation of aerosols and their climatic effects are complicated and uncertainties in results can slightly be larger. However, this study is limited in many ways; First, the biogenic particles, dust particles and sea salt aerosols are not included in this study which is proved to be important over the South Asia (Lau et al, 2006). Second, the emission inventory does not include the inter-annual variability of aerosols but it includes the seasonal transformation of aerosols for year 2000. Third, the indirect aerosol effects of cloud properties are not taken into account in this study. Finally, the lateral boundaries are restricted in advection from outside. The Sud et al, (2009) showed that limited area models are unable to provide the reasonable assessment on the large scale aerosol anomalies of the climate without a two way feedback interaction at the boundaries of the regional domain.

The RegCM4 and the aerosol model improved as compared to the previous version of RegCM but it needs to be improved more in near future.

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References

Adam, J. C., D. P. Lettenmaier, 2003: Adjustment of global gridded precipitation for systematic bias. Journal of Geophysical Research, Vol. 108, 4257, Pp. 1-9.

Araújo, K., D. Mahajan, R. Kerr, M. da Silva, 2017: Global Biofuels at the Crossroads: An Overview of Technical, Policy, and Investment Complexities in the Sustainability of Biofuel Development, Agriculture, Vol. 7, Pp. 1-32.

Cooke, W. F., V. Ramaswamy, P. Kasibhatla, 2002: A general circulation model study of the global carbonaceous aerosol distribution, Journal of Geophysical Research, Vol. 107, 4279, Pp. 2-32.

Daniel, A. L., X. T. Xuexi, N. D. Bofinger, A. N. Wiegand, S. Madronich, 2004: Seasonal variability of secondary organic aerosol: A global modeling study, Journal of Geophysical Research, Vol. 109, D03203, Pp. 1-13.

Dickinson, R. E., R.M. Errico, F. Giorgi, G. T. Bates, 1989: A regional climate model for the western United States, Climate Change, Vol. 15, 3, Pp. 383-422

Dickinson, R. E., P. J. Kennedy, A. Henderson-Sellers, M. F. Wilson, 1986: Biosphere-atmosphere transfer scheme (BATS) for the ncar community climate model, NCAR Technical Note, doi:10.5065/D6668B58.

Giorgi, F., X. Bi, 2005: Regional changes in surface climate interannual variability for the 21st century from ensembles of global model simulations, Geophysical Research Letters, Vol. 32, L13701, Pp. 1-5.

Giorgi, F., P. H. Whetton, R. G. Jones, J. H. Christensen, L. O. Mearns, B. Hewitson, H. vonStorch, R. Francisco, C. Jack, 2001: Emerging patterns of simulated regional climatic changes for the 21st century due to anthropogenic forcings, Geophysical Research Letters, Vol. 28, 17, Pp. 3317– 3320.

Grell, G. A., J. Dudhia, D. R. Stauffer, 1994: A description of the fifthgeneration penn state/ncar mesoscale model (mm5), NCAR Technical Note, doi:10.5065/D60Z716B.

Grell, G. A, 1993: Prognostic evaluation of assumptions used by cumulus parameterizations, Monthly Weather Review, Vol. 121, Pp. 764–787.

IPCC, 2007: The Physical Science Basis, Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change, Cambridge University Press, Pp. 996.

Kaleem M., 2016: A sensitivity study of decadal climate prediction to aerosol variability using ECHAM6-HAM (GCM), Bonner Meteorologisches Abhandlungen, thesis no. 74, Pp. 98, http://hss.ulb.uni-bonn.de/2016/4539/4539.pdf.

Kalnay, E., M. Kanamitsu, R. Kistler, W. Collins, D. Deaven, L. Gandin, M. Iredell, S. Saha, G. White, J. Woollen, Y.Zhu, A. Leetmaa, B. Reynolds, M. Chelliah, W. Ebisuzaki, W. Higgins, J. Janowiak, K. Mo, C. Ropelewski, J. Wang, R. Jenne,. D. Joseph, 1996: The NCEP/NCAR 40-Year Reanalysis Project, Bulletin of the American Meteorological Society, Vol. **77**, Pp. 437–471.

Kieu, T. X., L. Duc, H. T. Minh Ha, 2004: Vietnam National University of Hanoi Simulation of Southeast Asia Rainfall using RegCM3 and Problems, Vietnam National University of Hanoi, Pp. 1-4.

Kiehl, J. T., J. J. Hack, G. B. Bonan, B. A. Boville, B. P. Briegleb, D. L. Williamson, P. J. Rasch, 1996: Description of the NCAR Community Climate Model (CCM3), NCAR Technical Note, Pp. 1-152.

Kirkevåg, A., T. Iverson, 2002: Global direct radiative forcing by process-parameterized aerosol optical properties, Journal of Geophysical Research, Vol. 107, 4433, Pp. 1-16.

Lau K. M., M. K. Kim, K. M. Kim, 2006: Asian summer monsoon anomalies induced by aerosol direct forcing: the role of the Tibetan Plateau, Climate Dynamics, Vol. 26, Pp. 855-864.

Leon, J. F., P. Chazette, F. Dulac, J. Pelon, C. Flamant, M. Bonazzola, G. Foret, S. C. Alfaro, H. Cachier, S. Cautenet, E. Hamounou, A. Gaudichet, L. Gomes, J. L. Rajot, F. Lavenu, S. R. Inamdar, P. R. Sarode, J. S. Kadadevarmath, 2001: Large-scale advection of continental aerosols during INDOEX, Journal of Geophysical Research, Vol. 106, 28427-28439.

Lelieveld, J., P. J. Crutzen, V. Ramanathan, M. O. Andreae, C. M. Brenninkmeijer, T. Campos, G. R. Cass, R. R. Dickerson, H. Fischer, J. A. de Gouw, A. Hansel, A. Jefferson, D. Kley,A. T. de Laat, S. Lal, M. G. Lawrence, J. M. Lobert, O. L. Mayol-Bracero, A. P. Mitra, T. Novakov, S. J. Oltmans, K. A. Prather, T. Reiner, H. Rodhe, H. A. Scheeren, D. Sikka, J. Williams, 2001: The Indian Ocean Experiment: Widespread air pollution from South and Southeast Asia, Science, Vol. 5506, Pp. 1031-1036.

Loveland, T. R., B. C. Reed, J. F. Brown, D. O. Ohlen, Z. Zhu, L. Yang, J. W. Merchant, 2000: Development of a global land cover characteristics database and IGBP DISCover from 1-km AVHRR data, International Journal of Remote Sensing, Vol. 21, Pp. 1303–1330.

Lohmann, U., J. Feichter, 2004: Global indirect aerosol effects: a review, Atmospheric Chemistry and Physics, Vol. 4, Pp. 7561-7614.

Mohanty, U. C., D. S. Niyogi, S. Raman, A. Sarkar, 2001: Numerical study of the role of land-air sea interactions for the northeasterly monsoon circulations over Indian Ocean during INDOEX, Current Science (Supplement), Vol. 80, Pp. 60-68.

New, M., D. Lister, M. Hulme, I. Makin, 2002: A high-resolution data set of surface climate over global land areas, Climate Research, Vol. 21, Pp. 1-25.

Pal, J. S., E. E. Small, E. A. B. Eltahir, 2000: Simulation of regional scale water and energy budgets:Representation of subgrid cloud and precipitation processes within RegCM, Journal of Geophysical Research, Vol. 105, D24, Pp. 29579–29594.

Pant, G. B., K. Rupa Kumar, 1997: Climates of South Asia, International Journal of Climatology, Vol. 18, 5, Pp. 581-581.

Qian, Y., F. Giorgi, Y. Huang, 2001: Regional simulation of anthropogenic sulfur over east asia and its sensitivity to model parameters, Tellus B, Vol. 53, Pp. 171–191.

Raman, S., D. S. Niyogi, M. Simpson, J. Pelon, 2002: Dynamics of the elevated land plume over the Arabian Sea and the northern Indian Ocean during northeasterly monsoons and during the Indian Ocean experiment (INDOEX), Geophysical Research Letters, Vol. 29, 16, Pp. 1-4, doi:10.1029/2001GL014193.

Ramanathan, V., P. J. Crutzen, J. Lelieveld, A. P. Mitra, D. Althausen, J. Anderson, M. O. Andreae, W. Cantrell, G. R. Cass, C. E. Chung, A. D. Clarke, J. A. Coakley, W. D. Collins, W. C. Conant, F. Dulac, J. Heintzenberg, A. J. Heymsfield, B. Holben, S. Howell, J. Hudson, A. Jayaraman, J. T. Kiehl, T. N. Krishnamurti, D. Lubin, G. Mcfarquhar, T. Novakov, J. A. Orgen, I. A. Podgorny, K. Prather, K. Preistley, J. M. Prospero, P. K. Quinn, K. Rajeev, P. Rasch, S. Rupert, R. Sadourny, S. K. Satheesh, G. E. Shaw, P. Sheridan, F. P. J. Valero, 2001: Indian Ocean Expermient: An integrated analysis of the climate forcing and effects of the great Indo-Asian haze, Journal of Geophysical Research, Vol. 106, D22, Pp. 28371-28398.

Ramanathan, V., M. V. Ramana, 2005: Persistent, widespread, and strongly absorbing haze over the Himalayan foothills and the Indo-Gangetic Plains, Pure and Applied Geophysics, Vol. 162, Pp. 1609-1626.

Ramanathan, V., F. Li, M. V. Ramana, P. S. Praveen, D. Kim, C. E. Corrigan, H. Nguyen, E. A. Stone, J. J. Schauer, G. R. Carmichael, B. Adhikary, S. C. Yoon, 2007: Atmospheric brown clouds: Hemispherical and regional variations in long-range transport, absorption, and radiative forcing, Journal of Geophysical Research, Vol. 112, D22S21, Pp. 1-26, doi:10.1029/2006JD008124.

Reynolds, R. W., T. M. Smith, 1994: Improved global sea surface temperature analyses using optimum interpolation, Journal of Climate, Vol. 7, Pp. 929–948.

Solmon, F., F. Giorgi, C. Liousse, 2006: Aerosol modelling for regional climate studies: application to anthropogenic particles and evaluation over a European/African domain, Tellus B, Vol. 58, Pp. 51–72.

Slingo, A., 1989: A GCM parameterization for the shortwave radiative properties of water clouds, Journal of the Atmospheric Sciences, Vol. 46, 10, Pp. 1419-1427.

Sud, Y. C., W. K. M. Lau, E. Wilcox, G. K. Walker, X. -H.Liu, A. Nenes, D. Lee, K. –M. Kim, Y. Zhou, P. S. Bhattacharjee, 2009: Sensitivity of Boreal-Summer Circulation and Precipitation to Atmospheric Aerosols in Selected Regions – Part 1: Africa and India,Annales Geophysicae, Vol. 27, Pp. 3989-4007.

Syed, F. S., F. Giorgi, J. S. Pal, K. Keay, 2009: Regional climate model simulation of winter climate over Central-South Asia, with emphasis on NAO and ENSO effects, International Journal of Climatology, Vol. 30, Pp. 220-235, doi: 10.1002/joc.1887.

Wang, Y., O. L. SEN, B. Wang, 2002: A Highly Resolved Regional Climate Model (IPRC-RegCM) and its Simulation of the 1998 Severe Precipitation Event over China. Part I: Model Description and Verification of Simulation, Journal of climate, Vol. 16, Pp. 1721-1738.

Wu, J., C. Fu, Y. Xu, J. P. Tang, W. Wang, Z. Wang, 2008: Simulation of direct effects of black carbon aerosol on temperature and hydrological cycle in Asia by a regional climate model. Meteorology and Atmospheric Physics, Vol. 100, 1-4, Pp. 179-193.

Zanis, P., 2009: A study on the direct effect of anthropogenic aerosols on near surface air temperature over Southeastern Europe during summer 2000 based on regional climate modeling, Annales Geophysicae, Vol. 27, Pp. 3977-3988.

Zeng, X., M. Zhao, R. E. Dickinson, 1998: Intercomparison of bulk aerodynamic algorithms for the computation of sea surface fluxes using toga coare and tao data, Journal of Climate, Vol. 11, Pp. 2628–2644.