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The effect of aerosol on surface cloud radiative forcing in the Arctic

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Abstract

Cloud radiative forcing is a very important concept to understand what kind of role the clouds play in climate change with thermal effect or albedo effect. In spite of that much progress has been achieved, the clouds are still poorly described in the climate models. Due to the complex aerosol-cloud-radiation interactions, high surface albedo of snow and ice cover, and without solar radiation in long period of the year, the Arctic strong warming caused by increasing greenhouse gases (as most GCMs suggested) has not been verified by the observations. In this study, we were dedicated to quantify the aerosol effect on the Arctic cloud radiative forcing by Northern Aerosol Regional Climate Model (NARCM). Major aerosol species such as Arctic haze sulphate, black carbon, sea salt, organics and dust have been included during our simulations. By inter-comparisons with the Atmospheric Radiation Measurement (ARM) data, we find surface cloud radiative forcing (SCRF) is $-22\,W/m^2$ for shortwave and $36\,W/m^2$ for longwave. Total cloud forcing is $14\,W/m^2$ with minimum of $-35\,W/m^2$ in early July. If aerosols are taken into account, the SCRF has been increased during winter while negative SCRF has been enhanced during summer. Our estimate of aerosol forcing is about $-6\,W/m^2$ in the Arctic.

1. Introduction

Clouds have significant impacts on energy budget and water cycle in the earth-atmosphere system. They can produce surface cooling effect by reflecting the sunlight and surface warming effect by trapping the longwave radiation lost to space. Low clouds such as stratocumulus clouds exert strong influence on the surface solar radiation budget (SSRB) through their large extent, longer existence and high reflectivity of solar radiation. On the other hand, cirrus clouds exert different influence on the earth's climate through their effect on outgoing longwave radiation. As both types of these clouds have complicated inhomogeneity in microphysics and morphology, it is extremely difficult for remote sensing and modelling of their optical and microphysical properties. With recognition of its significance, numerous ground-based and space-borne programs are designed to monitor the surface radiation budget and determine the cloud properties. There are ground-based observation networks such as world Baseline Radiation Network (Ohmura et al., 1998), the Atmospheric Radiation Measurements (ARM) program (Stokes and Schwartz, 1994). Several satellite missions conducted intensive field observation such as the Moderate Resolution Imaging Spectroradiometer (MODIS; King et al., 1992), the International Satellite Cloud Climatology Project (ISCCP, Rossow and Schiffer, 1999), and the Earth Radiation Budget Experiment (ERBE, Barkstrom and Smith, 1986; Harrison et al., 1990; Li and Leighton, 1993).

From the important product of ERBE experiment, the concept of Cloud Radiative Forcing (CRF) was proposed to understand what kind of role the clouds play in climate change with thermal effect or albedo effect (Ramanathan et al., 1989). This concept can be directly applied to examine the effectiveness of cloud and radiation schemes in the climate models. Unfortunately, a series of the reported discrepancy between measurements and simulations have been puzzled by atmospheric radiation community for many years (Stephens et al., 1978; Herman and Curry, 1984; Ramanathan et al., 1995; Chou et al., 1995; Li, 1998). The basic theory of atmospheric radiation had even been doubted. Many hypotheses such as absorbing aerosol effect, side effect of clouds, size distribution effect of droplets or unobserved water vapor absorption have been presented to explain the 'anomalous absorption paradox'. In spite of much progress achieved in past years, the stunning clouds or those involving with clouds in current numerical models are still poorly described (Cess et al., 1989; Li et al., 1997; Li and Fu, 2000; Barker et al., 2003; Wang et al., 2004). In the Arctic, clouds can influence the surface radiation budget more strongly, reducing wintertime cooling of surface by 40–50 W/m^2 and summertime heating of the surface by 20–30 W/m^2 (Curry et al., 1996; Briegleb et al., 1998). The sensitivity of CRF to uncertainties in special surface albedo, atmospheric composition and cloud properties has caused a large variability of current estimates for polar CRF by measurements (Curry and Ebert, 1992; Rossow and Zhang, 1995; Doelling et al., 2001).

Most General Circulation Models (GCMs) simulations predict that any warming in global climate will be enhanced at the polar regions (Washington and Meel, 1989; Manabe et al., 1992, 1994). However, measurements have no evidence for greenhouse warming over the Arctic ocean in past decades (Kahl et al., 1994). One of reasons is that Arctic haze aerosols may alter the number concentration and mean diameter of cloud droplets or particles and consequently increase the precipitation. This dehydration effect will lead to a decrease of the downward longwave radiation and cooling at the surface (Blanchet and Girard, 1994). Other indirect effects of aerosols such as Twomay and Albrecht effects (Twomay, 1991; Albrecht, 1989) may strongly affect cloud microphysical and radiative properties. These indirect effects of aerosols can be contributed greatly to the uncertainties of quantifying the Arctic CRF.

In this study, we examine the surface CRF in the Arctic by using Northern Aerosol Regional Climate Model (NARCM) simulations. The model results will be compared with the observational data obtained from the Surface Heat Budget of Arctic Ocean (SHEBA) field experiment relative to Arctic Regional Climate Model Inter-comparison Project (ARCMIP). We begin with the definition of cloud radiative forcing and describe the modelling system and initialization in Sect. 2. In Sects. 3 and 4, we present the results and conclusion respectively.

2. Methodology

2.1. Cloud radiative forcing

To assess the impact of clouds on the radiation budget at the surface, cloud radiative forcing (CRF) is a simple means first introduced by Ramanathan et al. (1989). If CRF is positive, the clouds act to warm the surface(i.e. the thermal effect is dominant). If CRF is negative, then the clouds act to cool the surface(i.e. the albedo effect is dominant). The net radiation at the surface is the difference between absorbed solar radiation and emitted longwave radiation.

$$R = (1 - \alpha_s)F_{sd} - \epsilon\sigma T_s{}^4 + \epsilon F_{ld}, \tag{1}$$

Here R is the net radiation. α_s is surface albedo, F_{sd} is downward solar radiation, Ts is surface temperature, F_{ld} is downward longwave radiation, and the emissivity ϵ. So we define the net CRF as the difference between the average and clear-sky radiative flux:

$$CF = R - R_{clear}, \tag{2}$$

Aerosol particles and clouds can exert a cooling (parasol) effect on climate (Crutzen and Ramanathan, 2003). In order to easily quantify the aerosol effect on the CRF, we use the concept of Aerosol Radiative Forcing (ARF) includes the direct and indirect radiative forcing at the surface:

$$AF = CF_{aerosol} - CF, \tag{3}$$

Here $CF_{aerosol}$ is the CRF with aerosols and CF is the CRF without aerosols.

2.2. Model description

NARCM is built based on the Canadian Regional Climate Model (CRCM) and Canadian Aerosol Module (CAM) for simulating the impact of aerosol on climate in Northern hemisphere. The meteorological module of NARCM (Spacek et al., unpublished manuscript, 1999) is a limited area model, which used the fully elastic, non-hydrostatic Euler equations solved with semi-implicit and semi-lagrangian method (Laprise et al., 1997). The physical parameterization package is imported from the Canadian General Circulation Model (CGCM) (McFarlane at al., 1992).

CAM is a size-segregated multi-component aerosol module including the physical and chemical processes that determine the aerosol composition and size distribution (Gong et al., 2003). Due to the limited area of ARCMIP domain which does not include regions with important aerosol sources, we initialized the model with 5 aerosol species (sulphate, black carbon, dust, sea salt and organics) provided by chemical transport models (CTMs) (Penner et al., 1992; Chin et al., 1996). The vertical distribution of aerosol is fitted from the 9-layer 3-D CTM (Chin et al., 1996) with modifications according to recent aerosol observational network such as AERONET (Holben et al., 1998) and AEROCAN (Bokoye et al., unpublished manuscript, 2003).

The radiation scheme is developed based on the scheme used by ECMWF (Morcrette, 1990). For better treating the aerosol effect, we use four wave bands in shortwave radiation and six wave bands in longwave radiation. The optical properties of aerosols such as the particle scattering or extinction efficiency, single scattering albedo and asymmetry factor are calculated from Mie scattering theory. Refractive indices are based on the data from Ghan et al. (2001). The number concentration and size distribution of cloud droplets and ice particles have been introduced to calculate the radiative flux. For water clouds, the cloud optical depth is related to the liquid water path and effective radius. The asymmetry factor and single scattering albedo are parameterized as the function of optical depth and the effect radius based on the data from Hu and Stamnes (1993). For ice clouds, we have parameterized the optical depth, the asymmetry factor and single scattering albedo as function of effective radius and crystal shape.

For the indirect effect of aerosols, it is very uncertain to quantify this kind of forcing because the nucleation processes of cloud droplets or ice particles are still not fully known. NARCM uses the Lohmann and Roeckner (1996) microphysics scheme. It is a 2-moment scheme with 6 prognostic cloud variables. Aerosols are accounted for in this scheme for water droplet and ice crystal nucleation. As the scheme only considered the sulfate aerosols, we have added black carbon (“soot”), sea salt, and organic aerosol to be the candidates of cloud condensation nuclei (CCN).

Surface albedo has been described in different surface types such as bare soil, snow and surface canopy. In order to consider the spectral influence, the surface albedo in the near-infrared band is assumed to be twice that in the visible band.

3. Results and discussion

We performed two experiments with and without aerosols in the period between September 1997 to September 1998. During this time period, the enhanced measurements such as the Atmospheric Radiation Measurements (ARM) and the surface Heat Budget of Arctic Ocean (SHEBA) data are available. The atmospheric boundary forcing data is supplied by ECMWF data centre. We calculated the surface radiative parameters along the ship track of a year long extensive set of measurements using the Canadian Coast Guard ice breaker Des Groseilliers as a permanent ice station.

3.1. Aerosol radiative forcing

The radiative forcing of aerosols has been quantified in General Circulation Models (Houghton et al., 2001), but the value is still uncertain. One of uncertainties is due to the complicated composition of aerosols. For example, the radiative forcing of soot aerosols behaves completely different way from the sulfate aerosols. Figure 1 presents the global average values of aerosol forcing calculated by single cloudy column model. It is interesting to find that the radiative forcing of soot aerosol is enhanced when clouds exist, but it is diminished for sulfate aerosols. The other uncertainties come from the size distribution and concentration of aerosols. In the Arctic, such kind of measurement is so scarce that we have to use the outputs of CTMs as climatology. Figure 2 presents the optical depth of total five kinds of aerosols. The vertical distribution of aerosol is used by a formula fitted from observation data (Hu et al., 2001).

For pure sulfate aerosols, strong scattering of this kind of aerosol leads to large reduction of downward solar radiative flux at surface and large negative surface radiative forcing (Fig. 3). However, when there are soot aerosols included, they reduce the single scattering albedo of low clouds and the cloud amount, hence increase the downward solar radiative flux at surface and lead to smaller negative surface radiative forcing in comparison with pure sulfate aerosols (Fig. 4).

3.2. Intercomparison with in situ measurements

Figure 5 illustrates the surface albedo used in the simulations and measured by the Atmospheric Surface Flux Group (ASFG) radiometers. The values of model albedo have good agreement with the ASFG radiometer albedos for the fall, winter and spring. During summer, our model values are lower than ASFG albedos. However, our albedos for summer are much nearer to the Cold Regions Research and Engineering Laboratory (CRREL) values. As there were many different ice types including melt ponds and open water in that time, the albedos measured by CRREL are different from ASFG values.

The downward, upward solar radiation (SW) and longwave radiation (LW) Fluxes at the surface are shown in Figs. 6, 7, 8 and 9, respectively. The downwelling and upwelling SW display strong seasonal cycle. The downward SW is underestimated during summer and overestimated during spring as compared to the observed in situ data. The upward SW is consistent well with the observations during spring, but smaller during summer. This is due to the lower values of the surface albedo used in our simulations for summer. The LW radiation plays an important role in the surface energy balance for without SW radiation for a long time period in the Arctic. The upwelling LW radiation values are similar between the simulations and measurements, especially during summer. The downwelling LW values have larger discrepancy between simulations and observations, especially during winter. This is partly due to the clouds which have not been simulated well in our model. The fact is that low clouds are often warmer than the surface because of strong Arctic temperature inversions.

Figure 10 presents the annual cycle of cloud fraction during SHEBA year, averaged over 20 days. We find that the cloudiness is overestimated during winter and underestimated during spring as compared with the measurements. Amazingly, the cloudiness has good agreement between the model and measurements when we carried out the experiment with aerosols.

The SW and LW surface CRFs simulated by NARCM are shown in Fig. 11. Apparently, there is no solar CRF during winter. The longwave CRF is positive with averaged value 25 W/m^2 during winter and spring. As the low clouds are dominated over the Arctic, these clouds inhibit longwave radiation from escaping to space and warm the surface during fall, winter and spring. The solar CRF is negative during summer with average value −22 W/m^2 and the minimum value occurs in late June and early July. Although the strong positive longwave CRF with average value 36 W/m^2 occurs, the solar albedo effect of clouds is dominating during summer.

Figure 12 presents the surface cloud radiative forcing (SCRF) from NARCM simulations and SHEBA observations. The SCRF simulated by NARCM is positive during winter and negative during summer. Our values are lower than SHEBA observations during summer and a little higher during winter. If our results are compared to the satellite results derived from ISCCP (International Satellite Cloud Climatology Project) (Schweiger and Key, 1994), there is a good agreement. One of reasons to cause large discrepancies between modelling and observation during summer is due to different surface albedo used. As we mentioned before, the surface albedo used in our model is lower than the SHEBA observations during summer. For aerosol case, we find the SCRF has been a little increased during winter while negative SCRF has been enhanced during summer. The best estimation of aerosol forcing is −6 W/m^2. All results present a strong spring-summer transition during the melting of snow and sea ice.

The negative SCRF during summer can lead to strong cooling effect at the surface and trigger the sea ice to form earlier in the Arctic ocean due to the feedbacks. On the other hand, the positive SCRF can induce the warming effect at the surface during winter and result in melting of snow and sea ice earlier due to the feedbacks. Thus, the aerosol-cloud forcing will induce warming during the winter season and cooling during the summer season. Moreover, it could trigger the change of salinity and sea ice concentration in the Arctic Ocean, and influence on the natural variability of Arctic Oscillation (AO).

4. Conclusions

It is most favourable to detect the fingerprints of global warming over the Arctic region because the polar climate is so sensitive to anthropogenic trace gases and pollutants. Cloud radiative forcing is a simple and effective means to understand what kind of role the clouds play in Arctic climate change. In this study, we use NARCM to examine the aerosol effect on surface cloud radiative forcing. After comparing with SHEBA measurements, we find the net effect of Arctic clouds is to warm the surface during autumn, winter and spring, and to cool the surface during summer. Aerosols can amplify the above cloud effects and the total aerosol forcing is estimated to be $-6\,W/m^2$ in annual average. For absorbing aerosols such as soot aerosols, the reduction of scattering albedo of low clouds can partly explain the 'anomalous absorption paradox'.

The SCRF is very sensitive to surface albedo, solar zenith angle and microphysical properties of clouds. The inaccurate surface albedo and unknown properties of mixed-phase clouds in our model can be the main source of errors for quantifying the SCRF. Also, the interaction between aerosols and clouds can be very important in the Arctic given the predominance of solid and mixed-phase clouds. Further understanding the microphysical and optical properties of different types of aerosols and clouds over the Arctic needs to be highlighted as the improvements of aerosols, clouds and radiative processes have been tested to be necessary for better simulation of surface temperature (Hu et al., 2003).

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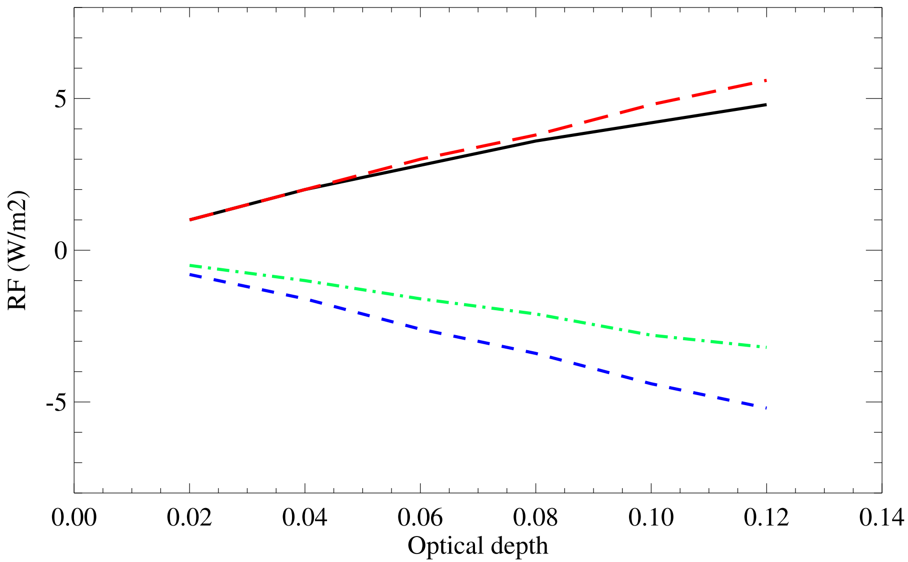

Fig. 1. The radiative forcing of aerosols in clear and cloud sky. Red: soot; green: sulfate; solid and dashed: clear sky; long dashes and dash dot: cloudy sky (Cloudiness=0.48).

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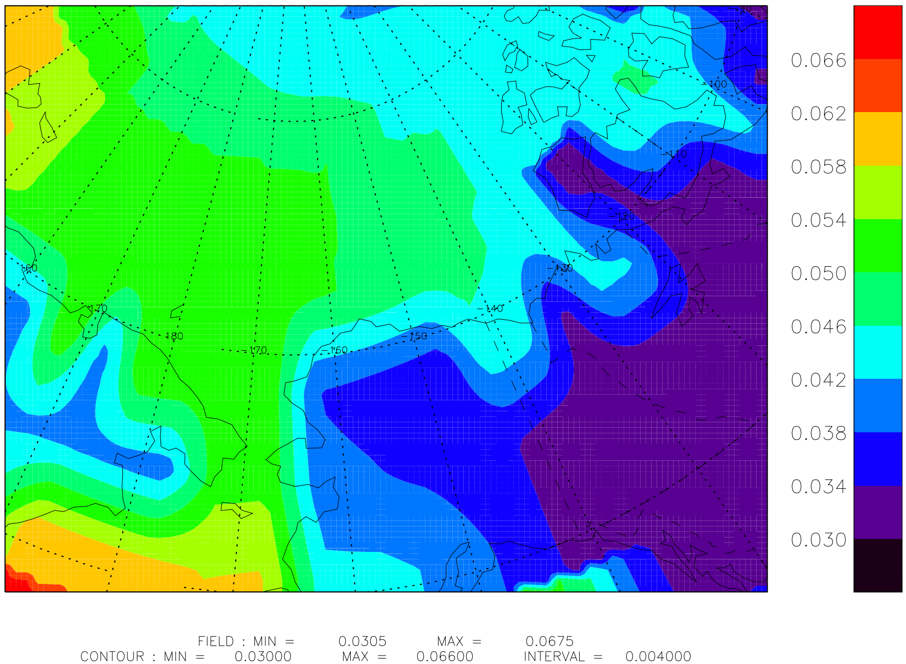

Fig. 2. The optical depth of aerosols in climatology.

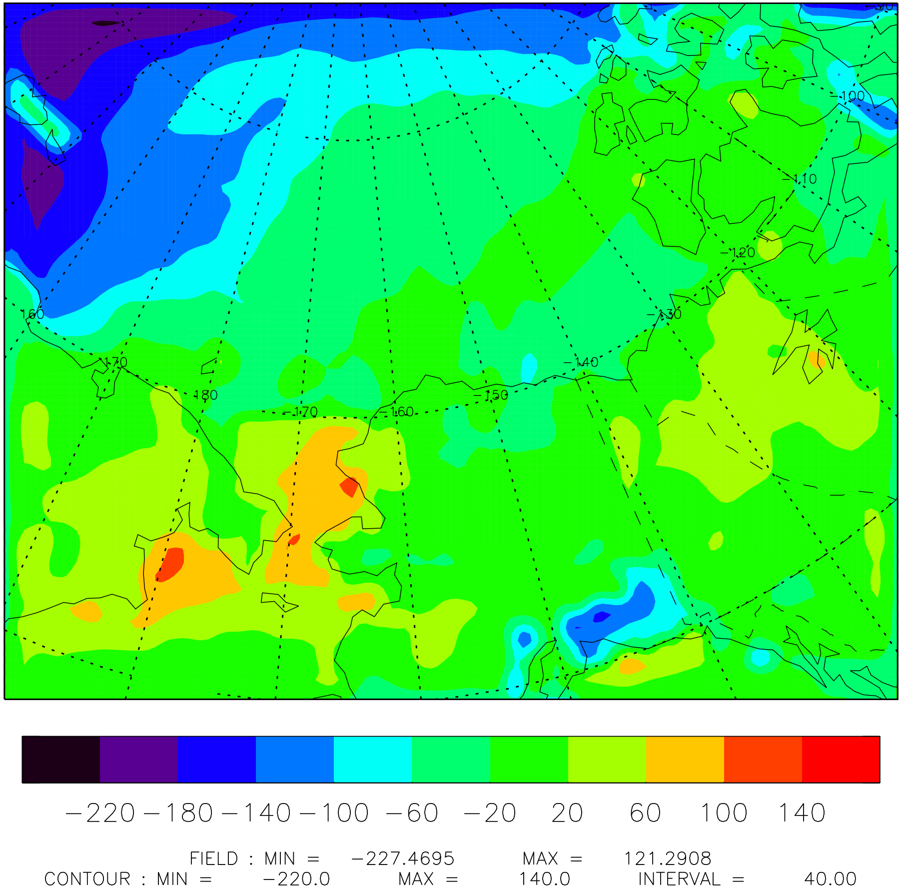

Fig. 3. The surface radiative forcing of sulfate aerosols.

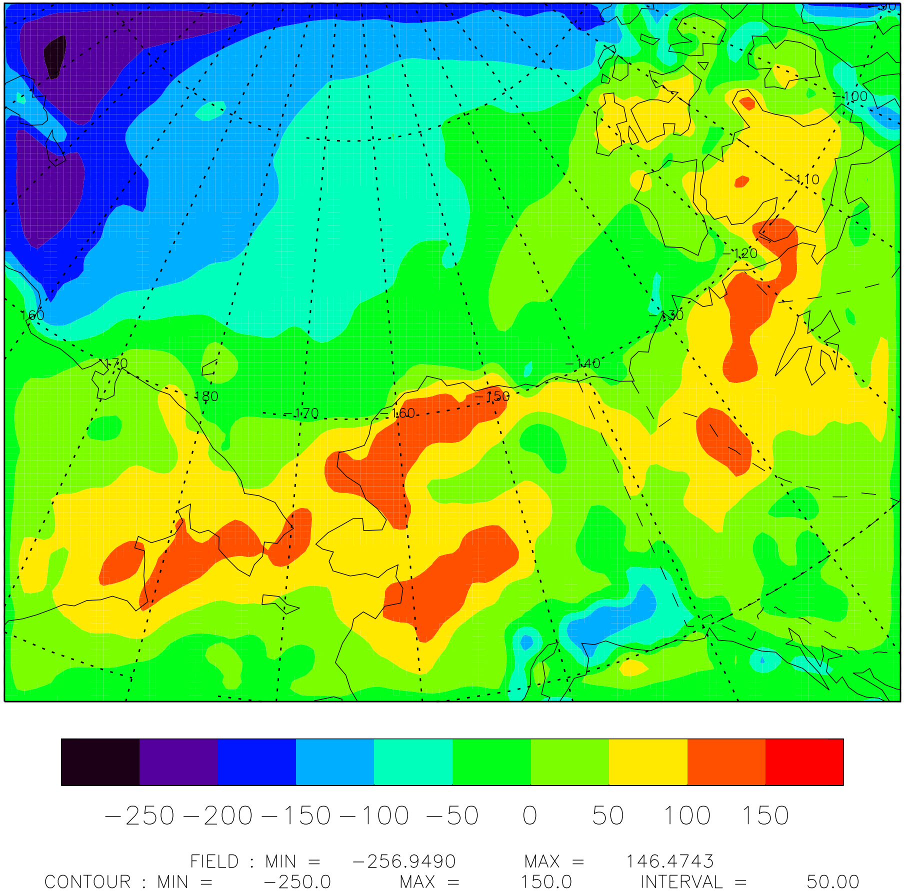

Fig. 4. The surface radiative forcing of 5 kinds of aerosols.

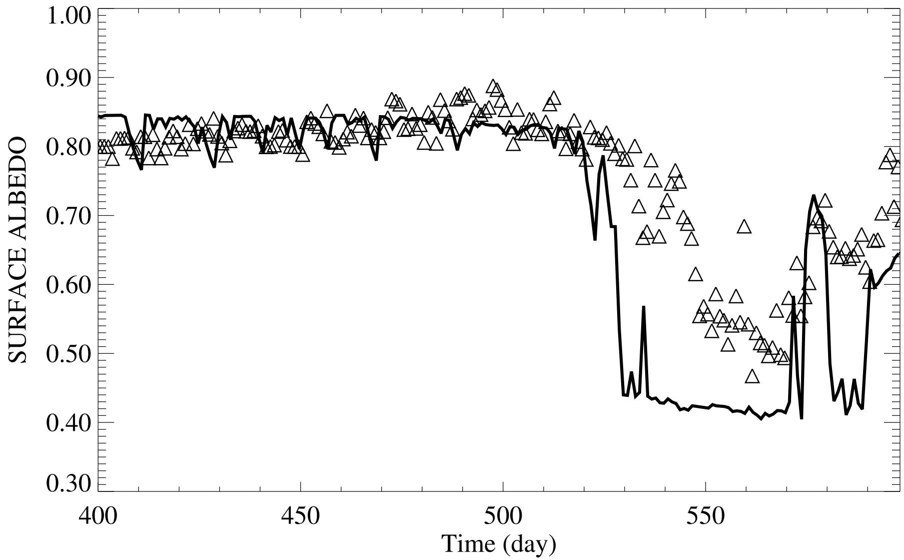

Fig. 5. Annual cycle of surface albedo modeled in NARCM (line) and observed by ASFG (asterisks). Time is in Julian day and plotting begins in early February 1998.

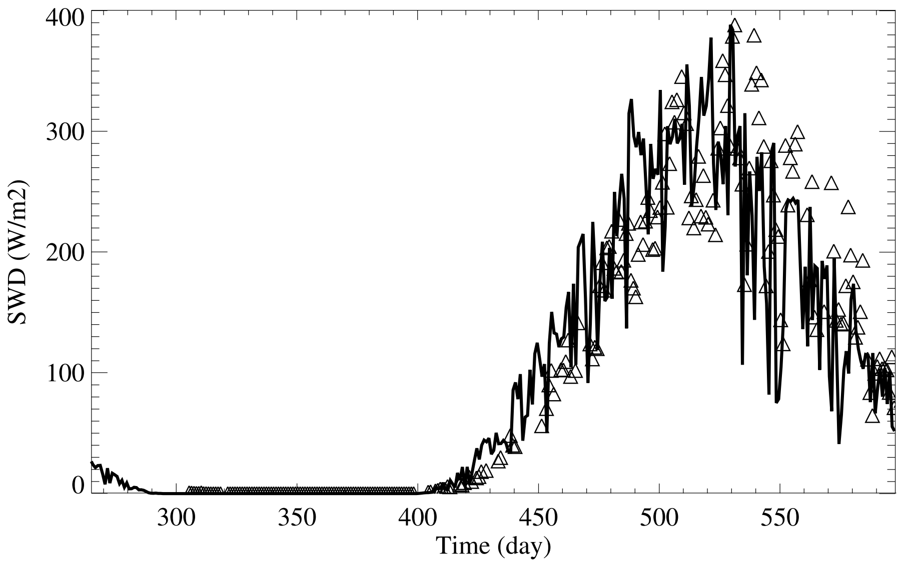

Fig. 6. The downwelling solar radiation at surface simulated by NARCM (line) and measured by ASFG (asterisks).

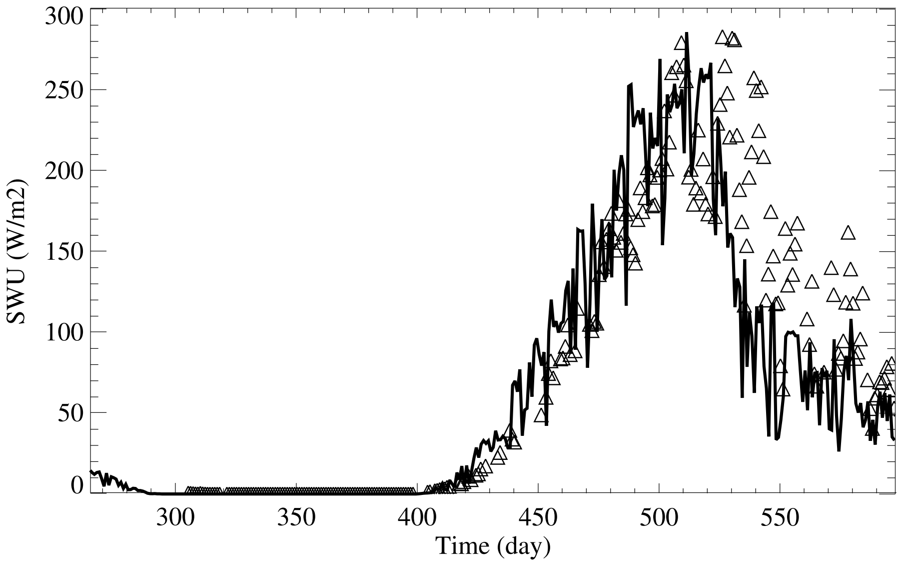

Fig. 7. The upwelling solar radiation at surface simulated by NARCM (line) and measured by ASFG (asterisks).

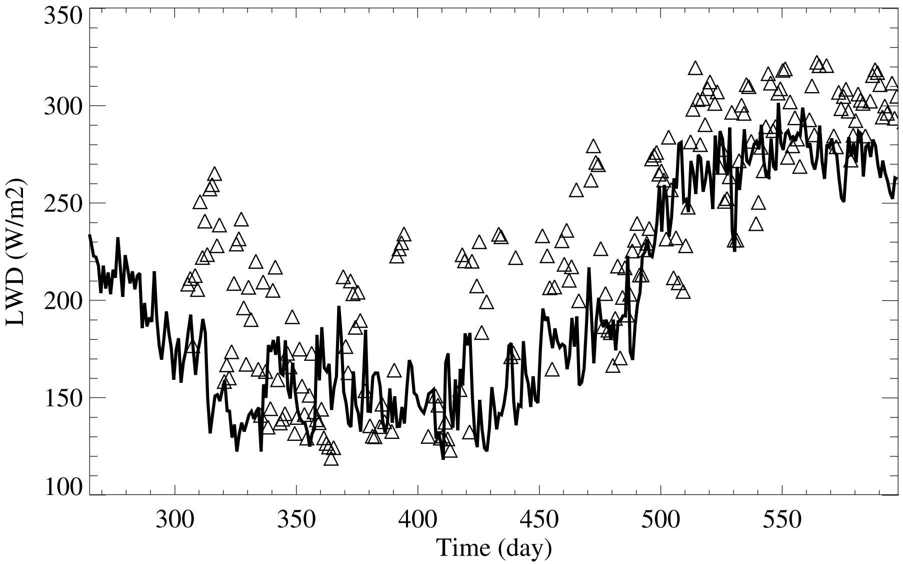

Fig. 8. The downwelling longwave radiation at surface simulated by NARCM (line) and measured by ASFG (asterisks).

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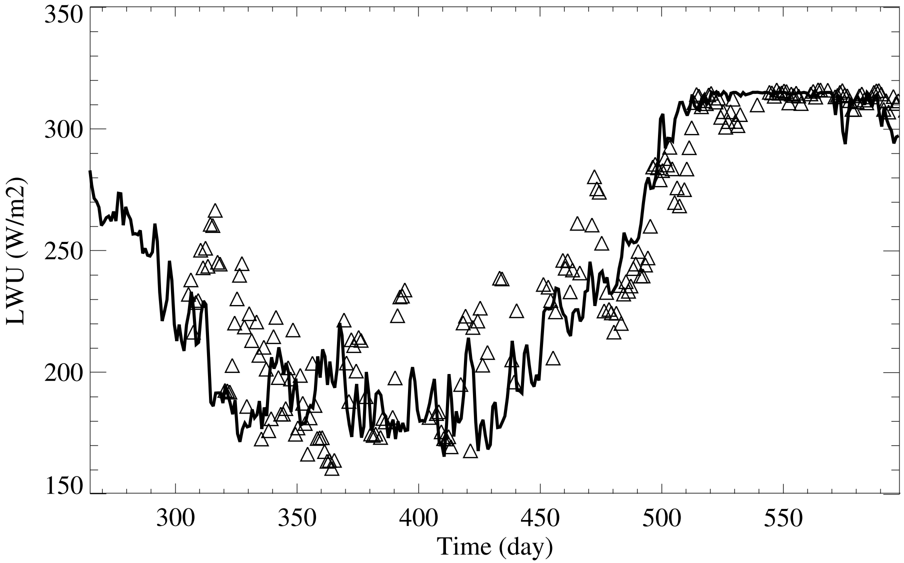

Fig. 9. The upwelling longwave radiation at surface simulated by NARCM (line) and measured by ASFG (asterisks).

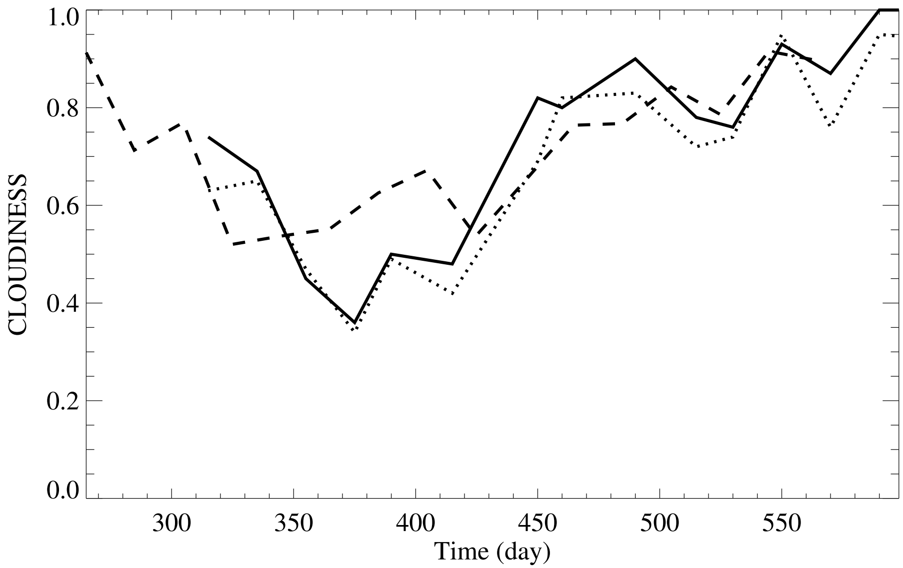

Fig. 10. Annual cycle of cloudiness averaged over 20 day blocks. Solid: ASFG measurements; dashed: model simulation without aerosols; dotted: model simulation with five kinds of aerosols.

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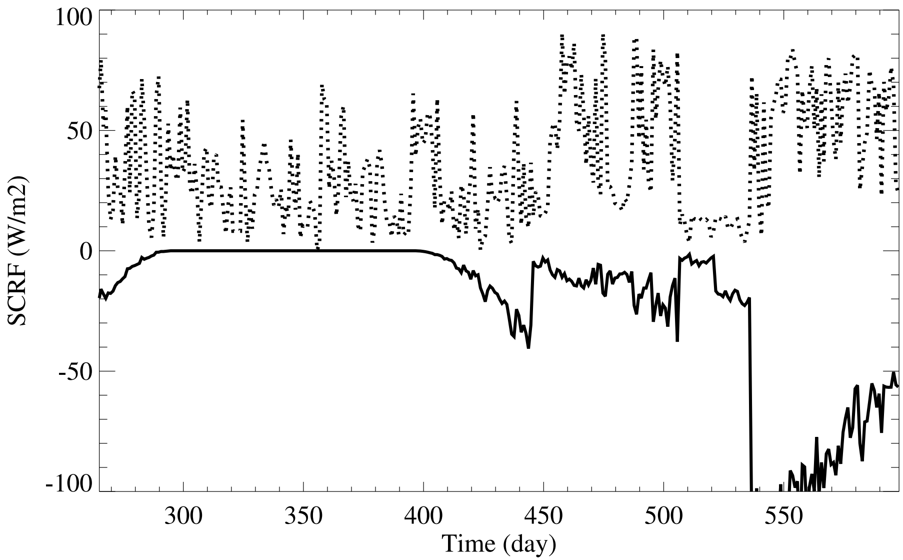

Fig. 11. Annual cycle of solar surface cloud forcing (solid line) and longwave surface cloud forcing (dotted line).

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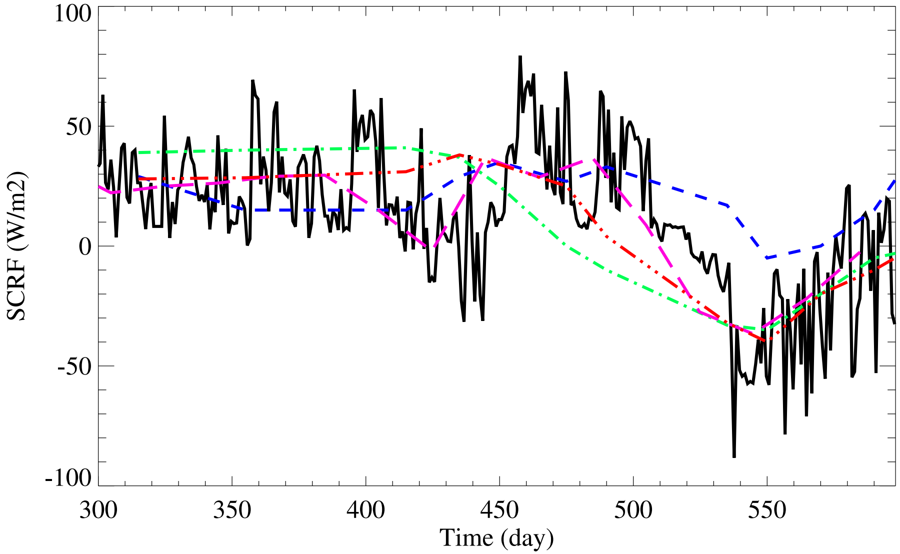

Fig. 12. Comparison of total cloud radiative forcing data. Solid: daily averaged model results; long dashes: 20 days averaged model results; dashed: measurements from ASFG; dash dot: satellite-derived results; dash dot dot: 20 days averaged model simulations with five kinds of aerosols.