Time evolutions of various radiative forcings for the past 150 years estimated by a general circulation model

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[1] Time series of the instantaneous radiative forcings for main anthropogenic and natural forcing agents from the year 1850 to 2000 are evaluated at the Earth's surface as well as at the tropopause with an atmospheric general circulation model. This evaluation corresponds to a simulation of 20th century climate with a synthetic coupled atmosphere-ocean general circulation model. The evaluation indicates that the positive radiative forcing at the tropopause rapidly increases from 1910 to 1950 and after 1970 principally due to long-lived greenhouse gases, while the negative radiative forcing at the surface sharply increases between 1955 and 1965 mainly due to the aerosol direct and indirect effects. This study suggests that a simultaneous analysis of changing rates of the radiative forcing both at the tropopause and surface can explain tendencies of changes in the surface air temperature.
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1. Introduction

[2] There are a lot of natural and anthropogenic climate forcing agents that cause climate change through each specific mechanism. The radiative forcing is one of the popular indices for quantitative impacts of external factors on the climate system, of which we can make comparisons among different agents such as changes in greenhouse gases, ozone (O_3), aerosols, solar radiation, volcanic activities, and surface albedo. In the Intergovernmental Panel on Climate Change (IPCC), it is defined as the change in net irradiance at the tropopause after allowing for stratospheric temperature to readjust to radiative equilibrium with surface and tropospheric temperatures and state held fixed [*Ramaswamy et al.*, 2001], which is coincident with the "adjusted forcing" of *Hansen et al.* [2005a]. Almost studies on the radiative forcing have estimated the "instantaneous forcing" which is defined with the temperature and state held fixed throughout the atmosphere. Fixed surface and atmospheric temperatures and state avoid climate responses and feedbacks with changes in any forcing agents, therefore the radiative forcing is an independent index of the climate sensitivity that is different among global climate models.

[3] Several studies have well simulated a change in the surface air temperature, which is one of the most significant parameters to evaluate the climate change, in the 20th century with coupled atmosphere-ocean general circulation models (AOGCMs) including all of main external natural and anthropogenic climate forcings [e.g., *Stott et al.*, 2000; *Broccoli et al.*, 2003]. Figure 1 shows anomalies of annual global mean surface air temperature from the pre-industrial era by the observations and the AOGCM [*Nozawa et al.*, 2005; *Nagashima et al.*, 2006]. It may be instructive to look at a time evolution of the radiative forcing by external climate forcing agents in the 20th century in order to analyze the temperature trend simulated by the AOGCMs. Several past studies showed the time evolution of the radiative forcing at the tropopause or the top of the atmosphere [e.g., *Myhre et al.*, 2001; *Hansen et al.*, 2002, 2005b]. The radiative forcings at the tropopause may generally explain changes in the overall tropospheric temperatures. On the other hand, *Wild et al.* [2004, 2005] reported that observed long-term variations of solar irradiance at the surface correspond to changes in the surface air temperature. Black carbon aerosol (BC), for example, has positive forcing at the tropopause and consequently increases the atmospheric temperature, while it has negative forcing at the surface and therefore may has an effect on reducing temperature near the surface [*Ramanathan et al.*, 2005].

[4] In this study, changes in the instantaneous radiative forcings from 1850 to the present are demonstrated for long-lived greenhouse gases (LLGHGs), O_3, aerosols, volcanoes, solar variability, and land-use change not only at the tropopause but also at the surface with an atmospheric component of an AOGCM. Time series of the radiative forcings for each climate forcing agent at the surface has not been revealed in past studies. Relationship between changes in the radiative forcings both at the tropopause and surface and a change in the surface air temperature is also discussed.

2. Model Description and Experimental Configurations

[5] The instantaneous radiative forcing from 1850 to 2000 is evaluated with the atmospheric component of MIROC, i.e., the atmospheric general circulation model (AGCM), in this study. The MIROC is an AOGCM developed by Center for Climate System Research, University of Tokyo (CCSR), National Institute for Environmental Studies (NIES), and Frontier Research Center for Global Change (FRCGC) [*K-1 Model Developers*, 2004]. The spatial resolution of the atmosphere is T42 (approximately 2.8° by 2.8°

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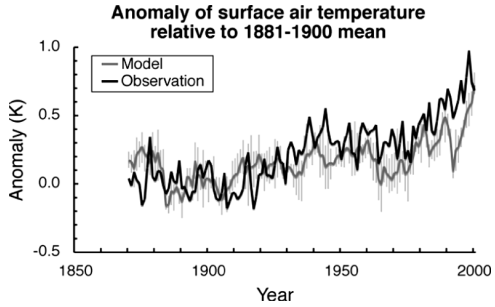

Figure 1. Anomalies of annual global mean surface air temperature relative to 1881–1900 mean by the observations [*Jones and Moberg*, 2003] (black line) and the ensemble mean of the MIROC AOGCM [*Nozawa et al.*, 2005; *Nagashima et al.*, 2006] (gray line). Maximum and minimum ranges from the ensemble simulation are shown by error bars.

in longitude and latitude) with 20 layers of sigma coordinate. The MIROC includes the radiative effects of LLGHGs (CO_2, CH_4, N_2O, and sixteen species of halocarbons) and O_3 [*Nakajima et al.*, 2000]. Changes in LLGHGs concentrations are based on work by *Johns et al.* [2003], and the three-dimensional O_3 data from 1850 to 2000 are prescribed by the simulation of a global chemical climate model, CHASER [*Sudo et al.*, 2002], in the troposphere and based on work by *Randel and Wu* [1999] in the stratosphere. The aerosol transport-radiation model, SPRINTARS, is interactively coupled with the MIROC, which includes explicit calculation of the direct, first indirect, and second indirect effects of aerosols [*Takemura et al.*, 2005]. The anthropogenic emission of BC is edited by T. Nozawa and J. Kurokawa (Historical and future emissions of sulfur dioxide and black carbon for global and regional climate change studies, manuscript in preparation, 2006) and its summary is in the work by *Takemura et al.* [2005]. The anthropogenic organic carbon (OC) emission is proportional to BC with mass ratios for each aerosol source [*Takemura et al.*, 2002]. The historical data of SO_2 emission is in work by *Lefohn et al.* [1999]. The influence of volcanic stratospheric aerosols is included as changes in the optical thickness in the radiative process [*Sato et al.*, 1993]. The simulation from 1850 to 2000 is also forced with changes in solar irradiance [*Lean et al.*, 1995] and land-use [*Hirabayashi et al.*, 2005].

[6] The AGCM is executed to calculate the radiative forcings with assimilating the 6-hourly data of the atmospheric temperature, cloud water content, and cloud fraction and monthly data of the sea surface temperature and sea ice which separately calculated in the 20th century simulation by the AOGCM MIROC [*Nozawa et al.*, 2005]. The radiative forcing of a certain agent is derived from the difference in the radiation budget between the experiments with its concentration for LLGHGs and O_3 or emission for aerosols in the corresponding year and with those in the year 1850. The radiative forcing of LLGHGs in 1950, for example, is estimated by a difference between the experiments with LLGHGs concentrations in 1950 and 1850 using the same prescribed meteorological data and the same concentrations and emissions for the other agents in 1950. This method is consistent with the general estimation of the present-day instantaneous radiative forcing using the present-day climate condition. In the evaluation of the radiative forcing of the aerosol second indirect effect, the cloud water content and cloud fraction are diagnosed and allowed to vary in the AGCM because it could not be calculated if the cloud field was fixed. There is not the consensus on the meteorological field in the calculation of temporal evolution for the radiative forcing, while use of different meteorological data may results in an error of estimation for the radiative forcing even if changes in concentrations of forcing agents are the same.

3. Results

[7] Figure 2a shows a time series of the global mean instantaneous radiative forcing relative to 1850 under all-sky condition at the tropopause. The radiative forcing of

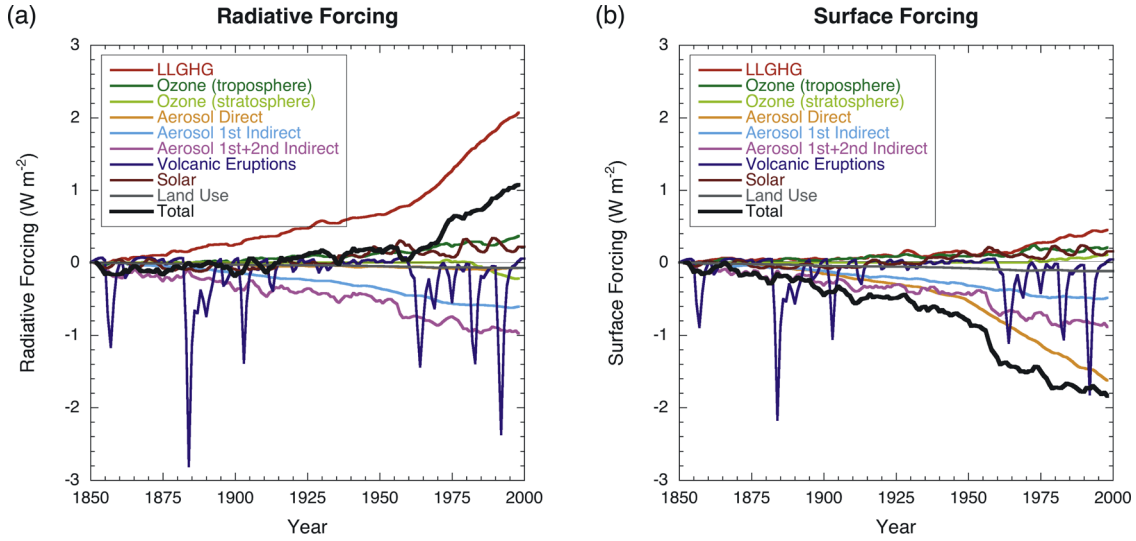

Figure 2. Time evolution of global mean instantaneous radiative forcings from 1850 to 2000 under all-sky condition due to various climate forcing agents at the (a) tropopause and (b) surface in W m^{-2}. They are 5-years running means for annual averages except for stratospheric volcanic aerosols. The total forcing includes effects of all climate forcing agents except for stratospheric volcanic aerosols.

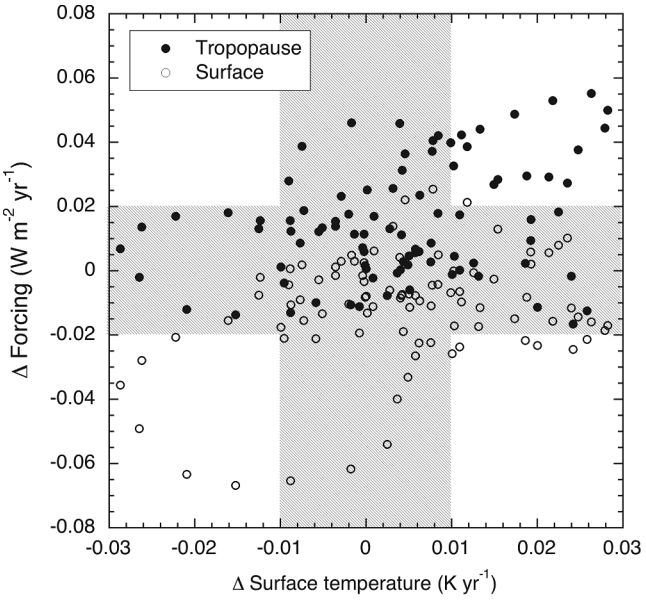

Figure 3. Relationship between annual global mean rates of changes in the simulated surface air temperature in K yr^{-1} and changes in total radiative forcings excluding stratospheric volcanic aerosols at the tropopause (filled circles) and surface (open circles) in W m^{-2} yr^{-1}. Each rate is estimated by a slope of linear regression from five years before to four years after. Rates of changes in the simulated surface air temperature from −0.01 to +0.01 K yr^{-1} and radiative forcings from −0.02 to +0.02 W m^{-2} yr^{-1} are shaded.

LLGHGs is calculated to be +2.1 W m^{-2} in 2000, which is the lower limit of the uncertainty in the IPCC estimation [*Ramaswamy et al.*, 2001]. It is strongly increasing from 1960, 2.6 times in 2000 as large as in 1960. The radiative forcings of an increase in tropospheric O_3 and a decrease in stratospheric O_3 are estimated to be +0.4 and −0.2 W m^{-2}, respectively, which are within a range of the IPCC estimation [*Ramaswamy et al.*, 2001]. Negative forcings of sulfate and OC and a positive forcing of BC by the direct effect nearly cancel the global mean radiative forcing each other, consequently the total forcing due to the aerosol direct effect is small negative forcing of −0.1 W m^{-2} in 2000. The each aerosol direct radiative forcing for sulfate, OC, and BC is estimated to be −0.2, −0.3, and +0.4 W m^{-2}, respectively [*Takemura et al.*, 2005]. The radiative forcings of the first and second aerosol indirect effects in 2000 are estimated to be −0.6 and −0.4 W m^{-2}, respectively. The changing rate of the aerosol indirect effect increases after the mid-20th century. Stratospheric aerosols from explosive volcanic eruptions at Cotopaxi (1855–1856), Krakatau (1883), Santa Maria (1902–1904), Agung (1963), El Chichon (1982), and Pinatubo (1991) made the radiative forcing in excess of −1 W m^{-2}.

[8] The estimated total radiative forcing of LLGHGs, O_3, and aerosols at the tropopause is +1.2 W m^{-2} at the present. The present total radiative forcing relative to 1950 (+1.0 W m^{-2}) is smaller than the estimation by *Hansen et al.* [2002] because the aerosol indirect effect is included in this study. It is principally controlled by LLGHGs and the aerosol indirect effect. The IPCC estimated uncertainties of the radiative forcing by LLGHGs with 10 to 15 %, while it had a wider range from 0 to −2 W m^{-2} at the present without a plausible value for the first aerosol indirect effect and could not be evaluated for the second one because of a much lower confidence level [*Ramaswamy et al.*, 2001]. However, in this study, the estimation is fully linked to the simulation by the AOGCM that successfully reproduces the change in the surface air temperature for past 150 years, so that the radiative forcing due to the aerosol indirect effect in this study can be more reliable than past studies [e.g., *Myhre et al.*, 2001; *Hansen et al.*, 2002]. The estimated contribution of the first indirect effect to the total one is a little larger than the second one.

[9] On the other hand, as shown in Figure 2b, the positive radiative forcing of LLGHGs (+0.4 W m^{-2} in 2000) and O_3 (0.2 W m^{-2} in 2000) at the surface is much smaller than that at the tropopause. Magnitude of the negative forcings of the aerosol indirect effect and volcanic stratospheric aerosols at the surface also slightly decrease in comparison with those at the tropopause. However, the negative forcing of the aerosol direct effect at the surface is much larger, which is estimated to be −1.6 W m^{-2} in 2000. There is little difference in the direct radiative forcing between the tropopause and the surface for OC and sulfate aerosols that principally scatter the incident solar radiation back to the space, whereas BC strongly absorbs the radiation in the troposphere and attenuates the solar radiation reaching the surface [*Takemura et al.*, 2005].

[10] The total radiative forcing of LLGHGs, O_3, and aerosols at the surface is estimated to be −1.7 W m^{-2} in 2000 and rapidly descends between 1955 and 1965. The most important factor on the surface forcing is the aerosol direct effect, and next is the aerosol indirect effect since 1960. The radiative absorption in the troposphere can be understood by consideration of the radiative forcing not only at the tropopause but also at the surface, so that it is significant to show the time evolution of the radiative forcing at the surface which has not presented in past studies [e.g., *Myhre et al.*, 2001; *Hansen et al.*, 2005b]. The global mean of the radiative absorption in the troposphere is estimated to be +3.0 W m^{-2} in this study.

4. Discussion and Conclusions

[11] A remarkable increase in the surface air temperature occurred from 1910 to 1950 and after 1970 (Figure 1), which corresponds with the periods when the positive total radiative forcing at the tropopause rapidly increases and the negative one at the surface does not change much as shown in Figure 2. The increase in the LLGHGs radiative forcing is principally responsible for the increase in the total one at the tropopause after 1970. In contrast, the surface air temperature is stable or slightly reduces from 1950 to 1970 (Figure 1) when the total radiative forcing is fairly flat at the tropopause and sharply goes down at the surface (Figure 2). Both the aerosol direct and indirect effects mainly affect the sharp decline of the total radiative forcing at the surface. *Nagashima et al.* [2006] suggested that carbonaceous aerosols largely contribute to suppression of global warming near the surface in the mid-20th century. Figure 3 clearly summarizes these tendencies. If the large positive changing rates of the surface air temperature occur, the large changing rates of the radiative forcing (top right and bottom right corners without shade in Figure 3)

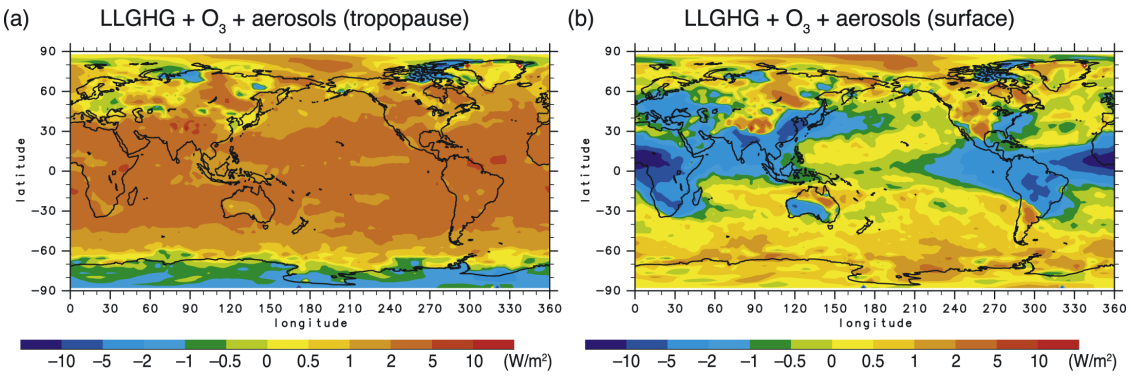

Figure 4. Global distributions of instantaneous radiative forcing from 1850–1854 to 1996–2000 means under all-sky condition due to LLGHGs, O_3, and aerosols at the (a) tropopause and (b) surface in W m^{-2}.

are only in the positive forcing at the tropopause. Figure 3 also indicates that the large changing rates of the radiative forcing are only in the negative forcing at the surface if the large negative changing rates of the surface air temperature occur (top left and bottom left corners without shade in Figure 3). Figures 1–3 suggest that the simultaneous analysis of the rates of changes in the radiative forcing both at the tropopause and surface can indicate changes in the surface air temperature, though it depends on a lot of complicated dynamical and physical feedback mechanisms.

[12] In this study, the time evolution of the radiative forcing from the pre-industrial era to the present day for main climate forcing agents is calculated in conjunction with the simulation of 20th climate by the CCSR/NIES/FRCGC AOGCM. It is indicated that the total radiative forcing is principally controlled by LLGHGs and the aerosol indirect effect at the tropopause and by the aerosol direct and indirect effects at the surface. It is important to clearly present the time evolution of the radiative forcing at the surface as well as at the tropopause in this paper, because variations of the radiation budget in the troposphere can be understood.

[13] While the global distribution of the radiative forcing for LLGHGs is considerably homogeneous, the strong negative aerosol radiative forcing is concentrated around emission source regions because of short lifetime of aerosol particles about several days in the atmosphere (Figure 4). The negative forcing at the surface is remarkable over densely populated areas in the Northern Hemisphere and over biomass burning areas in Africa and South America principally due to the aerosol direct and indirect effects. Therefore, in the next study, we need to analyze the radiative forcing at the surface as well as at the tropopause for each region, because aerosol amounts and species are spatially heterogeneous and then the aerosol effects on atmospheric radiation and clouds are largely different.

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