Aerosol radiative forcing and climate sensitivity deduced from the Last Glacial Maximum to Holocene transition

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[1] We use the temperature, carbon dioxide, methane, and dust concentration record from the Vostok ice core to deduce the aerosol radiative forcing during the Last Glacial Maximum (LGM) to Holocene transition and the climate sensitivity. A novel feature of our analysis is the use of a cooling period between about 42 KYBP (thousand years before present) and LGM to provide a constraint on the aerosol radiative forcing. We find the change in aerosol radiative forcing during the LGM to Holocene transition to be 3.3 ± 0.8 W/m^2 and the climate sensitivity between 0.36 and 0.68 K/Wm^{-2} with a mean value of 0.49 ± 0.07 K/Wm^{-2}. This suggests a 95% likelihood of warming between 1.3 and 2.3 K due to doubling of atmospheric concentration of CO_2. The ECHAM5 model simulation suggests that the aerosol optical depth during the LGM may have been almost twice the current value (increase from 0.17 to 0.32).
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1. Introduction

[2] The increasing atmospheric concentration of carbon dioxide and other atmospheric greenhouse gases has been well documented. Their positive radiative forcing at the top-of-atmosphere can be reasonably well estimated using detailed radiative transfer codes. However, a large uncertainty remains when we try to translate this top-of-atmosphere radiative forcing to changes of the global average of the surface temperature.

[3] Climate sensitivity in K/Wm^{-2} specifies the equilibrium temperature change (in K) per unit change of the radiative forcing (in W/m^2). If the climate system behaves like a black body radiator the climate sensitivity is 0.30 K/Wm^{-2}. Various climate feedbacks (including the water vapor feedback and cloud feedback) modify the climate sensitivity with the positive feedbacks amplifying the temperature change and negative feedbacks reducing it.

[4] An alternative definition of the climate sensitivity, used predominantly by the climate modeling community to compare performance of individual climate models, is the global mean equilibrium temperature increase in response to a doubling of the atmospheric CO_2 concentration from its pre-industrial value. This equilibrium climate sensitivity has been deduced on the basis of simulations performed using different climate models coupled to ocean models in which CO_2 was doubled and the models were run to equilibrium. The equilibrium climate sensitivity is currently believed to be (with 66% probability) within the range of a temperature increase between 2 and 4.5 K [*Solomon et al.*, 2007] for doubling of CO_2. Considering that the radiative forcing for the doubling of CO_2 is 3.7 ± 0.5 W/m^2 [*Solomon et al.*, 2007], the corresponding likely range of the IPCC suggested climate sensitivity is between 0.48 and 1.4 K/Wm^{-2}.

[5] One possible way to narrow the limits of climate sensitivity is to use the reconstruction of the past climate based on ice core records. The climate transition from the last glacial maximum (LGM) to the present interglacial period (Holocene) has been used for this purpose in the past [*Harvey*, 1988; *Hoffert and Covey*, 1992; *Hewitt and Mitchell*, 1997; *Harrison et al.*, 2001; *Claquin et al.*, 2003]. In these studies, the considered radiative forcing at the LGM to Holocene transition includes the solar output, radiative forcing due to the increase in greenhouse gases, radiative forcing due to changes in Earth's albedo (including ice sheets extent and vegetation changes), and radiative forcing due to changes in atmospheric aerosol loading. Effects of aerosols on the hydrological cycle and on the radiative forcing still have large uncertainties [*Forster et al.*, 2007; *Lohmann and Feichter*, 2005; *Penner et al.*, 2004]. One of the uncertainties in the radiative forcing calculation during the LGM to the Holocene transition is the radiative forcing due to increased aerosol optical depth during the peak of the last ice age (LGM). *Harvey* [1988] estimates the change in the aerosol radiative forcing between the present-day and the LGM to be between -1.9 and -3.3 W/m^2. *Hoffert and Covey* [1992] and *Hansen et al.* [2002] estimate the change in aerosol forcing to be around -1 W/m^2, while *Hewitt and Mitchell* [1997] neglect the aerosol forcing all together. *Harrison et al.* [2001] and *Claquin et al.* [2003] re-emphasize the importance of dust aerosol forcing during the LGM to Holocene transition and suggest that at least in the tropics (for 30°S to 30°N) the dust aerosol forcing may be as strong as the forcing due to an increase in greenhouse gases.

[6] In the following analysis we use the two adjacent time periods, the LGM to Holocene transition and the cooling period between 41710 YBP and the LGM to deduce the change in aerosol radiative forcing and to estimate climate sensitivity. We use the temperature, CO_2, CH_4 and dust concentration data provided by the Vostok ice core [*Petit et al.*, 1999]. We assume that the climate sensitivity is the same for both time periods. From two separate data sets (LGM to Holocene, and the warm period around 42 KYBP

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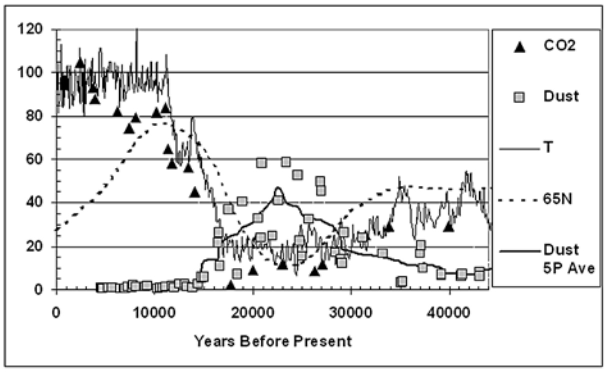

Figure 1. Vostok ice core data for changes in temperature (in units of 0.1 K), carbon dioxide atmospheric concentration (in ppmv), and dust aerosols (in arbitrary units normalized to value of one for an average Holocene concentration), and relative changes in summer solar insulation (in W/m^2) at the latitude of 65°N (dashed line). A solid thick black line shows a five point running average of dust aerosol concentration. Ice core data from *Petit et al.* [1999]; insolation from *Berger et al.* [1993].

to LGM transition) we compute the aerosol radiative forcing and estimate climate sensitivity.

2. LGM to Holocene Climate Transition

[7] In our analysis we use the reconstruction of the past climate during the last 42,000 years based on the Vostok ice core records. The Vostok record represents polar conditions for the last 420,000 years, it includes four glacial to interglacial transitions and it demonstrates the universality of temperature, greenhouse gases and aerosol concentration changes during these transitions [*Petit et al.*, 1999].

[8] To estimate the aerosol forcing during the LGM to Holocene transition we use the temperature, GHGs and aerosol concentrations at three defined time intervals. Figure 1 shows a period of cooling from about 42 KYBP to the LGM, followed by a warming leading to the current Holocene interglacial period. The changes in temperature, CO_2, and CH_4 atmospheric concentrations are fairly accurately characterized by the Vostok ice core data. Variability of dust concentration is determined only qualitatively in relative units, which cannot easily be translated into the changes of aerosol optical depth or aerosol radiative forcing. The changes in surface albedo are the most uncertain because they cannot be deduced from the ice core and have to be approximated from past model results. The estimated changes in temperature, CO_2, CH_4, dust concentration, and surface albedo for these periods are discussed in more detail below.

2.1. Temperature

[9] The temperature changes during the past 44,000 years derived from the Vostok ice core [*Petit et al.*, 1999] are shown in Figure 1 (in units of 0.1 K). The maximum interglacial Holocene temperature is about 10.2 K above the LGM minimum temperature. A relatively warm epoch (lasting over 600 years) centered around the year 41710 YBP is about 4.8 K above the minimum LGM temperature. The polar region temperature changes are about twice as large as the global temperature [*Chylek and Lohmann*, 2005]; however, the exact ratio of the global to Vostok temperature change is not known. To account for this uncertainty, we consider three scenarios in which the global temperature difference between the Holocene and the LGM is taken to be 4.1, 4.6 and 5.1 K. This is in agreement with past investigations suggesting the global temperature difference between the Holocene and the LGM to be at most 5 K [*Hewitt and Mitchell*, 1997]. The global temperature change between the warm maximum near 42 KYBP and the LGM is set to be 2.4 K (half of the observed difference at the Vostok site) and then the difference is decreased to 2.16 and 1.93 K, keeping the ratio of the temperature difference between the two considered climate transitions (LGM to Holocene and the warm period 42 KYBP to LGM) constant.

2.2. Greenhouse Gases

[10] Carbon dioxide concentration in air bubbles from the Vostok ice core suggests CO_2 levels of 285, 182 and 209 ppmv for the pre-industrial Holocene, the LGM, and the warm period around 42 KYBP, respectively. The change in radiative forcing due to the change in carbon dioxide concentration is estimated using the approximation [*Myhre et al.*, 1998]

$$\Delta F_{CO2} = 5.35 \ln(C_1/C_2), \tag{1}$$

where the change in the radiative forcing, ΔF_{CO2}, is in W/m^2, and C_1 and C_2 are the CO_2 concentrations in ppmv before and after the considered climate transition. We obtain the radiative forcing due to CO_2 to be 2.40 W/m^2 for the LGM to Holocene transition and 0.74 W/m^2 for the transition between the LGM and the warm period 41710 YBP.

[11] The methane (CH_4) concentrations during the Holocene, LGM and the warm period at 42 KYBP are 667, 340 and 548 ppbv, respectively [*Petit et al.*, 1999]. The change in radiative forcing due to the change in CH_4 concentration is estimated using the approximation

$$\Delta F_{CH4} = 0.036\left[(M_1)^{1/2} - (M_2)^{1/2}\right], \tag{2}$$

where the change in the radiative forcing, ΔF_{CH4}, is in W/m^2, and M_1 and M_2 are the CH_4 concentrations in ppbv before and after the considered transition. We obtain the radiative forcing due to CH_4 to be 0.27 W/m^2 for the LGM to Holocene transition and 0.19 W/m^2 for the warm period at 42 KYBP to LGM transition.

[12] Thus the total radiative forcing of greenhouse gases (GHGs) between the LGM and Holocene is 2.67 W/m^2 and 0.93 W/m^2 for the warm period of 42 KYBP to LGM transition.

2.3. Surface Albedo

[13] Unfortunately, we have no direct information concerning the past global surface albedo from the ice core data. Consequently, we have to rely on past modeling results. The radiative forcing due to the surface albedo changes (extent of ice sheets, sea ice and snow cover, exposure of a new land in a low sea level state, change in surface characteristics and vegetation cover) has been esti-

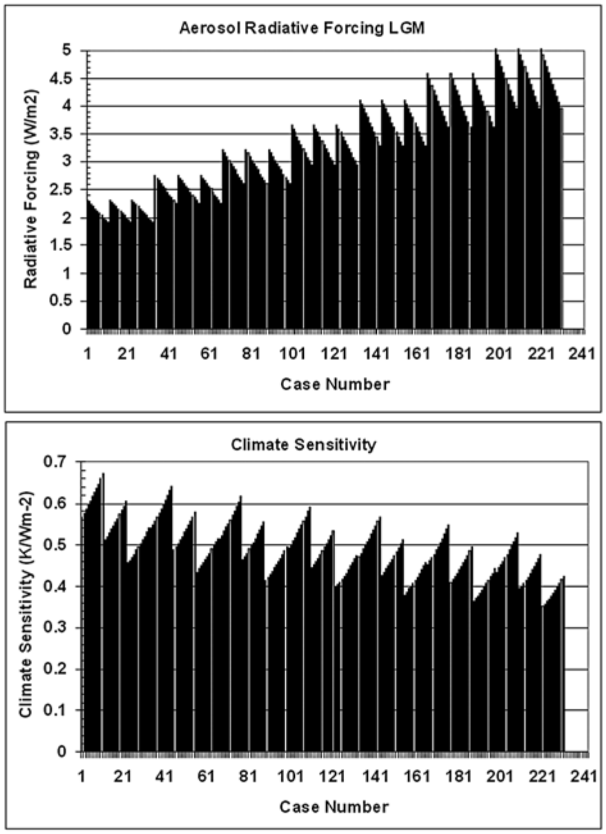

Figure 2. (top) Aerosol radiative forcing (W/m^2) during the LGM to Holocene transition for different values of LGM to Holocene temperature difference and different values of the surface albedo changes (as described in the text). (bottom) Climate sensitivity derived from our analysis for different values of LGM to Holocene temperature difference and different values of the surface albedo changes.

mated by several authors to be between 3 and 4 W/m^2 with most results clustering around 3.5 W/m^2 [*Hewitt and Mitchell*, 1997]. To account for a wide range of possible values we consider 11 different cases of the forcing with the values between 3 and 4 W/m^2 in increments of 0.1 W/m^2. For the change of surface albedo between the LGM and the warm period 42 KYBP we consider a wide range of possibilities assuming that the radiative forcing (due to surface albedo changes during this transition) did not change at all, or changed in proportion to the albedo radiative forcing of the LGM to Holocene transition with the proportionality factor of 0.05, 0.10, 0.15, 0.20, 0.25, or 0.30. Altogether we consider 77 possible combination of the forcing due to albedo changes. Multiplied by the three temperature differences for the LGM to Holocene transition, we consider a total of 231 different scenarios.

2.4. Aerosol Concentration

[14] Relative aerosol concentrations as measured in the Vostok ice core (in μg/g) are shown in Figure 1. The data are normalized (for the purpose of plotting) to the value of one for the average concentration during the years 0 to 10,000 YBP. The difference between the Holocene and the LGM is 58 units (Figure 1), while the difference between the LGM and the warm period 42 KYBP is 53 units. The translation between the radiative forcing in W/m^2 and the introduced arbitrary units is considered as an unknown to be determined in the following analysis.

[15] We note (Figure 1) that the decrease in aerosol concentration at the termination of the last glacial period started approximately at the same time as an increase in summer solar insolation at 65°N, which according to the Milankovitch hypothesis initiates the transition from the glacial to interglacial stage. It seems that the decrease in the dust atmospheric concentration had produced the first positive radiative forcing impulse that might have contributed to the termination of the glacial state.

2.5. Climate Sensitivity

[16] We assume that the climate system has reached a state close to equilibrium during the Holocene, during the LGM, and during about 600 years long period characterized by a peak in temperature close to 41710 YBP. The equilibrium climate sensitivity, λ, can be written as

$$\lambda = \Delta T / \Delta F \tag{3}$$

where ΔT is the difference between an average global temperature of two equilibrium states, and ΔF is the radiative forcing difference considered to be a sum of all agents facilitating the climate transition

$$\Delta F = \Delta F_{GHG} + \Delta F_{Albedo} + \Delta F_{Aerosol} \tag{4}$$

Assuming the climate sensitivity to remain the same during the two considered climate transitions, we obtain (from equation 3) an equation

$$\frac{\Delta T_1}{F_{GHG1} + F_{ALB1} + 58X} = \frac{\Delta T_2}{F_{GHG2} + F_{ALB2} + 53X} \tag{5}$$

from which we calculate X, the radiative forcing per arbitrarily chosen unit of aerosol concentration. We find the average value X = 0.056 W/m^2. The change in aerosol radiative forcing for the LGM to Holocene transition is then $\Delta F_1 = 58X$. The climate sensitivity is obtained from Equation (5). The results for all 231 considered cases are shown in Figure 2.

[17] We find the aerosol radiative forcing during the LGM to Holocene transition to be 3.3 W/m^2 with a standard deviation of 0.8 W/m^2. The climate sensitivity is thus limited to values between 0.36 and 0.68 K/Wm^{-2} (with the mean value of 0.49 K/Wm^{-2} and standard deviation of 0.07 K/Wm^{-2}). Thus the LGM to Holocene transition suggests the climate sensitivity between 1.3 and 2.5 K with the mean value of 1.8 K for doubling of the atmospheric CO_2.

3. Previous Glacial to Interglacial Transitions

[18] Vostok ice core also provides information (temperature, CO_2, CH_4 and dust concentrations) concerning the three previous glacial to interglacial transitions [*Petit et al.*, 1999]. Table 1 summarizes the appropriate dates and values of relevant parameters. We supplement these data with the deduced value of the radiative forcing per unit of aerosol

Table 1. Ice Core Data Used[a]

YBP	ΔT, K	CO_2, ppmv	CH_4, ppbv	ΔDust, relative units	λ, K/Wm^{-2}
11,191	9.2	182/285	340/667	58	0.49
138,193	12.4	190/287	318/710	40	0.67
245,834	9.4	185/279	380/650	33	0.60
333,602	11.2	186/299	342/773	34	0.60

[a]The first column specifies an approximate time in YBP of the beginning of the interglacial period, ΔT is the corresponding temperature difference, CO_2 column gives the CO_2 concentration (in ppmv) before and after the transition, CH_4 provides similar data for CH_4 concentration (in ppbv), and ΔDust indicates a change in relative dust units during the transition. The last column shows the climate sensitivity, λ, deduced from the considered glacial to interglacial transition.

concentration ($X = 0.056$ W/m^2) and calculate the appropriate change in radiative forcing between the corresponding glacial to interglacial transitions. The mean values of climate sensitivities deduced from these glacial to interglacial transitions that occurred about 140 KYBP, 250 KYBP and 330 KYBP are 0.67, 0.60 and 0.60 K/Wm^{-2}, respectively, with a standard deviation of 0.07 K/Wm^{-2} (corresponding to a temperature increase due to the doubling of CO_2 of 2.2 and 2.6 K, respectively, with an uncertainty of ± 0.3 K). At this time it is not clear whether these higher sensitivities, compared to the climate sensitivity deduced from the LGM to Holocene transition, really reflect higher climate sensitivity at the time of the considered climate transitions or whether they are artifacts due to imperfect ice core data and uncertainties in the used approximations.

4. General Circulation Model (GCM) Simulation

[19] To obtain a better understanding of the aerosol radiative forcing during the LGM to Holocene transition we have used the ECHAM5 GCM [*Lohmann et al.*, 2007] model to simulate aerosol radiative forcing and its geographical distribution. The simulations were carried out in T42 horizontal resolution ($\sim 2.8° \times 2.8°$) over 10 years. To simulate the high dust concentration in the ice core during the LGM, we have increased the strength of the current dust sources by a factor of 4, and the fluxes of DMS and sea salt both by a factor of 2,

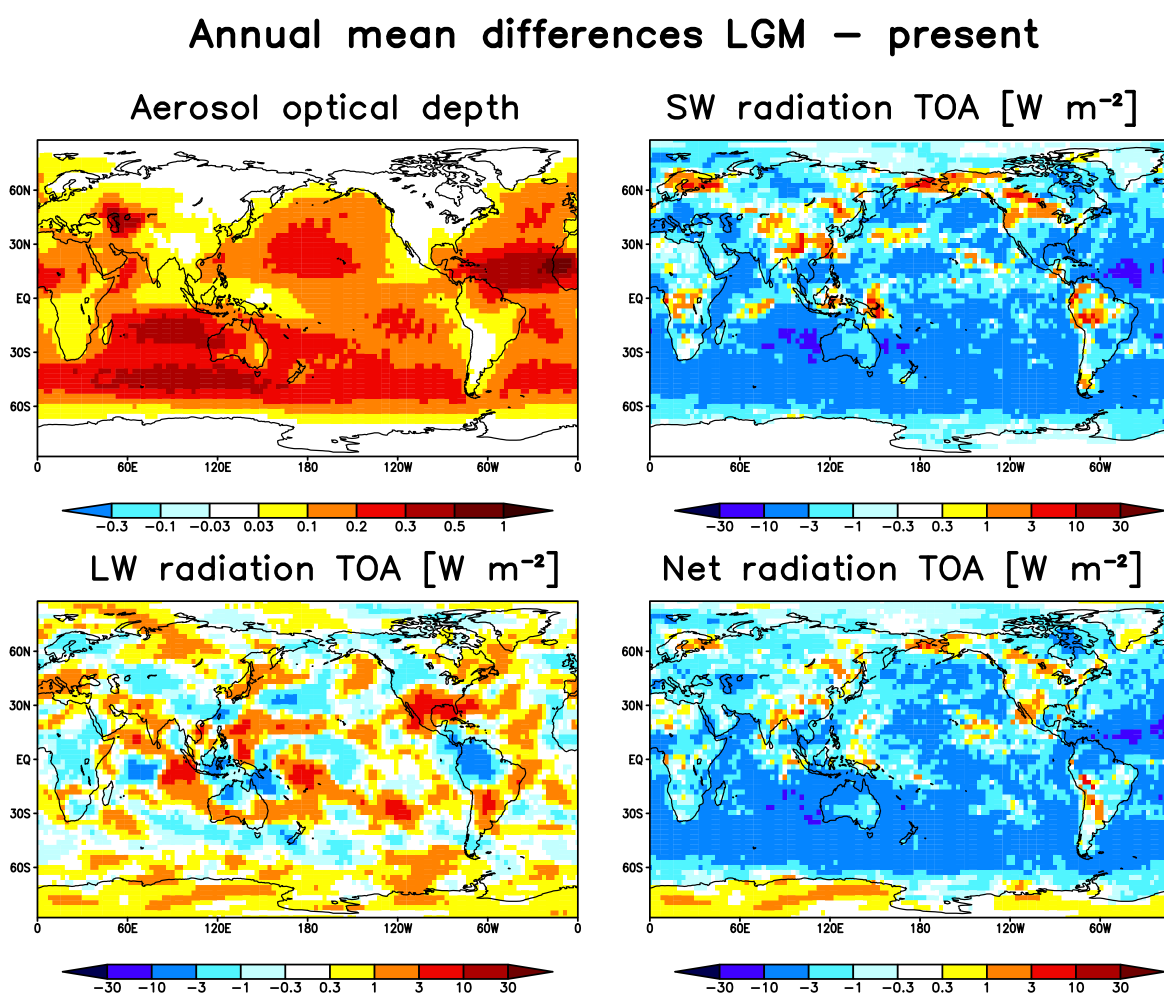

Figure 3. ECHAM5 model simulation of the annual mean difference in dust aerosol depth and the radiative forcing between the LGM with enhanced aerosol concentration and the present-day.

so that the average aerosol optical depth has almost doubled (increased from 0.17 to 0.32). The dust burden increased roughly by a factor of 3, which is comparable to the 2.5 higher dust loadings reported in the modeling study [*Mahowald et al.*, 1999]. The other aerosol sources and the greenhouse gas concentrations were held constant. The water vapor feedback causes the vertically integrated water vapor mass to be reduced by 0.15% during the LGM. The ECHAM5 model shows that the aerosols cause a strong regional radiative cooling especially over the oceans of up to 30 W/m^2 (Figure 3). The radiative forcing is dominated by the short-wave radiation that is responsible for the average radiative cooling of about 3.1 W/m^2, while the long-wave radiation increased by 0.1 W/m^2 in the global mean. The total aerosol radiative forcing in the ECHAM5 simulation produces an aerosol cooling of 3.0 W/m^2 which is within the standard deviation of the radiative forcing derived in our analysis of the ice core data.

5. Discussion and Summary

[20] We have shown that the ice core data from the warm period (around 42 KYBP) to the LGM and from the LGM to Holocene transition can be used to constrain the dust aerosol radiative forcing during these transitions. We find the dust radiative forcing to be 3.3 ± 0.8 W/m^2. Assuming that the climate sensitivity is the same for both transitions, we obtain $\lambda = 0.49 \pm 0.07$ K/Wm^{-2}. This suggests 95% likelihood of warming between 1.3 and 2.3 K due to doubling of atmospheric concentration of CO_2 (assuming that the CO_2 doubling produces the radiative forcing of 3.7 W/m^2 according to the IPCC 2007 report). The ECHAM5 model simulation suggests that during the LGM the global average aerosol optical depth might have been almost twice the current value.

[21] The results compatible with climate sensitivity around or below 2 K for doubling of CO_2 were recently deduced using cloud resolving models incorporated within GCMs [*Miura et al.*, 2005; *Wyant et al.*, 2006], from observational data [*Chylek et al.*, 2007; *Schwartz*, 2007], and from a set of GCM simulations constrained by the ERBE (Earth Radiation Budget Experiment) observations [*Forster and Gregory*, 2006]. All these results together with our work presented in this paper support the lower end of the climate sensitivity range of 2 to 4.5 K suggested by the IPCC 2007 report [*Solomon et al.*, 2007].

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