Limits on climate sensitivity derived from recent satellite and surface observations

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[1] An analysis of satellite and surface measurements of aerosol optical depth suggests that global average of aerosol optical depth has been recently decreasing at the rate of around 0.0014/a. This decrease is nonuniform with the fastest decrease observed over the United States and Europe. The observed rate of decreasing aerosol optical depth produces the top of the atmosphere radiative forcing that is comparable to forcing due to the current rate of increasing atmospheric concentration of carbon dioxide and other greenhouse gases. Consequently, both increasing atmospheric concentration of greenhouse gases and decreasing loading of atmospheric aerosols are major contributors to the top-of-atmosphere radiative forcing. We find that the climate sensitivity is reduced by at least a factor of 2 when direct and indirect effects of decreasing aerosols are included, compared to the case where the radiative forcing is ascribed only to increases in atmospheric concentrations of carbon dioxide. We find the empirical climate sensitivity to be between 0.29 and 0.48 K/Wm^{-2} when aerosol direct and indirect radiative forcing is included.

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1. Introduction

[2] A climate forcing is defined as a perturbation of the Earth's energy balance and it is usually quantified in W/m^2 at the top of the atmosphere. The climatological consequences of such forcing vary from region to region, however, the average climate response to the applied forcing is usually expressed as the change in the average global surface temperature. The climate sensitivity (in K/Wm^{-2}) is a measure characterizing how the global top-of-atmosphere radiative forcing is translated into a change in the annual mean global surface temperature. It is a conversion factor between implied forcing and global temperature response. Thus the climate sensitivity can be understood as climate response per unit radiative forcing. There are several definitions of the climate sensitivity depending on which of the feedback processes are included. Fast feedback (acting on a scale of a few years) include water vapor and cloud response, while ocean heat uptake is a dominant slow feedbacks acting on a scale of a few hundreds to thousands years.

[3] The equilibrium climate sensitivity (in K) is usually defined as a change in the equilibrium annual global surface temperature when carbon dioxide has been doubled and held constant after that. Alternatively, the equilibrium climate sensitivity can be express (in K/Wm^{-2}) as an equilibrium temperature change per unit radiative forcing when atmospheric CO_2 concentration is doubled. The climate sensitivity is not prescribed in global climate models but follows from the parameterization of various physical processes in these models. The equilibrium climate sensitivity is a useful parameter for comparing climate models. Climate system without feedbacks would have an equilibrium climate sensitivity of 0.3 K/Wm^{-2} corresponding to the global average warming of about 1.1 K for doubling of CO_2. Early radiative convective climate models [*Manabe and Strickler*, 1964; *Manabe and Wetherald*, 1967] suggested an increase in an equilibrium global average surface temperature after doubling of CO_2 between 1.3 (with no water vapor feedback) and 2.3 K (while keeping relative humidity constant). The National Research Council climate assessment [*Charney et al.*, 1979] suggested equilibrium climate sensitivity (ΔT for doubling of CO_2) to be within the range of 1.5 to 4.5 K. An early study [*Cess et al.*, 1989] comparing 14 different atmospheric general circulation models obtained equilibrium climate sensitivity parameters in the range from 0.39 K/Wm^{-2} to 1.11 K/Wm^{-2}. Recently the equilibrium climate sensitivity of climate models has been estimated by atmospheric general circulation models coupled to mixed layer oceans, as well as by fully coupled atmosphere-ocean models with dynamic oceans. According

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to *Intergovernmental Panel on Climate Change* (*IPCC*) [2007], the likely range of global equilibrium temperature increase for doubling of CO_2 is between 2.0 and 4.5K, with values below 1.5K considered very unlikely. Since the doubling of CO_2 causes direct radiative forcing of about 3.7 W/m^2 [*IPCC*, 2007], the range of 2.0 to 4.5K for doubling of CO_2 corresponds to climate sensitivity between 0.54 and 1.22 K/Wm^{-2}. However, some experiments with cloud resolving models embedded within GCMs [*Miura et al.*, 2005; *Wyant et al.*, 2006] suggested lower climate sensitivity with values of 0.44 and 0.41 K/Wm^{-2}, respectively. Unfortunately, the equilibrium climate sensitivity cannot be obtained directly from observations, since the Earth's climate system is always in the process of change, never reaching equilibrium.

[4] To eliminate the necessity to run the models up to equilibrium, an alternative in the form of a transient climate response has been introduced [*IPCC*, 2007]. The transient climate response is defined as the global mean temperature change which occurs at the time of carbon dioxide doubling when the CO_2 concentration is increased by 1% each year until the doubling is reached. The transient climate response is generally lower than equilibrium climate sensitivity, since the long time component of the climate response to considered forcing has not been yet fully realized.

[5] In this paper we derive empirical climate sensitivity (in K/Wm^{-2}) from the current climate observations. We allow for slow feedbacks [*Kim et al.*, 1992] using model estimates of the ocean heat uptake [*Raper et al.*, 2002]. We use the term empirical climate sensitivity to indicate that the climate sensitivity considered in this paper is based on empirical observations and that is not necessarily equal to equilibrium climate sensitivity deduced from climate models.

[6] Increasing atmospheric CO_2 causes a positive radiative forcing, leading to a warmer climate and a higher annual mean surface temperature. A part of the CO_2 induced warming since the beginning of industrialization has been compensated by a simultaneous cooling effect of aerosols, particularly sulfate from power plants. A potential overestimation of carbon dioxide warming in climate models can be compensated by a similar overestimate of aerosol cooling, still allowing good agreement with the observed global temperature change. Consequently, during the period of simultaneously increasing CO_2 and increasing atmospheric aerosol loading, there are fewer constraints imposed on estimates of climate sensitivity due to CO_2 and aerosol compensating effect.

[7] However, the situation has become quite different in recent years, when the global average aerosol optical depth, as indicated by satellite and ground-based instruments, has been decreasing. A decreasing aerosol optical depth is now causing a warming in a similar way as increasing carbon dioxide does. In this case an overestimate of the climate sensitivity would lead to a disagreement with the observed temperature rise. Additional simplifications are the lack of volcanic aerosols (the last significant volcanic eruption Mount Pinatubo occurred in 1991), stable atmospheric CH_4 concentration [*IPCC*, 2007], and relatively small changes in solar radiative output [*Fröhlich and Lean*, 2004]. This leaves the increasing concentration of carbon dioxide and the decreasing trend of atmospheric aerosol loading as major drivers for the recent global warming scenario.

2. Climate Sensitivity

[8] We start with the energy balance equation

$$\frac{d}{dt}(\Delta Q) = \Delta F - \frac{1}{\lambda}\Delta T \tag{1}$$

where ΔQ is the change in heat content of the climate system, ΔF is the radiative forcing at the top of the atmosphere, ΔT is the change in the annual global mean surface temperature, and λ is a climate sensitivity. For a system in equilibrium, $d(\Delta Q)/dt = 0$ and λ becomes the equilibrium climate sensitivity λ_{eq}, given by

$$\lambda_{eq} = \frac{\Delta T_{eq}}{\Delta F} \tag{2}$$

where ΔT_{eq} is the temperature difference between two equilibrium states.

[9] Since the heat storage of the climate system is dominated by the ocean, the $d(\Delta Q)/dt$ term in equation (1) can be approximated by the heat flux into the ocean. With this approximation the energy balance becomes

$$\Delta H = \Delta F - \frac{1}{\lambda}\Delta T \tag{3}$$

where ΔH is the change of the heat flux into the ocean. There is a considerable uncertainty in the ocean heat content changes during the past few decades [*Gouretski and Koltermann*, 2007] (because of limited geographical data available, short time series, and change of instrumentation). Following *Raper et al.* [2002], we assume that the ocean heat input term can be written in the form

$$\Delta H = \kappa \Delta T \tag{4}$$

with the value of the ocean heat uptake efficiency from zero to 1.4 Wm^{-2} in the range K. From equations (3) and (4) we obtain the climate sensitivity [*Knutti et al.*, 2002; *Hegerl et al.*, 2006; *Forster and Gregory*, 2006]

$$\lambda = \frac{\Delta T}{\Delta F - \kappa \Delta T} \tag{5}$$

In the following we estimate the climate sensitivity from the current changes of the radiative forcing, ΔF, different values for the ocean heat uptake efficiency and the current rate of the annual global mean surface temperature change, ΔT. We use term empirical climate sensitivity for sensitivity, λ, defined in the above-described way.

3. Radiative Forcing Due to Carbon Dioxide

[10] Radiative forcing due to increasing carbon dioxide can be estimated using a semiempirical formula [*Myhre et*

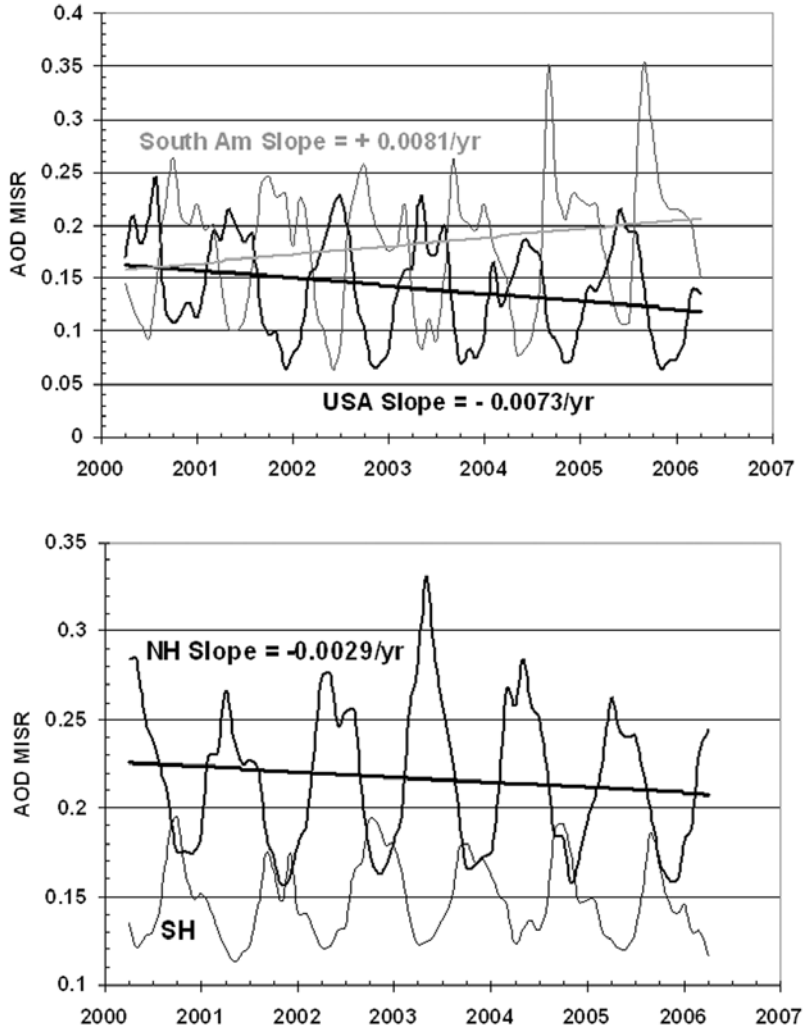

Figure 1. (top) A decreasing AOD trend over the USA (thick black line) and increasing trend over South America (gray line) from the MISR AOD data. (bottom) A decreasing AOD trend over the Northern Hemisphere (black thick line) and effectively no change over the Southern Hemisphere (gray line) suggested by the MISR data.

al., 1998] that approximates fairly accurately the detailed radiative transfer calculation

$$\Delta F = 5.35 \ln(C/C_0) \mathrm{W/m^2} \tag{6}$$

where C and C_o are the current and the reference carbon dioxide concentrations. The average annual increase of the CO_2 concentration within the 1960–2005 time span has been around 1.4 ppmv. The rate of increase has recently accelerated, and during the past decade the annual average increase has been about 1.9 ppmv [*IPCC*, 2007]. Taking the reference concentration, C_o, to be around 375 ppmv and an average recent annual increase of 1.9 ppmv, we obtain for the carbon dioxide induced top-of-atmosphere of radiative forcing rate:

$$\left(\frac{\Delta F}{\Delta t}\right)_{CO_2} = 5.35 \ln(376.9/375) = 0.027 \ \mathrm{Wm^{-2}/yr} \tag{7}$$

This yields a forcing for the past decade of $(\Delta F)_{CO2}$ = 0.27 $\mathrm{Wm^{-2}}$.

[11] Over the past 2 decades the methane growth rate in the atmosphere have decreased and the methane concentration has been stable over at least the past 10 a [*IPCC*, 2007]. Consequently, for the past decade, the CO_2 radiative forcing represents almost all greenhouse gas forcing.

[12] The annual rate of mean surface temperature increase over the past decade has been observed to be [*IPCC*, 2007]

$$\frac{\Delta T}{\Delta t} = 0.018 \ \mathrm{K/yr} \tag{8}$$

that is ΔT = 0.18 K for the past decade. Taking an average value of the ocean heat uptake efficiency [*Raper et al.*, 2002] of κ = 0.7 $\mathrm{Wm^{-2}/K}$, and hypothetically assuming that all the warming observed during the past decade has been

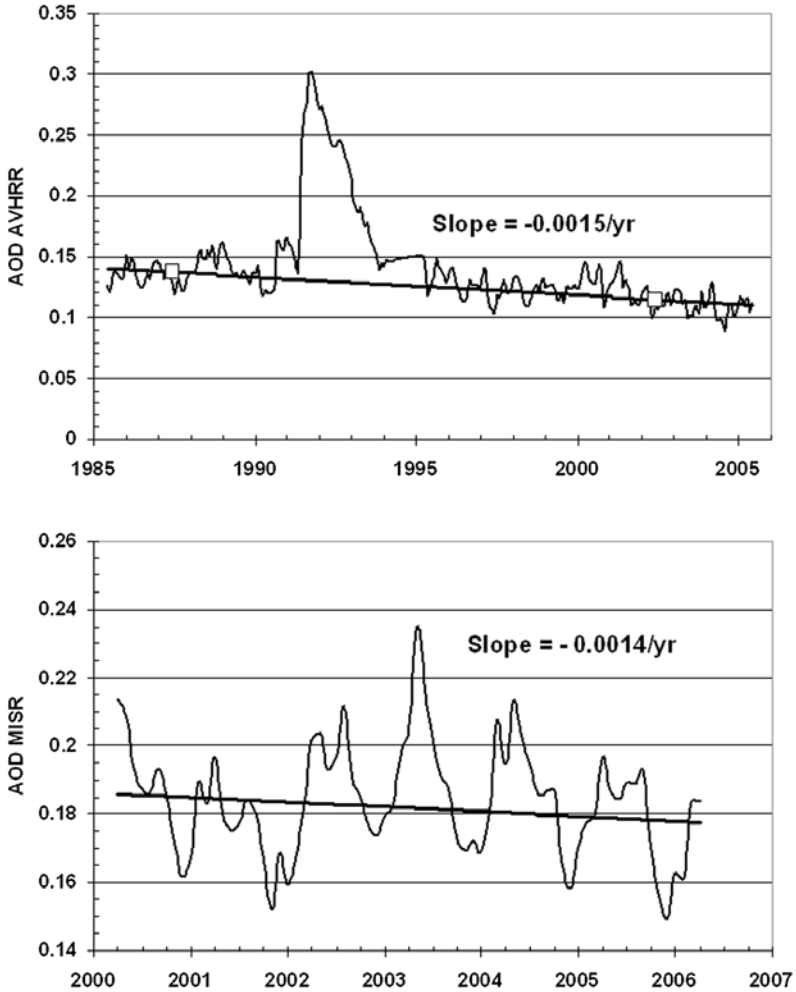

Figure 2. (top) The AVHRR data since 1985 suggests a declining global average of the AOD at the rate of −0.0015/a, while (bottom) the MISR data, available since early 2000, show a declining trend of −0.0014/a. To eliminate the Mt. Pinatubo effect on the AVHRR data, we have calculated an average AOD for years 1985–1990 and 2000–2005 (squares) and interpolated a linear trend through these two points. The AVHRR data are for AOD over the ocean, while the MISR data are for combined ocean and land.

due to CO_2 increase (neglecting effect of aerosols), we obtain from equation (5) the empirical climate sensitivity

$$\lambda = 1.25\ \mathrm{K/Wm^2} \tag{9}$$

4. Aerosol Optical Depth

[13] The global average aerosol optical depth (AOD) has been found to be decreasing during the past decade [*Mishchenko et al.*, 2007; *Mishchenko and Geogdzhayev*, 2007], probably because of efforts to clean the atmosphere of sulfate aerosols in North America and Europe. The observed decrease is consistent with reduction in anthropogenic emissions of SO_2 across North America and Europe [*Holland et al.*, 1998; *Stern*, 2006; *Manktelow et al.*, 2007]. Because of the short aerosol atmospheric lifetime, the decrease is not expected to be uniform over the globe.

[14] A significant decrease of aerosol optical depth is found (Figure 1) over the USA, whereas an increase has been documented over South America. The Northern Hemisphere data as a whole suggest a decreasing aerosol trend, whereas the Southern Hemisphere does not show any change in AOD during the 2000–2006 time span. The NASA Earth Observing System's Multiangle Imaging SptecroRadiometer (MISR) provides validated midvisible AOD globally over land and water [e.g., *Kahn et al.*, 2005], but only since the year 2000, and with processing that is undergoing further refinement. However, the 25-a AVHRR AOD record was recently analyzed, and after elimination of Mt. Pinatubo aerosol peak, a statistically significant decreasing trend has been established [*Mishchenko et al.*, 2007]. We note that the rate of the MISR global AOD decrease of −0.0014/a (since 2000) is essentially identical to the rate of −0.0015/a suggested by the AVHRR data (Figure 2).

[15] An additional confirmation of the AOD decreasing tendency during the past decades comes from the reversal of "solar dimming" to "solar brightening" [*Alpert et al.*,

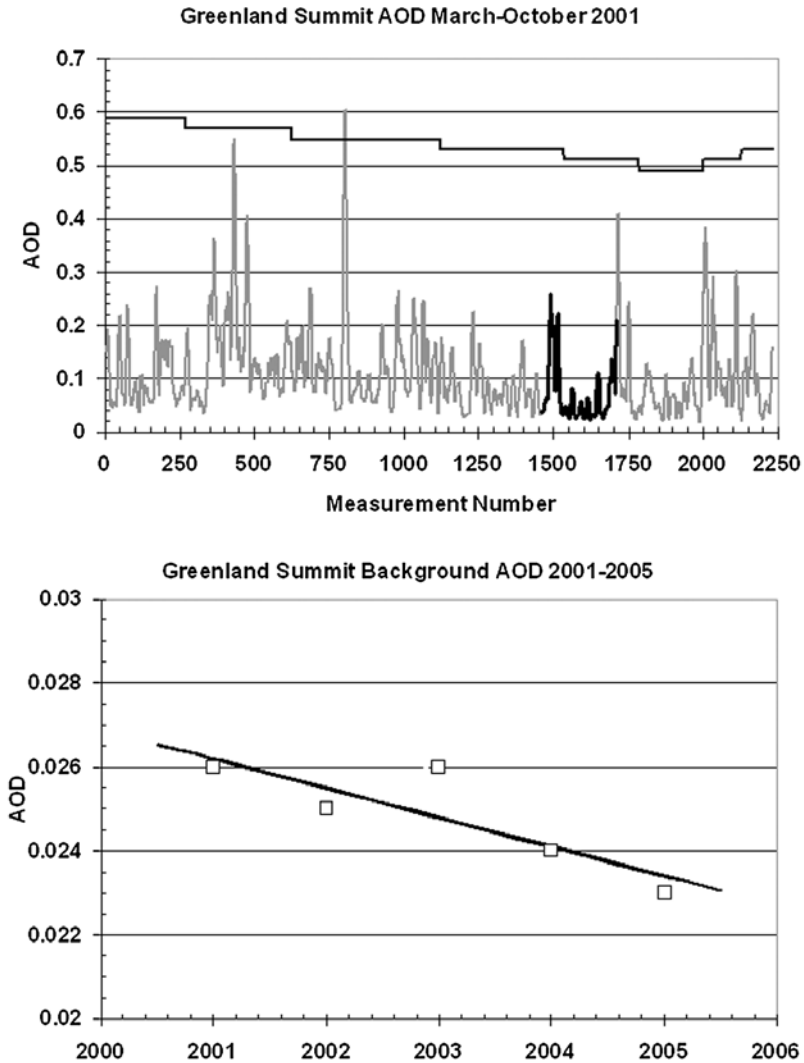

Figure 3. Aerosol optical depth (at the wavelength of 500 nm) measurements at the Summit of the Greenland ice sheet. (top) The AOD from March to October 2001. The thick black part of the curve (20 June to 20 July) is a section of the year that we use to estimate the background aerosol optical depth. The upper broken line indicates months from March till October. (bottom) Decreasing background aerosol optical depth during the 2001–2005 period defined as an average of observations within 20 June to 20 July with AOD < 0.03.

2005; *Streets et al.*, 2006; *Wild et al.*, 2007] and from the surface background AOD measurements. In situ pyrheliometer measurements of background AOD at Point Barrow, Alaska, and Sun photometer measurements from Ny Alesund, Spitsbergen [*Tomasi et al.*, 2007] suggest a decreasing background AOD at the rate of 1.5%/a at Point Barrow (data available for 1977–2002 time period) and at the rate of 2.2%/a at Ny Alesund (data available for 1991–2006). Similarly, the AOD of background aerosols at the Summit of the Greenland ice sheet (Figure 3) suggests a decreasing trend of the background aerosol optical depth. Several independent measurements suggest a decreasing AOD during at least the past decade. Consequently the decreasing AOD is a factor that should be included in the evaluation of the anthropogenic influences on climate. For our further consideration, on the basis of the AVHRR and MISR satellite records, we take

$$\Delta\tau/\Delta t = -0.0014/\mathrm{a} \tag{10}$$

as an approximate rate of decrease of global aerosol optical depth during the past decade.

[16] While it is possible that some of the downward trend in AOD is from climate feedbacks (increase of precipitation or reduction of winds), the fact that the strongest decrease is observed over North America and western Europe suggests that the trend is dominated by a

Table 1. List of Parameters Used in Estimate of Empirical Effective Climate Sensitivity

Parameter	Rate of Change	Source
Carbon dioxide rate of change	1.9 ppmv/a	*IPCC* [2007]
Aerosol optical depth rate of change	0.0014/a	AVHRR [*Mishchenko et al.*, 2007], MISR (this paper), ground measurements (this paper)
Surface temperature rate of change	0.018/a	*IPCC* [2007]
Efficiency of ocean heat uptake, κ	1.4, 0.90, 0.70, 0.50 Wm^{-2}/K	*Raper et al.* [2002]

reduction in anthropogenic emissions which can be considered an external climate forcing.

5. Aerosol Radiative Forcing

[17] The effect of atmospheric aerosols on climate is usually divided into direct and indirect aerosol effect. The direct effect [*Chylek and Coakley*, 1974; *Schulz et al.*, 2006] consists of absorption and scattering of radiation by aerosol particles. The indirect effect [*Lohmann and Lesins*, 2002; *Lohmann and Feichter*, 2005] includes aerosol modification of cloud radiative properties (cloud albedo effect) and cloud life cycle (cloud lifetime effect).

[18] Nonabsorbing aerosols always produce a cooling whereas absorbing aerosols can cause either cooling or warming, depending on aerosol properties (refractive index, size and shape) and on the properties of the underlying surface [*Chylek and Coakley*, 1974; *Chylek and Wong*, 1995]. The radiative forcing due to the direct aerosol effect can be estimated using the global average aerosol optical depth, single scattering albedo and an asymmetry parameter. The rate of aerosol optical depth decrease can be translated into a rate of the top-of-atmosphere radiative forcing using an approximation valid for a thin aerosol layer [*Chylek and Wong*, 1995]

$$\frac{\Delta F}{\Delta t} = -\frac{S_0}{4} T^2 (1-N) \frac{\Delta \tau}{\Delta t} \left[(1-a)^2 (1-g)\omega - 4a(1-\omega) \right] \tag{11}$$

where S_o is the solar constant, T is the average atmospheric transmission for solar radiation, N is the average cloudiness, a is the surface albedo, g is the asymmetry parameter of the aerosol size distribution and ω is the average aerosol single scattering albedo. We calculate the rate of radiative forcing change separately for oceans and for land. Taking typical values for S_o = 1368 W/m^2, T = 0.8, ω = 0.98, g = 0.75 [*d'Almeida et al.*, 1991], N = 0.77 over oceans and N = 0.52 over land, a = 0.06 over oceans and a = 0.22 over land [*Penner et al.*, 1992], and accepting the above deduced value $\Delta\tau/\Delta t = -0.0014$/a for both ocean and land we obtain the rate of a direct global aerosol radiative forcing

$$\left(\frac{\Delta F}{\Delta t}\right)_{DIRECT} = 0.018 \ \mathrm{Wm^{-2}/yr} \tag{12}$$

The aerosol indirect effect (sum of cloud albedo and cloud lifetime effect) is not yet well understood, but it has been estimated to be approximately as large as the direct aerosol forcing and possibly up to twice the direct aerosol forcing [*Rotstayn and Penner*, 2001; *IPCC*, 2007]. For our estimate we assume that the indirect effect is approximately equal to the direct effect. Thus we have

$$\left(\frac{\Delta F}{\Delta t}\right)_{INDIRECT} = 0.018 \ \mathrm{Wm^{-2}/yr} \tag{13}$$

With the current trend of decreasing aerosol optical depth the aerosol direct and indirect forcing is positive and thus contributes to the global climate warming.

[19] We obtain the combined effect of the dominating forcing agents assuming that the forcings and responses are additive. This may not be exactly correct [*Hansen et al.*, 2006; *Lohmann and Feichter*, 2005] but is a necessary simplification when trying to empirically determine the climate sensitivity. Thus the combined contribution of carbon dioxide and aerosols leads to the rate of the top-of-atmosphere radiative forcing:

$$\left(\frac{\Delta F}{\Delta t}\right)_{DIRECT+INDIRECT+CO_2} = 0.063 \ \mathrm{Wm^{-2}/yr} \tag{14}$$

6. Empirical Climate Sensitivity

[20] Using equations (14), (9), and (5), the range of the ocean uptake efficiency, from Table 2, and both the radiative forcing of increasing concentration of CO_2 and reduced AOD, the empirical climate sensitivity is

$$\lambda = 0.29 \text{ to } 0.48 \pm 0.12 \ \mathrm{K/Wm^{-2}} \tag{15}$$

where ± 0.12 K/Wm^{-2} represents our estimate of uncertainty due to approximations used and due to uncertainty in selected parameters (Table 1). If the aerosol indirect effect is neglected the climate sensitivity increases to 0.56 K/Wm^{-2}. Table 2 provides a summary of empirical climate sensitivity obtained under different assumptions concerning the radiative forcing and ocean heat flux. The deduced climate sensitivity is lower than suggestions of most climate models, however, it is in the same range as the climate sensitivity obtained with cloud resolving models embedded in global climate models [*Miura et al.*, 2005; *Wyant et al.*, 2006], and it is considerably higher than the climate sensitivity derived from the ocean heat capacity and climate time lag response [*Schwartz*, 2007].

7. Summary and Discussion

[21] We have deduced empirical climate sensitivity using the currently observed rates of change of global surface temperature, carbon dioxide atmospheric concentration and

Table 2. Empirical Climate Sensitivity, λ, Obtained Under Different Assumptions of Radiative Forcing, ΔF, and Ocean Heat Flux, $\Delta H = \kappa \Delta T$

	Carbon Dioxide Only	CO_2 + Direct Aerosol Effect	CO_2 + Direct + Indirect Aerosol Effect
ΔF per decade	+0.27 Wm^{-2}	+0.45 Wm^{-2}	+0.63 Wm^{-2}
κ = 1.40 Wm^{-2}/K	λ = 10 K/Wm^{-2}	λ = 0.91 K/Wm^{-2}	λ = 0.48 ± 0.12 K/Wm^{-2}
κ = 0.90 Wm^{-2}/K	λ = 1.67 K/Wm^{-2}	λ = 0.62 K/Wm^{-2}	λ = 0.38 ± 0.12 K/Wm^{-2}
κ = 0.70 Wm^{-2}/K	λ = 1.25 K/Wm^{-2}	λ = 0.56 K/Wm^{-2}	λ = 0.36 ± 0.12 K/Wm^{-2}
κ = 0.50 Wm^{-2}/K	λ = 1.00 K/Wm^{-2}	λ = 0.50 K/Wm^{-2}	λ = 0.33 ± 0.12 K/Wm^{-2}
ΔH = 0	λ = 0.67 K/Wm^{-2}	λ = 0.40 K/Wm^{-2}	λ = 0.29 ± 0.12 K/Wm^{-2}

aerosol optical depth. We have allowed for slow feedbacks by using model-based values for ocean heat uptake. Considering a hypothetical case that all the current radiative forcing is due to increasing carbon dioxide (at the rate of 1.9 ppmv/a) the estimated empirical climate sensitivity is between 0.67 and 10 K/Wm^{-2} (Table 2) depending on the heat flux into the ocean. Considering satellite and ground observations of aerosol optical depth changes, we have used a decreasing AOD rate of 0.0014/a. When both the carbon dioxide and aerosol forcing (direct and indirect) are included the deduced climate sensitivity is reduced to a value between 0.29 and 0.48 K/Wm^{-2} depending on the ocean heat uptake. Thus neglecting the radiative forcing due to decreasing aerosol optical depth leads to an overestimate of the empirical climate sensitivity by at least a factor of 2. It will be important to include currently decreasing aerosol optical depth in future three-dimensional climate simulations to explore the region-specific climate effects of AOD trends.

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