

Radiative forcing and the ice core greenhouse gas record

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Many books and articles deal with the numerous feedbacks and links in the climate-biogeochemical system and their consequences for our society. Here, the focus is on one particular link between the physical climate system and biogeochemical cycles, that is the radiative forcing imposed on the physical climate system through altered greenhouse gas concentrations and other radiative perturbations.

Polar ice cores analyzed by different groups around the world leave no doubt that cold polar ice faithfully records past atmospheric concentrations of major greenhouse gases (Figure 1) [*Stauffer et al.*, 2002]. The well-known finding is that the rate of increase in atmospheric carbon dioxide and methane is without precedence at least over the past few tens of millennia and that CO₂ and CH₄ concentrations today are larger than at any time over the past 650,000 years, the period spanned by the ice core records. Atmospheric concentration of N₂O, another greenhouse gas, is also higher than ever measured, though its ice core record is less complete. Humans, by burning fossil fuels and through land use changes and other activities, are altering the state of the atmosphere.

Increasing concentrations of greenhouse gases cause global warming. Greenhouse gases such as water and CO₂ absorb part of the long-wave radiation emitted from the earth surface, thereby altering the temperature on the ground and the temperature distribution in the atmosphere. Without greenhouse gases the temperature on the earth surface would be well below the freezing point of water. The greenhouse gas theory was already well established in the 19th century when Arrhenius calculated that a doubling of the atmospheric CO₂ concentration would cause a global-mean surface warming of about 4°C. In comparison, this climate sensitivity, termed ΔT_{2x} , evaluated with the current set of comprehensive and spatially-resolved climate models falls generally within the range from 1.5 to 4.5°C.

Climate scientists use a concept termed ‘radiative forcing’ to compare the climatic influence of a variety of greenhouse gases as well as that of other radiative agents such as aerosols or other externally imposed perturbations on the radiative energy budget of the planet [*Ramaswamy et al.*, 2001]. The basic concept of radiative forcing is relatively simple: It is the change in the radiation flux entering the lower atmosphere-surface system through the tropopause immediately after a perturbation (e.g. after a step-like increase in atmospheric CO₂).

The definition of radiative forcing is not without complexity and slightly different versions are found in the literature. IPCC uses the following “The radiative forcing of the surface-troposphere system due to the perturbation in or the introduction of an agent (say, a change in greenhouse gas concentrations) is the change in net (down minus up) irradiance (solar plus long-wave; in Wm⁻²) at the tropopause after allowing for stratospheric temperatures to readjust to radiative equilibrium, but with surface and tropo-

spheric temperatures and state held fixed at the unperturbed values". Radiative forcing is typically on the order of a few Watts per square meter (W m^{-2}). For example, a doubling of the preindustrial CO_2 concentration from 280 ppm to 560 ppm causes a radiative forcing, $RF(2x\text{CO}_2)$, of 3.7 W m^{-2} .

Climate feedbacks such as the change in water vapor and cloud cover in response to climate change are not included in the definition of radiative forcing. The strength of these climate feedbacks defines the climate sensitivity. The change in global mean surface temperature, $\Delta T_{s,\infty}$, and radiative forcing, RF , are linked through the equilibrium climate sensitivity, ΔT_{2x} . When the climate system has reached a new (quasi-) equilibrium, the change in surface temperature equals the product of the climate sensitivity and the change in radiative forcing divided by $RF(2x\text{CO}_2)$:

$$(1) \quad \Delta T_{s,\infty} = \Delta T_{2x} \frac{RF}{RF(2x\text{CO}_2)}$$

Calculations of radiative forcing are done in a 3-dimensional setting taking into account the non-uniform distribution of temperature, pressure and of other variables in the atmosphere as well as the absorption spectrum of individual molecules. The radiative forcing fields vary over space and time. It has been shown that the globally and annually averaged radiative forcing provides a good measure for the resulting surface temperature changes even for forcings with very different spatial distributions. Table 1 provides a set of equations that allows the reader to directly estimate the radiative forcing relative to the preindustrial period or any other reference for the most common greenhouse gases. These formulations summarize the results of comprehensive radiation models [Ramaswamy *et al.*, 2001]. CO_2 forcing increases logarithmically with concentration as CO_2 is relatively abundant and self-shading effects occur. Radiative forcing by CH_4 and N_2O increases with the square root of the concentration. This means that an identical concentration increase is more effective at lower concentrations than at higher concentrations for these three gases. On the other hand, radiative forcing increases linearly with concentration for gases with a low abundance such as halocarbons and SF_6 . These gases have been added to the atmosphere by humans just over the recent decades.

Radiative forcing can be calculated over the paleorecord from ice core concentrations. Figure 1 details the evolution of CO_2 and CH_4 and their radiative forcing over the last deglaciation and the current warm period, the Holocene. Atmospheric CO_2 increased from about 190 ppm at the Last Glacial Maximum to about 265 ppm at the beginning of the Holocene, the current warm period. Variations during the Holocene were small. Similarly, CH_4 increased by about 350 ppb over the transition and underwent modest changes during the Holocene. Slow natural climate and greenhouse gas variations over the past glacial-interglacial cycles are thought to be largely driven by changes in the earth's orbit around the sun. CO_2 and CH_4 concentrations have remained within relatively narrow limits during glacial/interglacial cycles. Their concentrations have stayed below 300 ppm and 800 ppb, respectively. In contrast, over less than two centuries atmospheric CO_2 increased by almost 100 ppm and CH_4 by 1000 ppb. The radiative forcing (relative to

1750 A.D.) by CO₂ is -2 W m⁻² at the Last Glacial Maximum and is +1.6 W m⁻² today. The forcing from the CH₄ changes is considerably smaller, reaching about half a W m⁻² today.

The warming commitment from today's radiative forcing by CO₂ and CH₄ of 2.1 W m⁻² is readily evaluated from equation (1). For a mid-range climate sensitivity of 3°C one expects an equilibrium warming of 1.7°C relative to preindustrial. However, a considerable part of the greenhouse gas forcing is offset by the cooling influence of aerosols and the climate system lags the forcing due to the large thermal inertia of the ocean.

Figure 2 puts the radiative forcing from greenhouse gases in the context of the known perturbations for the Last Glacial Maximum and for today. The increase in albedo due to larger ice sheets is the largest individual forcing at the last glacial maximum, whereas changes associated with the biogeochemical system (greenhouse gases, dust, vegetation changes) are responsible for more than half of the total Last Glacial Maximum forcing. The change in the total energy received by the earth from the sun through an altered orbital configuration was relatively small, however the energy was distributed differently over seasons and latitudes compared to today. Today's total radiative forcing is the result of partly offsetting contributions from the well-mixed greenhouse gases (CO₂, CH₄, N₂O, CFCs), stratospheric and tropospheric ozone, a mix of different aerosols, and changes in land use and solar irradiance. CO₂ is clearly the most important agent among the well-mixed greenhouse gases. Over the coming decades, atmospheric CO₂ will become of even greater relative importance, as emitted CO₂ accumulates over time in the climate system and recent sulfur emission control measures limit the growth in (cooling) aerosol forcing. In contrast to aerosols, a perturbation in atmospheric CO₂ is removed only very slowly on time scales ranging from decades to many millennia.

The rate of change in radiative forcing is dramatically accelerating. Radiative forcing by CO₂ changed on average by less than 0.1 milli-Watt m⁻² yr⁻¹ over the 6000 years preceeding the industrial era, already by about 12 milli-Watt m⁻² yr⁻¹ over the 20th century and by as much as 27 milli-Watt m⁻² yr⁻¹ over the most recent decade (1994 to 2003).

Figures and Tables

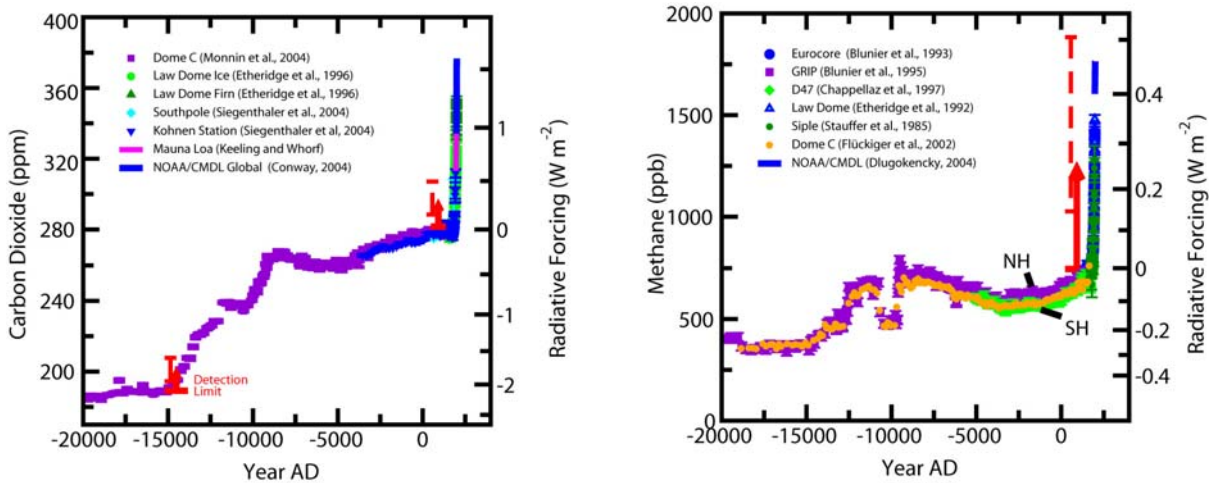


Figure 1: The evolution of atmospheric CO₂ (left) and CH₄ (right) over the last 20,000 years. The right-hand axes show the radiative forcing relative to 1750 A.D. associated with the concentration changes as calculated with the formulations in Table 1. Note that the radiative forcing scale is non-linear. The red arrows indicate the theoretical detection limit of a forcing (relative to last glacial maximum and preindustrial conditions) within a climate system model, assuming an internal climate variability or detection threshold of 0.2°C, climate equilibrium, and a mid-range climate sensitivity of 3°C. The dashed red arrows indicate the uncertainty of the detection threshold associated with the climate sensitivity range of 1.5 to 4.5°C.

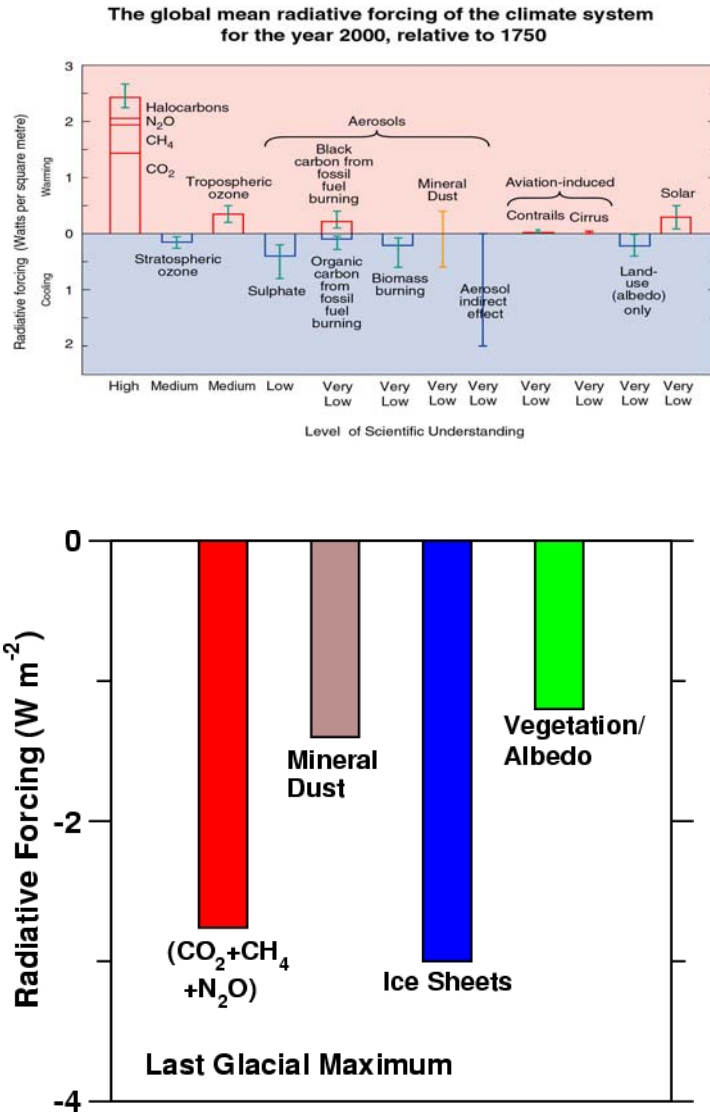


Figure 2: Radiative forcing for today [Ramaswamy *et al.*, 2001](top) and the Last Glacial Maximum (bottom) relative to 1750 A.D. for the range of known perturbations.

Table 1: Equations to calculate radiative forcing relative to a preindustrial (1750 A.D.) reference concentration (C_0). The overlap in absorption bands between N_2O and CH_4 is taken into account using the overlap function $f(M,N)=0.47 \ln(1+2.01 \times 10^{-5} (MN)^{0.75} + 5.31 \times 10^{-15} M(MN)^{1.52})$. For many practical purposes, this correction term can be neglected. Formulations for additional greenhouse gases can be found in [Joos *et al.*, 2001; Ramaswamy *et al.*, 2001]

agent	equation	C_0
CO_2	$RF = 5.35 \text{ W m}^{-2} \ln(CO_2/CO_{2,o})$	278 ppm
CH_4	$RF = 0.036 \text{ W m}^{-2} (\sqrt{CH_4} - \sqrt{CH_{4,0}}) - (f[CH_4, N_2O] - f[CH_{4,0}, N_2O])$	742 ppb
N_2O	$RF = 0.12 \text{ W m}^{-2} (\sqrt{N_2O} - \sqrt{N_{2O,0}}) - (f[CH_{4,o}, N_2O] - f[CH_{4,0}, N_{2O,0}])$	272 ppb
CFC-11	$RF = 0.25 \text{ W m}^{-2} (CFC-11 - CFC-11_0)$	0 ppt
CFC-12	$RF = 0.32 \text{ W m}^{-2} (CFC-12 - CFC-12_0)$	0 ppt

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