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DLR-Institut für Physik der Atmosphäre, Oberpfaffenhofen, Germany

Stratosphere adjusted radiative forcing calculations in a comprehensive climate model

N. Stuber, R. Sausen, and **M. Ponater**

With 3 Figures

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Summary

A new technical procedure is introduced to determine the stratosphere adjusted radiative forcing at the tropopause in the framework of the 3-D climate model ECHAM4. However, the procedure appears to be appropriate for other GCMs as well. It allows to study in a straightforward way the problem of the general usefulness of radiative forcing as a reliable predictor of climate change. Some examples are given for illustration. It is, once again, confirmed that the climate sensitivity is practically equal for experiments with increased solar insolation and increased CO_2 concentration. However, a higher climate sensitivity is obtained for ozone perturbations with a horizontally or vertically inhomogeneous distribution. The latter finding is in qualitative agreement with respective results reported in other studies, whereas the value of the climate sensitivity is exceptionally high in our model. The physical reasons for the unique model behaviour in the ozone experiments are currently not understood.

1. Introduction

Radiative Forcing (RF) has been established as a reliable predictor of climate change (IPCC 1990, 1992, 1995, 1999) and it will also be used in the forthcoming IPCC Third Assessment Report. Its high preference in climate research is due to the fact that it gives a reasonable first order approximation of global climate change without the need for time-consuming and (computationally) expensive 3-D simulations of the climate system. The concept is based on climate sensitivity experiments mainly with respect to changes of well-mixed greenhouse gases and solar insolation (e.g., Manabe and Wetherald, 1980; Hansen et al., 1984; Cess et al., 1985). It assumes an approximately linear relationship between the global mean radiative forcing and the equilibrium response of global mean surface temperature ΔT_{surf}. The climate sensitivity parameter λ relates the climate forcing to the climate response:

$$\Delta T_{\mathrm{surf}} = \lambda \cdot \mathrm{RF} \tag{1}$$

If the value of λ were independent from the strength of the radiative perturbation (RF), the horizontal and vertical distribution of the perturbation, and the spectral characteristics (solar and terrestrial partition) of the absorbers, RF would indeed be an ideal metric of climate change.

While the climate sensitivity has often been found to be constant in a given model framework, it is also known to be a highly model-dependent parameter. Cess et al. (1990) pointed out that this is mainly due to different acting of feedbacks in the various models, particularly due to the cloud feedback. Without the cloud feedback most models tend to produce similar responses to the same forcing.

There are several definitions of radiative forcing. All have in common that they quantify the global mean radiation imbalance following a perturbation before the atmosphere returns to a

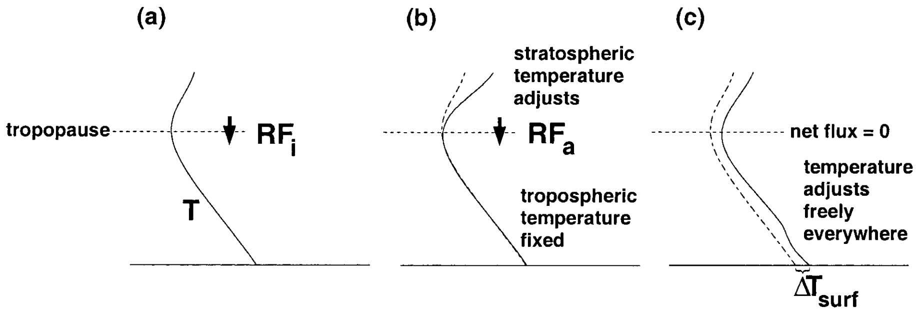

Fig. 1. Schematic representation of the different definitions of radiative forcing at the tropopause and the associated temperature change: (**a**) instantaneous radiative forcing (RF_i), (**b**) stratosphere adjusted forcing (RF_a), and (**c**) the full equilibrium response ($\Delta T, \Delta T_{\rm surf}$) including all feedbacks. Redrawn after Hansen et al. (1997)

new full equilibrium (IPCC, 1995; Hansen et al., 1997). The imbalance is determined either at the tropopause level or at the top of the atmosphere. We concentrate on those definitions that consider the radiative imbalance at the tropopause, which proved to be the best link to climate change. The definitions of radiative forcing differ as to which degree the atmosphere is allowed to react to the prescribed perturbation. The instantaneous radiative forcing (Fig. 1a) is calculated without further changes in any atmospheric variable. In contrast, the stratosphere adjusted radiative forcing (Fig. 1b) is calculated at the tropopause after allowing stratospheric temperatures to adjust to a new radiative equilibrium, without changes in tropospheric variables and stratospheric dynamics (fixed dynamic heating – FDH – concept). In the following we use this latter definition of RF, unless explicitly mentioned differently. This has been shown to be the most appropriate way for ensuring that the empirical relation (1) is valid with constant λ. It has to be stressed, that the bottom line of equation (1) is to relate a purely radiative response of the stratosphere (that will never occur in reality), from which the radiative forcing is derived, with a fully dynamic equilibrium change (Fig. 1c), from which the surface temperature response is derived.

RF was first calculated from radiative–convective models (Ramanathan, 1976; Ramanathan and Dickinson, 1979). In the course of time the concept has been continuously extended to more sophisticated models. Determining the stratosphere adjusted radiative forcing from 3-D models is still a comparatively new endeavor. As RF was originally a concept for 1-D models, its application to 3-D models introduces new, substantial complexity. Hitherto it has been common to calculate stratosphere adjusted RF forcing values off-line the framework of the 3-D model that is used to simulate the forcing perturbation (e.g., Chen and Ramaswamy, 1996; Haywood et al., 1998; Roelofs et al., 1998). This is not a fully satisfactory method, as it requires the calculation of an average perturbation to provide a suitable input for the offline radiative transfer model. Often, a set of monthly mean perturbations is created from the 3-D model results. However, in case of a perturbation highly variable in time substantial differences may result depending on whether the actual perturbation or the averaged perturbation is used (e.g., Feichter et al., 1997).

Recently, some doubt has been cast on the applicability of the radiative forcing concept on inhomogeneously distributed perturbations, especially ozone perturbations (Hansen et al., 1997; Forster et al., 1997; Christiansen, 1999). For example, Hansen et al. (1997) carried out a series of experiments with a simplified general circulation model (GCM). Adding a constant amount of ozone to each model layer individually, they showed that the climate response is far from being proportional to the radiative forcing, a fact they related to different cloud feedbacks. A maximum sensitivity occured for lower troposphere ozone perturbations. A significantly enhanced climate sensitivity in ozone change experiments was also found by Christiansen (1999), but in his GCM from a perturbation located in the higher stratosphere (above 35 hPa). In yet another GCM study Ponater et al. (1999) discovered enhanced climate sensitivity values for ozone perturbations related to aircraft emissions. However, they discussed only results derived from the instantaneous radiative forcing, which is considered a less adequate climate change predictor. Moreover, the ozone perturbations they applied were rather small, making a quantitative interpretation of their findings somewhat problematic.

In this paper, we will also (like Ponater et al., 1999) use the state of the art climate model ECHAM4 to investigate the characteristics of the GCM response to inhomogeneously distributed ozone perturbations. We will first point out a method of determining the stratosphere adjusted tropopause radiative forcing in the GCM framework (Sections 2 and 3). Section 4 gives an overview over the design of the experiments, and the respective results are presented in Section 5. The paper is closed with some conclusions in Section 6.

2. The ECHAM4/MLO model

ECHAM4 is a spectral atmosphere general circulation model, based on the primitive equations (Roeckner et al., 1996). Prognostic variables are vorticity, divergence, temperature, logarithm of surface pressure, and mixing ratios of water vapour and cloud water. Radiation, cloud formation, precipitation, convection, and vertical and horizontal diffusion are parameterized. Apart from water vapour, all radiatively active trace gases are prescribed. ECHAM4 has been used for a variety of climate sensitivity and change simulations (e.g., Lohmann and Feichter, 1997; Roelofs et al., 1997; Roeckner et al., 1999; Bengtsson et al., 1999; Ponater et al., 1999).

The radiation scheme of the ECHAM4 model has been adopted from the ECMWF forecast model (Fouquart and Bonnel, 1980; Morcrette et al., 1986). While the original scheme only accounts for H_2O, CO_2, and, in a simplified way, for O_3, the radiation scheme of ECHAM4 has been modified to include a number of additional greenhouse gases (methane, nitrous oxide and several chlorofluorocarbons; Roeckner et al., 1996). Furthermore, the inclusion of the 14.6 μm absorption band allows for a better representation of the radiative effects of stratospheric ozone (van Dorland et al., 1997). The parametrization of the water vapour continuum (Giorgetta and Wild, 1995) was also revised, and the treatment of cloud optical properties (Rockel et al., 1991; Roeckner, 1995) was upgraded. A detailed description of ECHAM4 and the simulated model climatology can be found in Roeckner et al. (1996). An evaluation of the simulated radiation balance has been given by Chen and Roeckner (1996) and by Wild et al. (1998). An extra check of the performance of the radiation scheme with respect to ozone perturbations revealed close agreement with a narrow band model both for the shortwave and the longwave response (P.M. de F. Forster, pers. comm.).

For studies involving climate sensitivity, ECHAM4 is available in a coupled configuration with a mixed layer ocean (MLO) module (Roeckner et al., 1995). The latter represents a mixed layer of 50 m depth. Only thermodynamics are considered, i.e., horizontal currents and vertical overturning are not explicitly represented, rather a prescribed heat flux (climatological annual cycle of 2-D longitude-latitude fields) accounts for horizontal and vertical heat exchange. A thermodynamic sea ice model is included, which determines sea ice depth and ice skin temperature from the energy balance at the top and the bottom of the ice layer.

The model was applied for the present study in T30 spectral horizontal resolution, corresponding to an isotropic resolution of 6°. Physical processes are calculated on the associated Gaussian latitude-longitude grid of approximately $3.75° \times 3.75°$ resolution. Vertically, 19 layers between the surface and the model top at 10 hPa are used.

Previous climate sensitivity and change studies in the ECHAM4 framework have either relied on the instantaneous RF at the tropopause and the associated climate sensitivity (Lohmann and Feichter, 1997; Roeckner et al., 1999; Roelofs, 1999; Ponater et al., 1999) or have made off-line calculations of the stratosphere adjusted tropopause RF (Roelofs et al., 1998), transferring the perturbation produced by the 3-D GCM to a 1-D column version of the ECHAM4 model that uses equivalent physics parametrizations.

3. Determining the stratosphere adjusted radiative forcing in a GCM

Calculating the stratosphere adjusted tropopause radiative forcing is based on the fixed dynamic heating (FDH) concept (Ramanathan and Dickinson, 1979; Fels et al., 1980). The concept assumes that the stratospheric temperature quickly adjusts to a new quasi-stationary state in response to the radiative perturbation before the troposphere and surface response is providing a substantial feedback. Following this assumption, the perturbed equilibrium temperature T^* is calculated such that

$$\frac{dT^*}{dt} = \left.\frac{dT^*}{dt}\right|_{\mathrm{dyn}} + \left.\frac{dT^*}{dt}\right|_{\mathrm{rad}} (T^*, m'_{\mathrm{H_2O}}, m'_{\mathrm{O_3}}, m'_{\mathrm{CO_2}}, \ldots) = 0 \text{ (stratosphere)} \tag{2}$$

$$T^* = T \text{ (troposphere)} \tag{3}$$

where the temperature tendency is separated in a dynamic and a radiative part (T is the temperature of the unperturbed atmosphere). The tendency due to radiation depends on externally perturbed mixing ratios of atmospheric constituents ($m'_{\mathrm{H_2O}}, m'_{\mathrm{O_3}}, m'_{\mathrm{CO_2}}, \ldots$). Following the FDH concept, the dynamic heating rates for the unperturbed and the perturbed temperature are equal, i.e.,

$$\left.\frac{dT^*}{dt}\right|_{\mathrm{dyn}} = \left.\frac{dT}{dt}\right|_{\mathrm{dyn}}. \tag{4}$$

Obviously, assuming fixed dynamic heating is only an approximation of the full stratospheric response, and the discussion whether the stratosphere should be allowed a purely radiative or a fully dynamic equilibrium before calculating the "adjusted forcing" is going on. Currently the FDH approximation is widely accepted (e.g., Christiansen et al., 1997; Hansen et al., 1997; IPCC, 1999), and we also rely on it. Christiansen (1999) provides evidence that the adjusted RF value arising from a radiatively adjusted stratosphere is preferable compared to the RF value resulting from a full radiative and dynamic adjustment process in the stratosphere. He finds that the assumption of constant climate sensitivity is better fulfilled in the former case.

Fels et al. (1980) report that it may take several months for the lower stratosphere to reach a new equilibrium. During this time the climate, as well as the perturbation, may have considerably evolved. We, therefore, do not assume a stationary equilibrium state, but a quasi-stationary evolving state of the stratosphere. The adjusted stratospheric temperature T^* develops according to

$$\frac{dT^*}{dt} = \left.\frac{dT}{dt}\right|_{\mathrm{dyn}} + \left.\frac{dT^*}{dt}\right|_{\mathrm{rad}} \tag{5}$$

where the dynamic heating is calculated from

$$\left.\frac{dT}{dt}\right|_{\mathrm{dyn}} = \frac{dT}{dt} - \left.\frac{dT}{dt}\right|_{\mathrm{rad}} \tag{6}$$

This method is similar to the "seasonally evolving fixed dynamical heating" (SEFDH) adjustment introduced by Forster et al. (1997). However, unlike Forster et al. we do not determine the right hand side of Eq. (6) using climatological data, but calculate the tendency of the unperturbed temperature and its radiative heating rate on-line at each time-step of the full three-dimensional model. In this way RF can be determined under the naturally varying conditions of the simulated annual cycle, and the adjusted temperature field T^* depends both on time and space. The global mean radiative forcing value essential for the climate sensitivity (Eq. 1) is obtained by eventually averaging over space (latitude and longitude) and time (annual cycle).

It is convenient to think of the additional temperature field T^* as of forming a "second" atmosphere. The "second" atmosphere feels the additional radiative heating of the prescribed perturbation in the stratosphere, while its dynamic heating is identical to that of the unperturbed atmosphere. In the troposphere the two atmospheres are completely identical. Employing this second atmosphere in a GCM requires an additional prognostic or quasi-prognostic variable (T^*) and a second calculation of the radiative heating rates according to Eq. (2). This procedure should be applicable to most GCMs.

The stratosphere adjusted forcing at the tropopause depends on how the tropopause is determined. Rather than a fixed model level or pressure level, we use a "thermal" tropopause, defined as the lowest level where the lapse rate exceeds -2K/km, according to the WMO definition (WMO, 1992). In the coarse vertical model grid the tropopause is actually calculated according to Dameris et al. (1995). The radiative fluxes are determined at the exact position of the tropopause. To calculate the stratospheric temperature adjustment the model layer containing the tropopause is included in the adjustment domain. A specific problem arises if the tropopause location is changing between various model layers during the simulation, as approaching a new stratospheric temperature equilibrium requires the adjustment domain to remain sufficiently constant during the whole integration. Hence, we decided to proceed as follows: First, a one year simulation is performed to determine the instantaneous radiative forcing at the actual tropopause. This run is also used to calculate an annual mean tropopause height. In a second simulation (12 evaluated months after a 6 months spin-up), the stratosphere adjusted forcing is calculated at the level of the annual mean tropopause height, according to the method described above.

4. Design of climate experiments

In the following sections we will apply the method described in Section 3 to a number of radiative perturbations to illustrate the usefulness of this approach. We have chosen perturbations to make our results a relevant contribution with respect to the problem to which extent the stratosphere adjusted radiative forcing is a reliable predictor of climate change. In all cases the equilibrium surface temperature response ΔT_{surf} (Eq. 1) was calculated from a multiyear simulation with the ECHAM4/MLO model described in Section 2.

First, we have upgraded the RF calculations described in Ponater et al. (1999), determining the adjusted RF for a couple of different perturbations. Ponater et al. used aircraft-induced ozone perturbations from several data sets to simulate the respective climate effect with ECHAM4/MLO. We consider three of their experiments:

- AC-MOG: aircraft-induced ozone perturbation as simulated by the MOGUNTIA chemical transport model (Zimmermann, 1994; 1∗MOG according to the notation of Ponater et al., 1999)
- AC-ECH3: aircraft-induced ozone perturbation as simulated by Dameris et al., 1998 (ECH3-92 according to the notation of Ponater et al., 1999)
- Equiv. CO_2: homogeneous CO_2 concentration increase by 7.5 ppmv, chosen to give the same global mean adjusted RF as in AC-MOG

Second, we carried out for this study experiments with idealized perturbations of stratospheric ozone, carbon dioxide, and solar insolation:

- Strat. O_3: horizontally uniform increase of the ozone concentration in model layers 2 to 4 (between 20 and 90 hPa) by about 7.0 ppmv, corresponding to approximately 315 DU
- Solar: increase of solar irradiance by 0.445%
- CO_2: homogeneous CO_2 concentration increase by 76 ppmv

The perturbations in this second experiment series were chosen to give a stratosphere adjusted RF of approximately 1 Wm^{-2}. The altitude of the stratospheric ozone perturbation was chosen in accordance with observed stratospheric ozone depletion (Forster, 1999), but it is opposite in sign (i.e., an ozone increase). The large compensation of the longwave and the shortwave component of the forcing requires a perturbation about ten times larger than observed to produce a RF of 1 Wm^{-2} at the tropopause. However, the larger RF makes it easier to separate the climate change signal from the internal climate variability, i.e. it reduces the statistical uncertainty of the climate sensitivity parameter as given by Ponater et al. (1999). In addition to the perturbation runs listed above, which have all been extended over 30 model years after spin-up, we performed a control run (CTRL) with no perturbation. The control run simulates the climate of the 1990ies. All response values are calculated relative to CTRL, which also serves to provide a measure of internal (“natural”) variability, particularly the interannual standard deviation of T_{surf} to calculate confidence intervals for its mean response.

5. Results and discussion

5.1 Aircraft induced ozone perturbations

Table 1 summarizes the values of the globally and yearly averaged climate change parameters for all experiments. We first discuss the characteristic features for the aircraft related results.

Consistent with previous experiments the adjusted RF from CO_2 is less than its instantaneous counterpart by about 6%, due to the stratospheric cooling caused by enhanced radiation to space. (For details on the effects of an

Table 1. Instantaneous radiative forcing at the tropopause (RF_i), stratosphere adjusted tropopause radiative forcing (RF_a), change of equilibrium surface temperature (ΔT_{surf}), climate sensitivity parameter calculated using RF_i (λ_i), and climate sensitivity parameter based on RF_a (λ_a) from six different radiative perturbations (see text)

	RF_i Wm^{-2}	RF_a Wm^{-2}	ΔT_{surf} K (mean)	λ_i $K/(Wm^{-2})$ (mean)	λ_a $K/(Wm^{-2})$ (mean)	λ_a $K/(Wm^{-2})$ (95% interval)
Aircraft related perturbations						
AC-MOG	0.068	0.065	0.096	1.41	1.48	[1.25, 1.71]
Equiv. CO_2	0.069	0.065	0.061	0.88	0.94	[0.71, 1.17]
AC-ECH3	0.031	0.038	0.062	2.01	1.63	[1.24, 2.03]
Idealized perturbations						
CO_2	1.066	1.004	0.808	0.758	0.805	[0.787, 0.823]
Solar	0.992	1.000	0.818	0.825	0.818	[0.800, 0.836]
Strat. O_3	−1.158	1.010	1.470	−1.269	1.455	[1.437, 1.473]

increased CO_2 content of the atmosphere, see Manabe and Wetherald, 1975.)

The effect of the AC-MOG ozone perturbation simulated by the MOGUNTIA CTM (Ponater et al., 1999; their Fig. 1), which is mainly an increase of tropospheric ozone, is a greenhouse warming of the troposphere and a cooling of the lower stratosphere (e.g., Clough and Iacono, 1995). Due to the latter effect the stratosphere adjusted forcing is again less than the instantaneous radiative forcing. In contrast, for the AC-ECH3 ozone perturbation the adjusted RF is larger than the instantaneous one. This ozone perturbation includes an ozone increase both in the troposphere and in the lower stratosphere, which leads to a considerable warming in the tropical lower stratosphere caused by enhanced absorption of solar radiation. The difference between the instantaneous and the adjusted radiation imbalance reaches 23% in the case of the AC-ECH3 ozone perturbation due to the large effect of the lower stratosphere temperature increase on downward thermal radiation. Nevertheless, although the radiative forcing is increased (compared to the instantaneous RF value) and the climate sensitivity λ_a derived from the stratosphere adjusted tropopause RF is smaller, the climate sensitivity remains substantially larger than for the CO_2 case. Thus, for both ozone perturbations considered the adjusted RF fails to be a quantitative predictor of the surface temperature change in the conventional way stated in, e.g., IPCC 1995. The respective conclusions drawn by Ponater et al. (1999) are thus confirmed.

5.2 Idealized perturbations

To discuss the results of the idealized perturbation experiments, we present zonal mean RF profiles and zonal mean cross sections of the stratospheric temperature adjustment in addition to the global mean parameters (Table 1). All values are annual averages. Figure 2 gives an overview of the zonal mean longwave, shortwave, and net stratosphere adjusted radiative flux changes at the tropopause.

The CO_2 perturbation (Fig. 2a) is most effective in the longwave part of the spectrum. The small shortwave contribution is due to absorption in the near infrared. The relative maxima and minima in the longwave (and net) zonal mean RF profile reflect the distribution of temperature, water vapour, and clouds in the troposphere. As already pointed out, the temperature decrease of up to 2 K in the stratosphere (Fig. 3a) reduces the longwave forcing at the tropopause, and leads to a global mean adjusted RF which is smaller than the instantaneous RF (Table 1). Note that the climate sensitivity value of $0.805\,K/(Wm^{-2})$ in this experiment is numerically different from the climate sensitivity found for the aircraft related CO_2 experiment. However, in the latter experiment both the surface temperature response and the climate sensitivity derived from it are associated with a rather high level of statistical uncertainty. Thus, as indicated by the confidence intervals, the difference between the climate sensitivity values is statistically not significant.

The increase of solar insolation causes by definition a purely shortwave instantaneous forcing. The global mean instantaneous forcing of

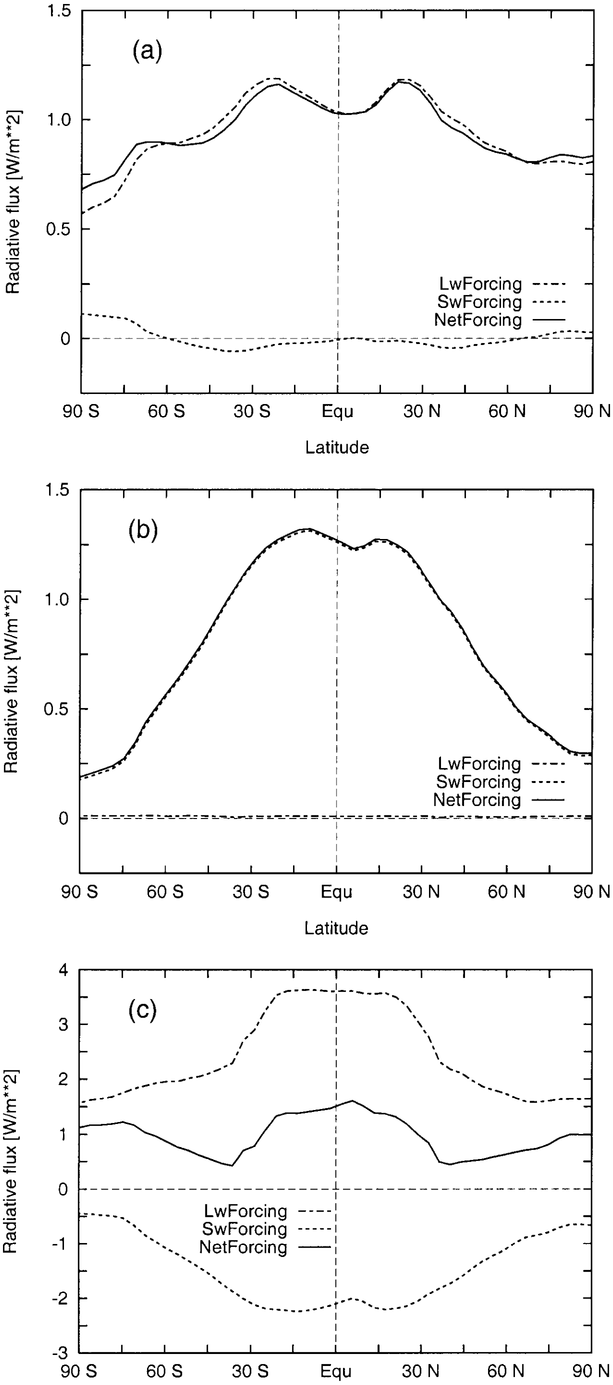

Fig. 2. Zonal and annual mean stratosphere adjusted radiative forcing at the tropopause for (**a**) the homogeneous CO_2 increase, (**b**) the increase of solar insolation, and (**c**) the upper stratospheric ozone increase

$0.992\,\mathrm{Wm^{-2}}$ is slightly increased to $1.000\,\mathrm{Wm^{-2}}$ after the stratosphere has been is allowed to adjust. Qualitatively, this adjustment means a warming as absorption of solar radiation by stratospheric ozone and water vapour increases. However, the magnitude of the stratospheric temperature adjustment is small and remains below 0.2 K everywhere. The relative minimum in forcing around 5° N (Fig. 2b) marks the annual and zonal mean position of the inner tropical convergence zone, where clouds reflect much solar radiation back to space. The meridional gradient of RF is considerably larger for the solar constant change experiment than for the CO_2 change experiment.

The stratospheric ozone perturbation causes a radiative forcing in the longwave as well as in the shortwave part of the spectrum. On the one hand the additional ozone absorbs shortwave solar radiation in the stratosphere, resulting in a negative shortwave forcing at the tropopause. On the other hand ozone is a greenhouse gas which implicates a positive forcing due to absorption and reemittance of longwave radiation. The overall effect is a negative instantaneous forcing, except for high polar latitudes (not shown). The warming due to the stratospheric adjustment (Fig. 3b), however, raises the global mean forcing from $-1.158\,\mathrm{Wm^{-2}}$ to $1.010\,\mathrm{Wm^{-2}}$ (Table 1). The warming is quite strong according to the large perturbation of about 7 ppmv that is required to induce an adjusted RF of $1\,\mathrm{Wm^{-2}}$. The enhanced downward longwave flux through the tropopause arising from the stratospheric temperature increase is far stronger than the direct shortwave forcing from the ozone increase, thus changing the sign of the net forcing. Incidentally, RF calculations by Fortuin et al. (1995) show this change in sign to be typical for ozone perturbations in the lower stratosphere, and a respective experiment by Christiansen (1999) yields the same result. However, this does no longer hold if the ozone change occurs at altitudes higher than about 15 km (Hansen et al., 1997; Christiansen, 1999).

Taking the statistical uncertainty of the monthly mean, globally averaged surface temperature in the simulations into account (0.05 K interannual standard deviation), the climate can be concluded to be equally sensitive to an increase in global atmospheric CO_2 content as to an increase in solar insolation. In this respect the assumption of a constant climate sensitivity for a given model framework is once more confirmed by our experiments with ECHAM4. However, for the stratospheric ozone perturbation the climate sensitivity parameter is larger by a factor of almost two.

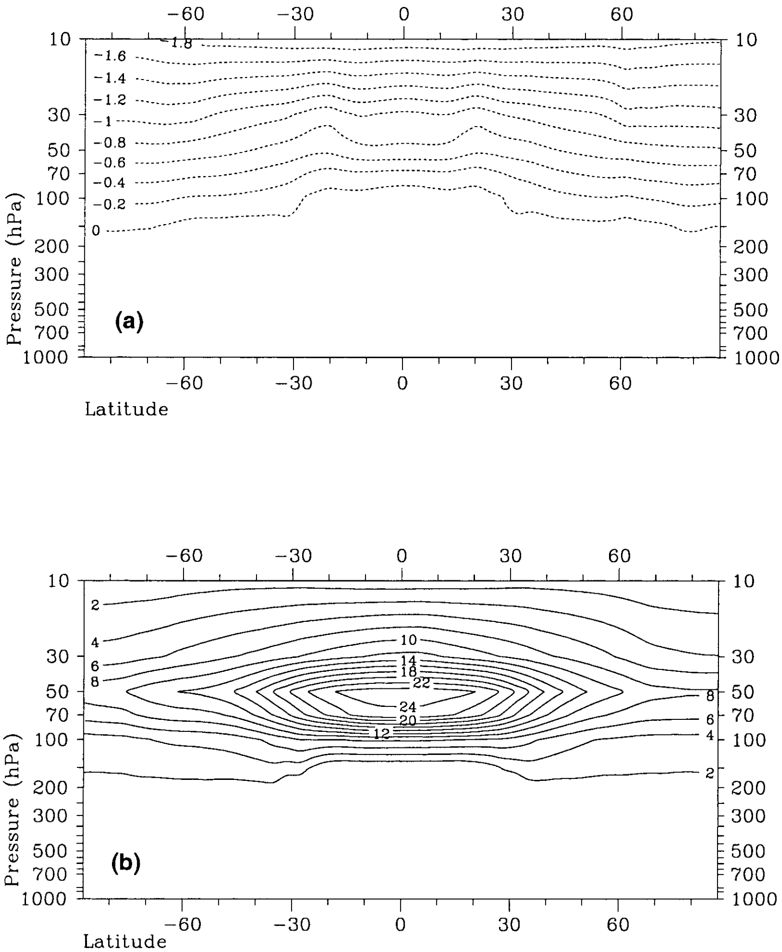

Fig. 3. Zonal and annual mean stratosphere adjusted temperature response (in K) for (**a**) the homogeneous CO_2 increase, and (**b**) the stratospheric ozone increase

While Christiansen (1999) and Forster and Shine (1999) also discovered anomalous λ values in ozone change experiments, this feature is much more distinct in our model.

6. Summary and conclusions

We have proposed a convenient procedure to determine the stratosphere adjusted radiative forcing in the framework of the ECHAM4 climate model. However, the procedure appears to be suited for other GCMs as well. The method is conceptually similar to the SEFDH adjustment technique used by Forster et al. (1997) in a column model.

The new procedure was applied in a number of experiments related to the question whether the stratosphere adjusted tropopause radiative forcing is a reliable predictor of climate change. First, it could be confirmed (as recently proposed by Ponater et al., 1999), that in ECHAM4 GCM experiments the climate sensitivity parameter is highly variable for different estimations of the aircraft induced ozone change distribution. In spite of the smallness of these perturbations, this result can not be explained by statistical uncertainties. The surface temperature appears to be significantly more sensitive to aircraft induced ozone perturbations than to an equivalent CO_2 perturbation.

In a second set of experiments we chose a number of physically different perturbations in such a way that each of them produces an annual mean stratosphere adjusted tropopause radiative forcing of 1 Wm^{-2}. Consistent with current theory the climate sensitivity to changes in atmospheric carbon dioxide content and solar insolation is remarkably constant (about 0.8 $K/(Wm^{-2})$ in our model), although the spectral properties of these perturbations are quite different. In contrast, the climate sensitivity for a stratospheric ozone perturbation is almost twice as large as that from a homogeneous greenhouse gas perturbation. Christiansen (1999), who identified an enhanced climate sensitivity for an upper stratospheric tenfold ozone increase, argued that the unique partitioning of the forcing in a shortwave and longwave part could be responsible for the effect. However, given the experience from our solar insolation and CO_2 experiments we do not think that this is a probable explanation in our case. Additionally it must be realized that the idealized ozone perturbation we have used is more similar to Christiansen's lower stratosphere ozone sensitivity experiment, for which he found no substantial deviation from the normal climate sensitivity at all. Large deviations between different models seem to be apparent in this respect, as it is known to be the case for the value of the climate sensitivity parameter in general. One has to be very cautious to generalize the response with respect to one individual ozone change profile to all kinds of stratospheric ozone perturbations. Hansen et al. (1997), e.g., pointed out that the altitude of the perturbation can be of utmost importance. For ozone perturbations in individual stratospheric model layers they got climate sensitivities ranging from 0.29 to $1.72\,K/(Wm^{-2})$, with the climate sensitivity of their simplified "Wonderland" climate model to CO_2 being $0.92\,K/(Wm^{-2})$.

Anyway, our results provide additional evidence that the stratosphere adjusted RF fails to reliably assess the response to vertically or horizontally inhomogeneous ozone perturbations. This evidence poses some challenge to the conclusion recently drawn by Forster et al. (2000), who stated that radiative forcing is more reliable to compare the climate impact of various perturbations than the temperature response simulated in GCMs. In our study the fluctuation of climate sensitivity due to the structure of the perturbation is of similar magnitude than the fluctuations one usually finds between the responses of different GCMs to the same perturbation. It is absolutely necessary to understand the physical reasons for the variation in temperature response and the climate sensitivity arising from the use of different forcings and models. We intend to conduct a series of systematic experiments with large ozone perturbations in various atmospheric regions. This will allow us an insight into the cause and effect mechanisms between climate forcing and response in a comprehensive climate model including all important feedbacks that may arise from a vertically or horizontally inhomogeneous perturbation.

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References

Bengtsson L, Roeckner E, Stendel M (1999) Why is the global warming proceeding much slower than expected? J Geophys Res 104: 3865–3876

Cess RD, Potter GL, Ghan SJ, Gates WL (1985) The climatic effects of large injections of atmospheric smoke and dust: a study of climate feedback mechanisms with one and three dimensional climate models. J Geophys Res 90: 12937–12950

Cess RD, Potter GL, Blanchet JP, Boer GJ, Del Genio AD, Déqué M, Dymnikov V, Galin V, Gates WL, Ghan SJ, Kiehl JT, Lacis AA, Le Treut H, Li Z-X, Liang X-Z, McAvaney BJ, Meleshko VP, Mitchell JFB, Morcrette J-J, Randall DA, Rikus L, Roeckner E, Royer JF, Schlese U, Sheinin DA, Slingo A, Solokov AP, Taylor KE, Washington WM, Wetherald RT, Yagai I, Zhangi M-H (1990) Intercomparison and interpretation of climate feedback processes in 19 atmospheric general circulation models. J Geophys Res 95: 16601–16615

Chen C-T, Ramaswamy V (1996) Sensitivity of simulated global climate to perturbations in low cloud microphysical properties. I. Global uniform perturbations. J Climate 9: 1385–1402

Chen C-T, Roeckner E (1996) Validation of the earth radiation budget as simulated by the Max Planck Institute for Meteorology general circulation model ECHAM4 using satellite observations of the Earth Radiation Budget Experiment (ERBE). J Geophys Res 101: 4269–4287

Christiansen B, Guldberg A, Hansen AW, Riishøjgaard LP (1997) On the response of a three-dimensional general circulation model to imposed changes in the ozone distribution. J Geophys Res 102: 13051–13077

Christiansen B (1999) Radiative forcing and climate sensitivity: The ozone experience. Quart J Roy Meteor Soc 125: 3011–3036

Clough SA, Iacono MJ (1995) Line-by-line calculation of atmospheric fluxes and cooling rates, 2. Application to carbon dioxide, ozone, methane, nitrous oxide and the halocarbons. J Geophys Res 100: 16519–16535

Dameris M, Nodorp D, Sausen R (1995) Correlation between Tropopause Height Pressure and TOMS-Data for the EASOE-winter 1991/1992. Beitr Phys Atmosph 68: 227–232

Dameris M, Grewe V, Köhler I, Sausen R, Brühl C, Grooss J-U, Steil B (1998) Impact of aircraft NO_x-emissions on tropospheric and stratospheric ozone. Part II: 3-D model results. Atmos Environ 32: 3185–3199

Feichter J, Lohmann U, Schult I (1997) The atmospheric sulfur cycle and its impact on the shortwave radiation. Climate Dyn 13: 235–246

Fels SB, Mahlman JD, Schwarzkopf MD, Sinclair RW (1980) Stratospheric sensitivity to perturbations in ozone and carbon dioxide: radiative and dynamical response. J Atmos Sci 37: 2265–2297

Forster PM de F, Freckleton RS, Shine KP (1997) On aspects of the concept of radiative forcing. Climate Dyn 13: 547–560

Forster PM de F (1999) Radiative forcing due to stratospheric ozone changes 1979–1997, using updated trend estimates. J Geophys Res 104: 24395–24400

Forster PM de F, Shine KP (1999) Stratospheric water vapour changes as a possible contributor to observed stratospheric cooling. Geophys Res Lett 26: 3309–3312

Forster PM de F, Blackburn M, Glover R, Shine KP (2000) An examination of climate sensitivity for idealised climate change experiments in an intermediate general circulation model. Climate Dyn 16: 833–849

Fortuin JPF, van Dorland R, Wauben WMF, Kelder H (1995) Green-house effect of aircraft emissions as calculated by a radiative transfer model. Ann Geophysicae 13: 413–418

Fouquart Y, Bonnel B (1980) Computations of solar heating of the earth's atmosphere: a new parametrization. Beitr Phys Atmosph 53: 35–62

Giorgetta M, Wild M (1995) The water vapour continuum and its representation in ECHAM4. Max-Planck Institut für Meteorologie, Rep 162, ISSN 0937-1060, Hamburg, 38 pp

Hansen JE, Lacis A, Rind D, Russell G, Stone P, Fung I, Ruedy R, Lerner J (1984) Climate sensitivity: analysis of feedback mechanisms. Geophysical Monographs 29: 130–163

Hansen JE, Sato M, Ruedy R (1997) Radiative forcing and climate response. J Geophys Res 102: 6831–6684

Haywood JM, Schwarzkopf MD, Ramaswamy V (1998) Estimates of radiative forcing due to modeled increases in tropospheric ozone. J Geophys Res 103: 16999–17007

IPCC (Intergovernmental Panel on Climate Change) (1990) Houghton JT, Jenkins GJ, Ephraums JJ (eds) Climate Change: the IPCC Scientific Assessment. Cambridge University Press, Cambridge, UK, 365 pp

IPCC (1992) Houghton JT, Callander BA, Varney SK (eds) Climate Change 1992: The Supplementary Report to the IPCC Scientific Assessment. WMO/UNEP, Cambridge University Press, Cambridge, UK, 200 pp

IPCC (1995) Houghton JT, Meira Filho LG, Bruce J, Lee H, Callander BA, Haites E, Harris N, Maskell K (eds) Climate Change 1994: Radiative Forcing of Climate Change. Cambridge University Press, Cambridge, UK, 393 pp

IPCC (1999) Penner JE, Lister DH, Griggs DJ, Dokken DJ, McFarland M (eds) Aviation and the Global Atmosphere; A Special Report of IPCC Working Groups I and III. Cambridge University Press, Cambridge, UK, 373 pp

Lohmann U, Feichter J (1997) Impact of sulfate aerosols on albedo and lifetime of clouds: a sensitivity study with the ECHAM4 GCM. J Geophys Res 102: 13685–13700

Manabe S, Wetherald RT (1975) The effects of doubling the CO_2 concentration on the climate of a general circulation model. J Atmos Sci 32: 3–15

Manabe S, Wetherald RT (1980) On the distribution of climate change resulting from an increase of CO_2 content of the atmosphere. J Atmos Sci 37: 99–118

Morcrette J-J, Smith L, Fouquart Y (1986) Pressure and temperature dependance of the absorption in longwave radiation parametrizations. Beitr Phys Atmosph 59: 455–469

Ponater M, Sausen R, Feneberg B, Roeckner E (1999) Climate effect of ozone changes caused by present and future air traffic. Climate Dyn 15: 631–642

Ramanathan V (1976) Radiative transfer within the earth's troposphere and stratosphere. J Atmos Sci 33: 1330–1346

Ramanathan V, Dickinson RE (1979) The role of stratospheric ozone in the zonal and seasonal radiative energy balance of the Earth-troposphere system. J Atmos Sci 36: 1084–1104

Rockel B, Raschke E, Weyres B (1991) A parametrization of broad band radiative transfer properties of water, ice and mixed clouds. Beitr Phys Atmosph 64: 1–12

Roeckner E (1995) Parameterization of cloud optical properties in the ECHAM4 model. In: Proceedings of the WCRP Workshop on ''Cloud Microphysics Parameterizations in Global Atmospheric Circulation Models'', May 23–25, 1995, Kananaskis, Alberta, Canada WCRP-Report No 90, 105–116, WMO/TD-No 713

Roeckner E, Siebert T, Feichter J (1995) Climate response to anthropogenic sulfate forcing simulated with a general circulation model. In: Charlson RJ, Heintzenberg J (eds) Aerosol Forcing of Climate. Wiley and Sons, Chichester, 349–362

Roeckner E, Arpe K, Bengtsson L, Christoph M, Claussen M, Dümenil L, Esch M, Giorgetta M, Schlese U, Schulzweida U (1996) The atmospheric general circulation model ECHAM-4: model description and simulation of present-day climate. Max-Planck Institut für Meteorologie, Rep 218, ISSN 0937-1060, Hamburg, 90 pp

Roeckner E, Bengtsson L, Feichter J (1999) Transient climate change simulations with a coupled atmosphere-ocean GCM including the tropospheric sulfur cycle. J Climate 12: 3004–3032

Roelofs G-J, Lelieveld J, van Dorland R (1997) A three-dimensional chemistry/general circulation model simulation of anthropogenically derived ozone in the troposphere and its radiative climate forcing. J Geophys Res 102: 23389–23401

Roelofs G-J, Lelieveld J, Feichter J (1998) Model simulations of the changing distribution of ozone and its radiative forcing of climate. Max-Planck Institut für Meteorologie, Rep 283, ISSN 0937-1060, Hamburg, 24 pp

Roelofs G-J (1999) Radiative forcing by tropospheric ozone: impact of cloud representation. Geophys Res Lett 26: 467–470

van Dorland R, Dentener FJ, Lelieveld J (1997) Radiative forcing due to tropospheric ozone and sulfate aerosols. J Geophys Res 102: 28079–28100

Wild M, Ohmura A, Gilgen H, Roeckner E, Giorgetta M, Morcrette J-J (1998) The disposition of radiative energy in the global climate system: GCM-calculated versus observational estimates. Climate Dyn 14: 853–869

WMO (1992) International meteorological vocabulary. World Meteorological Organization, ISBN 92-63-02182-1, Geneva, 784 pp

Zimmermann P (1994) The impact of aircraft released NO_x to the tropospheric ozone budget: sensitivity studies with a 3-D global transport/photochemistry model. In: Schumann U, Wurzel D (eds) Impact of emissions from aircraft and spacecraft upon the atmosphere. DLR Mitteilung 94-06, ISSN 0939-298X, Oberpfaffenhofen and Köln, 16 pp

Authors' address: Nicola Stuber (e-mail: Nicola.Stuber@dlr.de), Robert Sausen, Michael Ponater, DLR, Institut für Physik der Atmosphäre, Postfach 1116, D-82230 Wessling, Germany.