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Total aerosol effect: radiative forcing or radiative flux perturbation?

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Abstract

Uncertainties in aerosol radiative forcings, especially those associated with clouds, contribute to a large extent to uncertainties in the total anthropogenic forcing. The interaction of aerosols with clouds and radiation introduces feedbacks which can affect the rate of rain formation. In former assessments of aerosol radiative forcings, these effects have not been quantified. Also, with global aerosol-climate models simulating interactively aerosols and cloud microphysical properties, a quantification of the aerosol forcings in the traditional way is difficult to properly define. Here we argue that fast feedbacks should be included because they act quickly compared with the time scale of global warming. We show that for different forcing agents (aerosols and greenhouse gases) the radiative forcings as traditionally defined agree rather well with estimates from a method, here referred to as radiative flux perturbations (*RFP*), that takes these fast feedbacks and interactions into account. Based on our results, we recommend *RFP* as a valid option to compare different forcing agents, and to compare the effects of particular forcing agents in different models.

1 Introduction

Aerosols affect climate directly by scattering and absorption of shortwave and thermal radiation (direct effect). The global-mean net direct effect at the top-of-the-atmosphere (TOA) is a cooling that partly offsets the warming due to greenhouse gases. It is estimated as $-0.5\,\mathrm{W\,m^{-2}}$ with a 5 to 95% confidence range of -0.1 to $-0.9\,\mathrm{W\,m^{-2}}$ (Forster et al., 2007). In addition, aerosols modify the radiation budget indirectly by acting as cloud condensation nuclei and ice nuclei. The cloud albedo enhancement (first indirect effect, cloud albedo effect or indirect aerosol forcing) of warm stratiform clouds refers to an increase in cloud droplet number concentration due to anthropogenic aerosols for a constant liquid water content (Twomey, 1977). These more numerous and smaller

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cloud droplets increase the total surface area and thus cloud albedo. The cloud albedo effect can be calculated as a forcing because of the assumption of a constant liquid water content. Ensemble-averaged global-mean model estimates of the cloud albedo effect have remained rather constant over time (see Fig. 1) and amount to roughly $-0.9\,\mathrm{W\,m^{-2}}$. The $-0.9\,\mathrm{W\,m^{-2}}$ estimate that is obtained from the average over all published estimates, treating each of them equal (one paper one vote) is slightly larger than the estimate of the cloud albedo effect in the fourth assessment report of the Intergovernmental Panel on Climate Change (IPCC) where a different weighting procedure was used. There the median value of the indirect aerosol forcing was estimated as $-0.7\,\mathrm{W\,m^{-2}}$ with a 5 to 95% range of -0.3 to $-1.8\,\mathrm{W\,m^{-2}}$ (Forster et al., 2007). The rather large uncertainty in both the direct and indirect (cloud albedo effect) forcing accounts for a large fraction of the uncertainty in the total anthropogenic forcing (Kiehl, 2007).

In addition to the cloud albedo effect, there are multiple other effects of aerosols on clouds such as the cloud lifetime effect, the semi-direct effect and aerosol effects on mixed-phase, convective and cirrus clouds (Lohmann and Feichter, 2005; Denman et al., 2007). However, these effects cannot be evaluated via the usual definition of radiative forcing as the instantaneous change in radiative flux caused when the forcing agent is imposed, because these effects do not act "instantaneously". Also, if aerosols and/or cloud droplet number concentrations are calculated interactively in the model, the calculation of the aerosol radiative forcing is not straightforward because aerosols will then also influence the precipitation formation and with that cause an additional change in cloud properties. Hence these effects are usually evaluated as a radiative flux perturbation (*RFP*) (Haywood et al., 2009). The *RFP* is calculated as the difference in the top-of-the-atmosphere radiation budget between a present-day simulation and a pre-industrial simulation, both using the same sea surface temperatures. *RFP* estimates thus include fast changes and interactions in the climate system that induce changes in the meteorology. This does not conform to the pure definition of an instantaneous radiative forcing (Forster et al., 2007), in which only one radiatively active agent is changed, while leaving tropospheric profiles of temperature and other variables constant.

The issue of how to define radiative forcings is not new. This metric aims to estimate the influence of a particular climate perturbation on equilibrium global-mean surface temperature change, hence allowing comparison of different perturbations without the need to actually conduct equilibrium climate-change simulations. The concept of radiative forcing has been gradually refined, due to limitations that were found with the original idea of instantaneous radiative forcing. For forcing agents that affect stratospheric temperature, such as CO_2 and ozone, the procedure recommended by IPCC is to allow stratospheric temperatures to adjust to the imposed forcing agent (a process that takes a few months), before calculating the "adjusted" forcing at the tropopause (Shine et al., 1995). For increases of CO_2, this adjustment cools the stratosphere, reducing the net downwards flux at the tropopause by order 10% (Hansen et al., 2005). However, for stratospheric ozone depletion, omission of the adjustment has more drastic effects, changing the sign of the forcing from negative to positive (Shine et al., 1995; Hansen et al., 2005). Thus for ozone in particular, the stratospheric adjustment is essential if the radiative forcing is to be of any use as a predictor of the induced change in global-mean surface temperature.

More recent studies have shown that using the adjusted radiative forcing, the change in surface temperature per unit forcing, or climate sensitivity, is not strictly the same for different perturbations. To account for this, one approach suggested by Joshi et al. (2003) and Hansen et al. (2005) is to obtain an efficacy (E) and to display it next to forcing estimates. E is defined as the ratio of the climate sensitivity parameter for a given forcing agent to the climate sensitivity parameter for CO_2. E can vary markedly for different forcing agents and for different models, depending on how the forcing projects onto the various feedback mechanisms; see Forster et al. (2007) for a review. In particular, their Fig. 2.19 shows that for "realistic" perturbations of forcing agents in GCMs, E generally lies in the range of 0.6 to 1.3. The outlying point with $E\sim 1.65$ in that figure was derived by normalising ΔT obtained by Rotstayn and Penner (2001) in response to the combined cloud-albedo and lifetime effects by the forcing they calculated for the cloud-albedo effect only. If instead ΔT is normalised by the RFP estimated by Rotstayn and Penner (2001) for the combined effects, $E=0.86$ is obtained, in good agreement with the value of 0.83 they obtained for the cloud-albedo effect when it was calculated as an instantaneous forcing. This shows that the adjusted forcing concept does not work especially well for simulations that include indirect effects beyond the cloud-albedo effect. Further, the linear forcing-response concept may break down for certain idealised perturbations, especially involving absorbing aerosols. Aerosols within a certain range of single scattering albedo can even have negative adjusted forcing but induce a global-mean warming, i.e. E can be negative (Forster et al., 2007).

In the last few years, several studies have investigated yet another method of calculating radiative forcing, mainly in the context of CO_2 (Gregory et al., 2004; Forster and Taylor, 2006; Gregory and Webb, 2008; Andrews and Forster, 2008). The method is to regress the top-of-atmosphere radiative flux (N) against the global-mean surface air temperature change (ΔT). The forcing is taken as the intercept of the regression line, i.e. as the value of N when $\Delta T=0$. An interesting aspect of this method is that the efficacy is included in the forcing estimate (Forster and Taylor, 2006). Another important outcome from this work is that "fast feedbacks", such as cloud changes that respond directly to the forcing of CO_2 rather than to ΔT, are now regarded as part of the forcing (Gregory and Webb, 2008; Andrews and Forster, 2008). The "feedbacks" are considered to be those that operate on longer time scales (those on which T changes), and can be expressed as functions of ΔT. These conclusions are similar to those that have arisen in aerosol modelling, where it also seems desirable to treat "fast feedbacks" as part of the forcing. We note that the regression method may be useful for the evaluation of aerosol forcings in atmospheric models, but it also requires a mixed-layer or full ocean model, which not all groups have access to. A modification of the *RFP* method, in which land-surface temperature is fixed in addition to sea-surface temperature, was used by Shine et al. (2003) in an intermediate GCM. However, fixing land-surface temperature is difficult in a full GCM that includes a diurnal cycle. In this study we focus on the *RFP* method, which is straightforward to calculate in most global aerosol models.

2 Method

Two questions arise about the *RFP* method: (1) Is it a valid approach for comparing aerosol effects that includes fast feedbacks and interactions due to the cloud lifetime effect, semi-direct effect or aerosol interactions with mixed-phase and ice clouds with other forcings such as those from the well-mixed greenhouse gases (GHG) and (2) Can it be used for comparing these aerosol effects between different models?

The difference between the forcing (as traditionally defined) and the *RFP* due to the aerosol indirect effect was first investigated by Rotstayn and Penner (2001). They found from their atmospheric GCM coupled to a mixed layer ocean model that the differences in the climate sensitivity due to using the *RFP* method were smaller than the differences in the climate sensitivity due to different forcings. They hence argued that *RFP* estimates from aerosols should be compared to forcing estimates from GHG. The utility of the *RFP* method was further explored for a range of forcing agents by Hansen et al. (2002, 2005), also in the context of a single GCM; they similarly concluded that it was a useful approach. Put differently, because our interest is in the long-term climate response, which is delayed decades to centuries by the ocean's thermal inertia, it is reasonable to allow fast feedbacks to be included in the forcing (as in the *RFP* method), since these feedbacks are felt as forcings by the ocean and thus affect the long-term climate response (Hansen et al., 2005). This also makes sense from an energy balance perspective (Murphy et al., 2009) and is more suitable in the conceptual framework of radiative forcing and climate sensitivity (Gregory et al., 2004; Knutti and Hegerl, 2008; Quaas et al., 2009a).

For indirect aerosol effects, the advantage of the *RFP* method over the instantaneous forcing is that it allows the radiative impact of aerosols on both cloud albedo and precipitation efficiency to be evaluated. As shown in Fig. 1, if estimates of other aerosol-cloud interactions are considered in addition to the cloud albedo effect, then

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these estimates are mostly larger than the cloud albedo effect alone. This suggests that most of the model-calculated additional effects do not offset the cloud albedo effect, but rather constitute an additional cooling. Although the total indirect effect shows more scatter than the cloud albedo effect, more recent estimates indicate smaller (less negative) values. Some of the smallest estimates result from estimates of the indirect aerosol forcing from satellite data or from general circulation model (GCM) estimates that constrain the indirect aerosol effect using satellite data. Also, some aerosol interactions with mixed-phase clouds can partly offset the forcing due to the cloud albedo effect.

A complementary approach to estimate the total anthropogenic aerosol effect is to infer it as a residual using the observed temperature record over land, and estimates of the ocean heat uptake and the evolution of greenhouse gas and solar radiative forcing (Anderson et al., 2003; Hegerl et al., 2007) (dashed area in Fig. 1). One estimate includes only the indirect aerosol effect in which case additional assumptions about the direct aerosol effect were made (solid black line in Fig. 1). The so-derived total anthropogenic aerosol effect or indirect aerosol effect would, however, also include any other possible hitherto unknown cooling effect, but this is thought to be small. These so-called inverse estimates constrain the total cooling forcing over the 20th century, attributable to anthropogenic aerosols, to a likely range[1] of -0.1 to $-1.7\,\mathrm{W\,m^{-2}}$ (Hegerl et al., 2007). A total anthropogenic aerosol effect that is more negative than $-1.7\,\mathrm{W\,m^{-2}}$ would thus be inconsistent with the observed warming. An approach that constrains the total cooling effect since 1950 purely from an energy balance perspective limits it to between -0.7 to $-1.5\,\mathrm{W\,m^{-2}}$ (Murphy et al., 2009).

In this paper we compare the forcings due to two well-mixed greenhouse gases, the direct aerosol forcing and the cloud albedo effect as described in Table 1 from five atmospheric GCMs with the respective *RFP* that take fast feedbacks and interactions into account. Indirect aerosol effects beyond the cloud albedo effect cannot be compared this way because they comprise fast feedbacks and interactions and thus

[1]likely refers to a >66% probability of occurrence

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no forcing calculation can be done for them. The versions of the participating GCMs are: CSIRO in low resolution (Rotstayn et al., 2007; Rotstayn and Liu, 2009), EC-Earth (Storelvmo et al., 2009), ECHAM5 (Lohmann et al., 2008), GISS (Menon et al., 2008), and HadGEM2 (Collins et al., 2008). These models vary in the complexity with which they describe aerosol-cloud interactions and thus provide a reasonable spread in radiative forcing and radiative flux perturbation estimates. All models include anthropogenic emissions of sulfate precursors, organic and black carbon. Therefore the direct aerosol effect accounts for black carbon in all models and the semi-direct effect of black carbon is accounted for in the *RFP* calculations. However, only in the CSIRO and ECHAM5 GCMs does hydrophilic black carbon also contribute to the number of cloud droplets and thus to the cloud albedo effect. The radiative forcing and *RFP* calculations are conducted by using prescribed sea-surface temperature and sea ice extent, which is also referred to as the Hansen-style method or "quasi-forcing" (Rotstayn and Penner, 2001) to estimate forcing (Hansen et al., 2002).

For the forcing calculations using the traditional forcing definition, denoted F, the radiation code of the models was called twice keeping the meteorology fixed. The differences between two radiative transfer calculations due to pre-industrial GHG or aerosol concentrations versus their present-day values were extracted at the top-of-the-atmosphere and at the tropopause (or at 100 hPa which some GCMs took as a surrogate for the tropopause). The forcing calculation at the tropopause is the instantaneous value, which does not account for the fast stratospheric temperature adjustment as a response to the warming due to molecular absorption by greenhouse gases (Hansen et al., 1997). Calculation of the adjusted forcing in a GCM would require offline radiative computations or other elaborate procedures (Stuber et al., 2001), so we take the instantaneous value as an approximation to the adjusted value. Results shown in Table 1 of Hansen et al. (2005) suggest that the instantaneous forcing for present-day minus pre-industrial CO_2 is roughly 10% larger than the adjusted forcing (1.55 and 1.40 $W\,m^{-2}$ respectively, for a CO_2 change from 291 to 370 ppm). In the second set of experiments, the simulations were run for 5–10 years each after a spin-up period of

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several months under conditions appropriate for the present-day climate, a time scale which allows for all fast feedbacks to fully act. The *RFP* is then defined as the difference in global mean net TOA radiative fluxes between two such simulations with the same sea surface temperatures, one with perturbed, and one with unperturbed forcing agents. As the meteorology is different when varying greenhouse concentrations or aerosols, here the radiative effects of the forcing agents will be evaluated as *RFP*, defined as the difference in the net TOA radiation balance between the pre-industrial and present-day simulations.

In cases where GCMs have aerosols that interact with cloud microphysics and where the aerosols are radiatively active at the same time, *RFP* calculations for individual aerosol effects are more complicated. Here the interaction between aerosols and cloud droplets is artificially deactivated by prescribing a cloud droplet number concentration N_c for the calculation of precipitation formation. Moreover, aerosol concentrations were put to zero for the forward integration in time of the model. Then the forcings due to the direct aerosol effect and the cloud albedo effect are obtained from the difference of the forcing calculations in a simulation with present-day and one with pre-industrial emissions. Taking the difference between present-day and pre-industrial forcing is necessary as in each simulation the total forcing (present-day minus zero aerosols and pre-industrial minus zero aerosols) is calculated. *RFP* calculations are performed as for GHGs. For all radiative flux perturbations, the interannual standard deviation is calculated as $\sqrt{2/n \times SD_p}$, where n is the number of years in the simulation and SD_p is the pooled standard deviation (Snedecor and Cochran, 1989).

3 Radiative forcing versus radiative flux perturbation

The estimates of *RFP* vs. *F* at TOA and at the tropopause for the different forcing agents from the five GCMs are shown in Fig. 2. The difference between tropopause and TOA forcing is mainly important for CO_2 as an increase in CO_2 warms the troposphere

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but cools the stratosphere. If a stratospheric temperature adjustment would have been allowed in these simulations, then *F* at TOA would equal *F* at the tropopause. Therefore for CO_2 *RFP* at TOA should rather be compared to *F* at the tropopause (right panel), which is a reasonable approximation to the adjusted forcing. If the *F* values in the right panel were reduced by about 10%, to account for omission of the stratospheric adjustment in our runs (Hansen et al., 2005), the slope and the correlation coefficient of the least squares fit through the data would be further improved.

For the majority of these different estimates, the *F* values for the net radiation at the tropopause fall within the *RFP*± their interannual standard deviation. Deviations occur mainly for the larger forcings (carbon dioxide and the first indirect effect) especially for those models with larger forcings for a given species. For individual models explanations can be found that relate to the way the cloud feedback differs in these simulations. The negative *F* and *RFP* values for the aerosol effects and their deviations from the one-to-one line are reflected in the shortwave *F* and *RFP* values. The positive *F* and *RFP* values for the greenhouse gases and their deviations from the one-to-one line are dominated by their longwave signals (Fig. 2). The scatter plots of *F* versus *RFP* also include some earlier literature estimates by Rotstayn and Penner (2001) and Hansen et al. (2005).

The deviation from the 1:1 line in the CO_2 *RFP* vs. forcing at the tropopause may be indicative of a semi-direct cloud response to CO_2 forcing. This should be investigated in terms of the differences from the 1:1 line in the CO_2 *RFP* vs. forcing at the tropopause for all-sky minus clear-sky conditions. However, as no model saved the clear-sky forcing data at the tropopause, we attempt to estimate the semi-direct cloud response to CO_2 forcing from the comparison of the difference in net radiation (*RFP* – TOA forcing for all-sky conditions) – (*RFP* – TOA forcing for clear-sky conditions) assuming that this difference will not be that different at the tropopause and at TOA. The multi-model average amounts to $0.15\,\mathrm{W\,m^{-2}}$, in agreement with the small positive semi-direct cloud response to CO_2 forcing found by Gregory and Webb (2008).

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Deviations between the forcing and *RFP* estimates are smaller in the clear-sky case where the influence of cloud feedbacks is much smaller (Fig. 3). Unfortunately the clear-sky results are only available for the TOA forcing but not for the tropopause forcing. Changes in total cloud cover, liquid and ice water path remain below 1% of their present-day values in all *RFP* simulations and models (not shown). Thus, the zonal and annual mean pattern of the *RFP* estimates are a noisy version of the forcing distributions because of the inclusion of fast interactions and feedbacks in the latter but are not fundamentally different (Figs. 4–6).

4 Conclusions

In this paper we argue that feedbacks and interactions that are fast as compared to the time scale of global warming should be included when estimating the total anthropogenic aerosol effect. Doing so allows the total anthropogenic aerosol effect, which we cannot evaluate as a forcing precisely because it includes fast feedbacks and interactions and needs to be obtained from the *RFP* method, to be compared to the forcings due to well-mixed greenhouse gases.

We showed that the zonal and annual mean pattern of the *RFP* estimates are a noisy version of the forcing distributions but do not differ systematically. The global annual mean values mostly fall within the interannual standard deviation of the *RFP* simulations. This is a very powerful result as it shows that *RFP* estimates are consistent with forcing calculations using the traditional approach for all the species/effects considered here.

We thus conclude that assessing different forcing agents with the *RFP* method is a valid option to be considered in future IPCC reports. Moreover, replacing the global-mean aerosol forcing by its *RFP* has its merits because it is the overall aerosol flux perturbation that is needed for the global energy balance (Murphy et al., 2009).

Appendix A

References for Fig. 1

A1 Cloud albedo effect

Kaufman and Chou (1993), Jones et al. (1994), Boucher and Lohmann (1995), Chuang et al. (1997), Feichter et al. (1997), Lohmann and Feichter (1997), Rotstayn (1999), Lohmann et al. (2000), Kiehl et al. (2000), Jones et al. (2001), Williams et al. (2001), Ghan et al. (2001), Rotstayn and Penner (2001), Chuang et al. (2002), Kristjánsson (2002), Rotstayn and Liu (2003), Suzuki et al. (2004), Quaas et al. (2004), Dufresne et al. (2005), Ming et al. (2005), Chen and Penner (2005), Takemura et al. (2005), Quaas and Boucher (2005), Penner et al. (2006), Kvalevag and Myhre (2007), Quaas et al. (2008), Lebsock et al. (2008), Wang and Penner (2009), Storelvmo et al. (2009), Rotstayn and Liu (2009), Haerter et al. (2009)

A2 Total aerosol indirect effect

A2.1 Cloud albedo and cloud lifetime effect

Lohmann and Feichter (1997), Rotstayn (1999), Lohmann et al. (2000), Jones et al. (2001), Williams et al. (2001), Ghan et al. (2001), Lohmann and Lesins (2002), Menon et al. (2002), Kristjánsson (2002), Peng and Lohmann (2003), Kristjánsson et al. (2005), Ming et al. (2005), Rotstayn and Liu (2005), Takemura et al. (2005), Quaas et al. (2006), Storelvmo et al. (2006), Storelvmo et al. (2008a), Rotstayn and Liu (2009), Hoose et al. (2009)

A2.2 Cloud albedo, cloud lifetime, direct and semi-direct effect

Lohmann and Feichter (2001), Penner et al. (2003), Penner et al. (2006), Lohmann et al. (2007), Rotstayn et al. (2007), Posselt and Lohmann (2008), Posselt and Lohmann (2009), Quaas et al. (2009b)

A2.3 Cloud albedo, cloud lifetime, direct effect and aerosol effects on mixed-phase clouds

Lohmann and Diehl (2006), Jacobson (2006), Storelvmo et al. (2008a), Hoose et al. (2008b), Storelvmo et al. (2008b), Koch et al. (2009), Lohmann and Hoose (2009)

A2.4 Cloud albedo, cloud lifetime, direct effect and aerosol effects on convective clouds

Menon and Rotstayn (2006), Lohmann (2008), Unger et al. (2009)

A2.5 Inverse estimates of the direct and indirect aerosol effects

Andronova and Schlesinger (2001), Knutti et al. (2002), Gregory et al. (2002), Forest et al. (2002), Knutti et al. (2003), Forest et al. (2006), Stott et al. (2006), Shindell and Faluvegi (2009), Murphy et al. (2009)

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Table 1. Experimental set-up.

Forcing agent	pre-industrial concentration	present-day concentration
CO_2	280 ppm	379 ppm
CH_4	0.715 ppm	1.774 ppm
direct aerosol effect	pre-industrial emissions (1750 or 1860)	present-day (year 2000) emissions
cloud albedo effect	pre-industrial emissions (1750 or 1860)	present-day (year 2000) emissions

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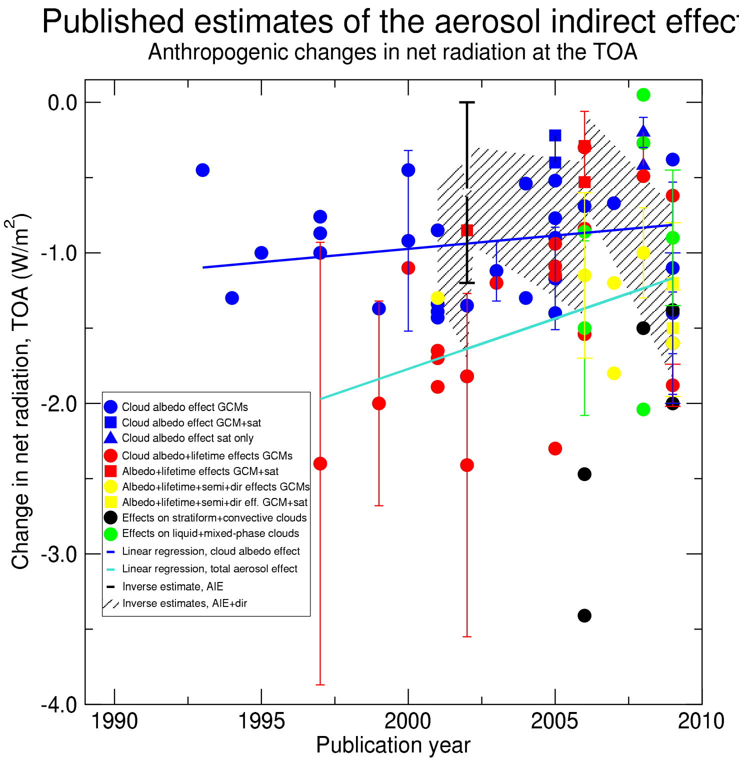

Fig. 1. Model, satellite and inverse estimates of the aerosol indirect effects over the last two decades. Per method or effects considered, each symbol represents one published estimate (one paper one vote). Blue represents estimates of the cloud albedo effect from GCMs (circles), GCMs combined with satellite measurements (squares) and satellite only (triangles). Red represents estimates of both the cloud albedo and cloud lifetime effect from GCMs (circles) and GCMs combined with satellite estimates (squares). The yellow circle represents an estimate of the cloud albedo, lifetime, direct and semi-direct effects. Black circles represent the aerosol effects on stratiform and convective clouds and green circles represent estimates of aerosol effects on liquid and mixed-phase clouds. The black stippled area refers to inverse estimates. In case of multiple estimates per paper, the vertical bars denote the standard deviation. See appendix for the individual papers, from which the estimates are obtained.

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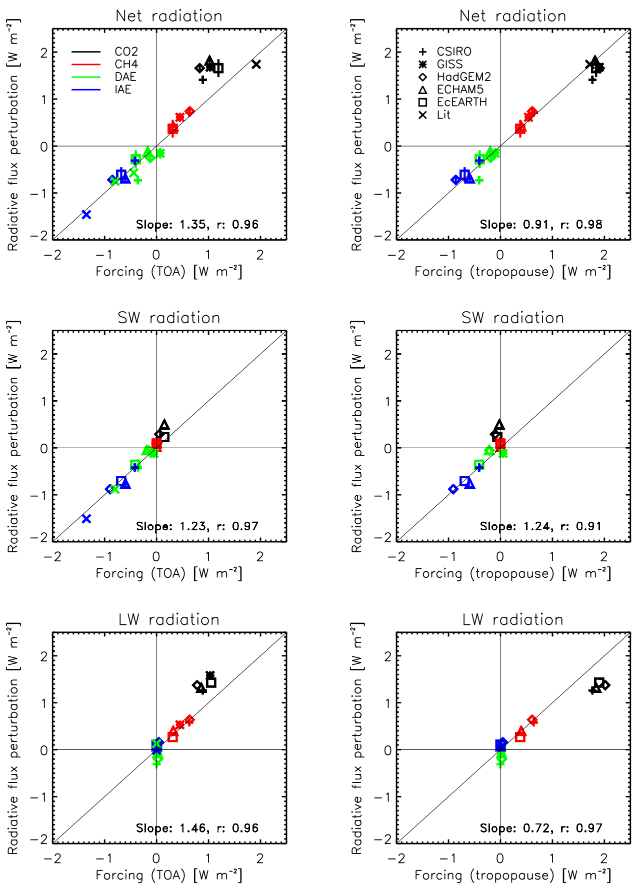

Fig. 2. Net, shortwave and longwave *RFP* versus TOA and tropopause *F*, respectively, from five GCMs. Vertical bars denote the interannual standard deviation in the radiative flux perturbation calculations. The slope of the least square fit through the data as well as the correlation coefficient *r* are shown at the bottom. *RFP* vs. *F* values from the literature (Rotstayn and Penner, 2001; Hansen et al., 2005) are added as well.

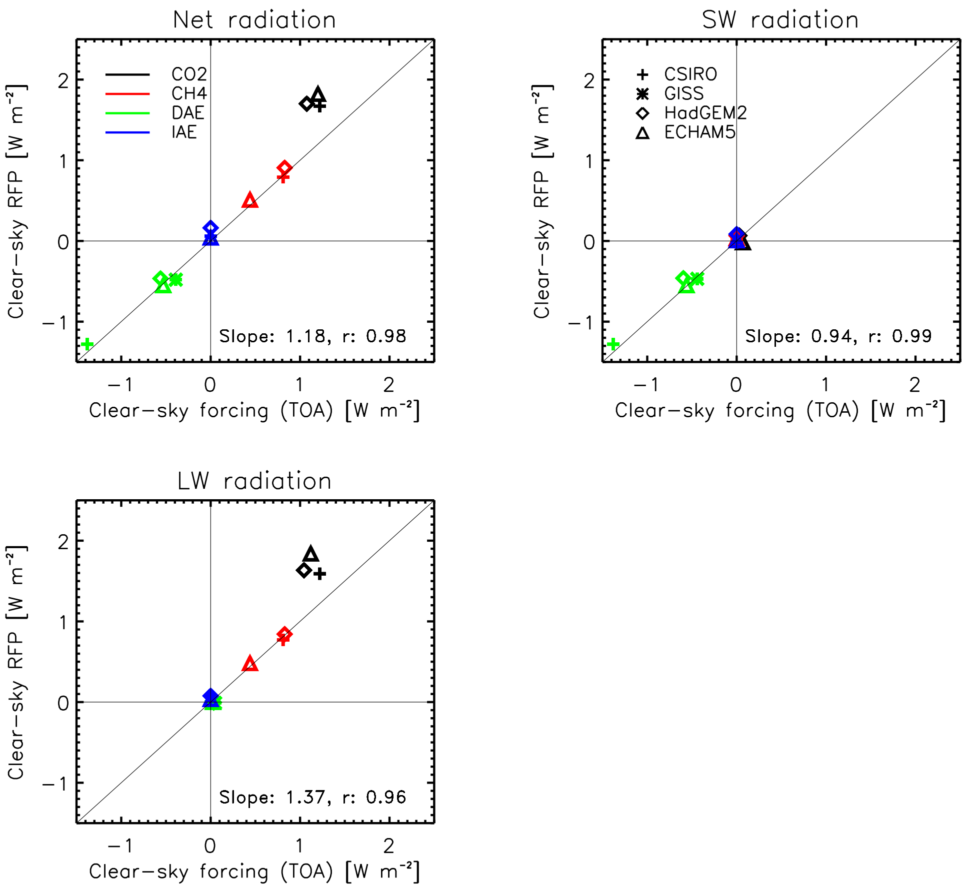

Fig. 3. As Fig. 2, but for the clear-sky net, shortwave and longwave *RFP* versus TOA *F* from four GCMs.

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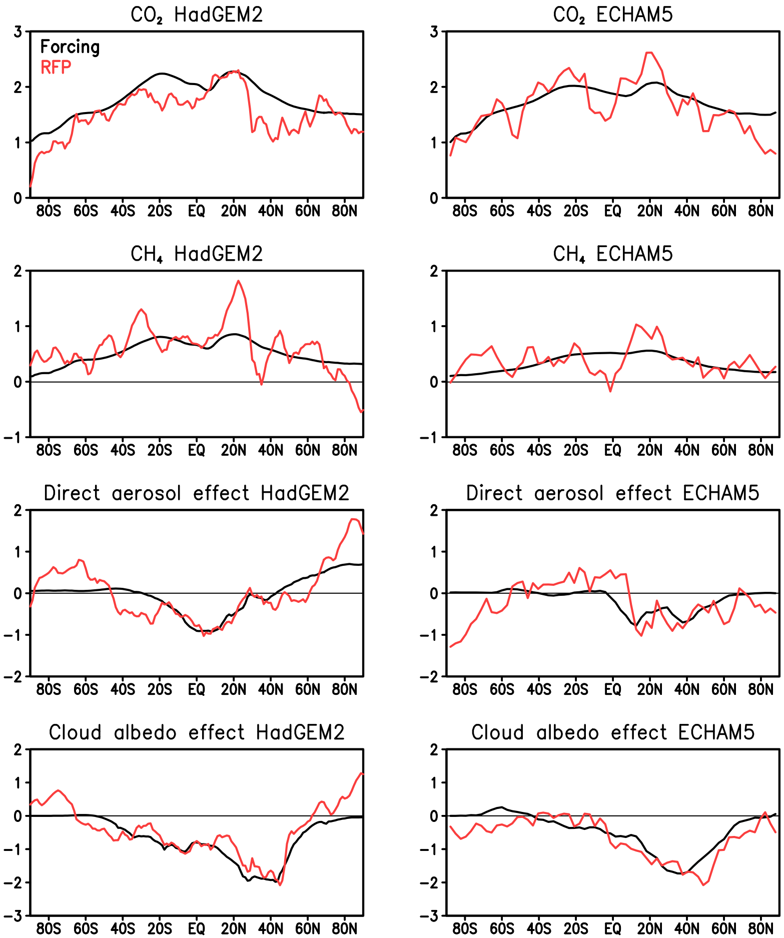

Fig. 4. Annual zonal means of *RFP* vs. *F* [$\mathrm{W\,m^{-2}}$] for the different forcing agents from the HadGEM2 and ECHAM5 GCMs.

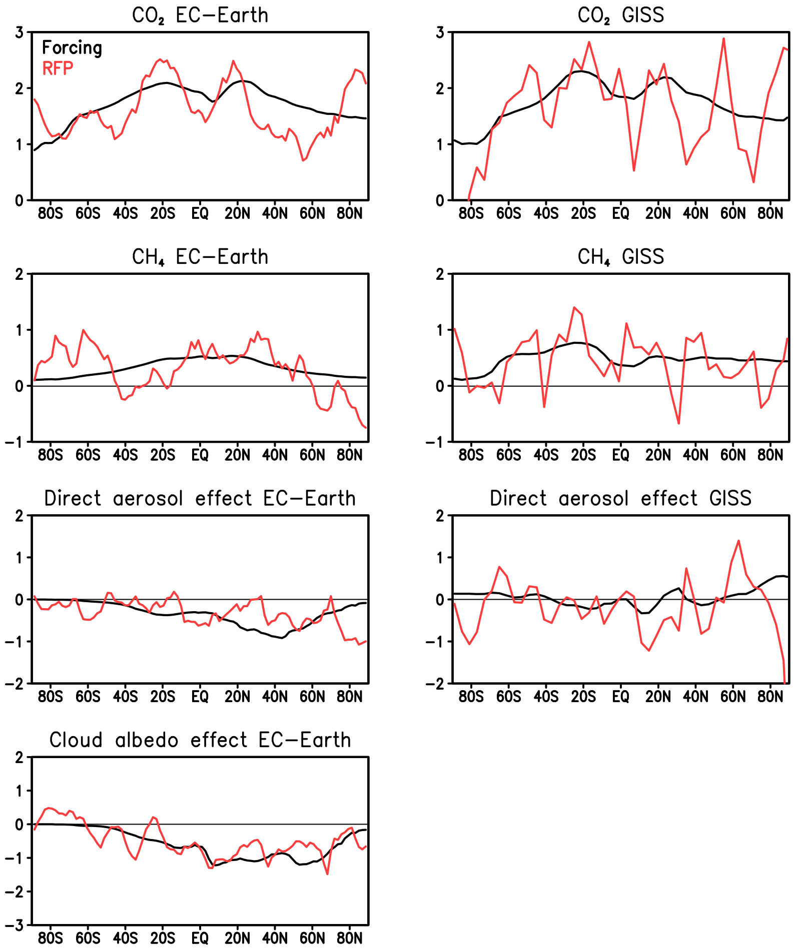

Fig. 5. Annual zonal means of *RFP* vs. *F* [$\mathrm{W\,m^{-2}}$] for the different forcing agents from the EC-Earth and GISS GCMs.

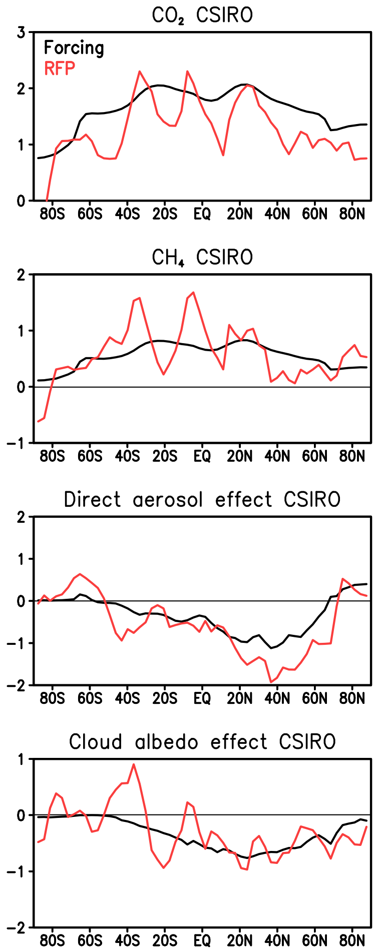

Fig. 6. Annual zonal means of *RFP* vs. *F* [W m^{-2}] for the different forcing agents from the CSIRO GCM.

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