

LETTERS

On the Use of Cloud Forcing to Estimate Cloud Feedback

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ABSTRACT

Uncertainty in cloud feedback is the leading cause of discrepancy in model predictions of climate change. The use of observed or model-simulated radiative fluxes to diagnose the effect of clouds on climate sensitivity requires an accurate understanding of the distinction between a change in cloud radiative forcing and a cloud feedback. This study compares simulations from different versions of the GFDL Atmospheric Model 2 (AM2) that have widely varying strengths of cloud feedback to illustrate the differences between the two and highlight the potential for changes in cloud radiative forcing to be misinterpreted.

1. Introduction

It is widely recognized that climate models exhibit a large range of sensitivities in response to increased greenhouse gas concentrations and that much of this discrepancy is attributable to differences in their treatment of clouds (Cess et al. 1990). A full understanding of the impact that clouds have on climate sensitivity requires an accurate and reliable method for quantifying its strength, that is, an accurate measure of cloud feedback.

Historically, there have been two different approaches for measuring cloud feedback. The first method, introduced by Wetherald and Manabe (1988), uses offline radiative transfer calculations to compute the partial radiative perturbation (PRP) that arises solely from the change in cloud properties between two climate states. This method has the advantage of measuring the differential behavior of the radiative fluxes in response to explicitly controlled independent variables, making the results easy to interpret. The disadvantages of this method are that it is computationally expensive, can be dif-

ficult to implement, and yields a quantity that is impossible to directly compare with observations.

A second, and much simpler, method was developed in a series of pioneering papers by Cess and Potter (1988) and Cess et al. (1990, 1996) and has since been adopted by many modeling groups for the routine evaluation of cloud feedbacks and climate sensitivity. This approach uses prescribed sea surface temperature (SST) perturbations to induce a change in top-of-atmosphere (TOA) fluxes. The resulting changes in clear-sky and total-sky radiative fluxes are then used to infer the clear-sky and total-sky sensitivity of the model, with the difference providing a measure of the contribution of clouds in altering the climate sensitivity. It has the advantage of being relatively straightforward to implement, requires little additional computational overhead, and provides a measure of cloud response whose definition is more consistent with observable quantities.

The two methods, however, are not the same. Zhang et al. (1994) point out that the cloud feedback obtained from the PRP differs from that inferred due to a change in cloud radiative forcing (ΔCRF) because the latter does not account for potential differences in the temperature and water vapor distributions between a clear-sky and a cloudy atmosphere, leading to a “small but non-negligible difference” between the two. More re-

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cently, Colman (2003) compared offline calculations of cloud feedback (from published analyses of solar and $2 \times \text{CO}_2$ climate perturbation experiments using mixed-layer GCMs) with ΔCRF calculations obtained from the idealized SST perturbation experiments reported by Cess et al. (1990). While a direct comparison of the ΔCRF and PRP feedbacks is complicated by the difference in climate perturbations between the two sets of experiments, Colman does note that “a substantial part of the cloud feedback term from Cess et al. (1990) may indeed be simply the cloud impact on the clear sky response.”

Yet, despite these results and the widespread use of SST perturbation experiments to diagnose model sensitivity, many in the climate community are often confused by the subtle distinctions between the ΔCRF and PRP methods. For example, both the 1990 and 1992 Intergovernmental Panel on Climate Change (IPCC) reports (Cubasch and Cess 1990; Gates et al. 1992) refer to cloud feedback as a change in cloud forcing. In this study, we apply both of these methods to the same set of idealized SST perturbation experiments using the Geophysical Fluid Dynamics Laboratory (GFDL) Atmospheric Model 2 (AM2). By comparing the results from different versions of AM2, which have widely varying strengths of cloud feedback, we illustrate more clearly the differences between the two methods and highlight the potential for ΔCRF metrics to be misinterpreted. In particular, we show how positive cloud feedbacks can be associated with negative values of ΔCRF and suggest that while almost half of the models in Cess et al. (1996) have negative values of ΔCRF , many of them, perhaps even all, actually have a positive cloud feedback.

2. The model and experimental design

a. The GFDL AM2

A full description of the relevant model physics in the GFDL AM2 and comparison of its simulated climate with observations are provided in a paper by the GFDL Global Atmospheric Model Development Team (2004). While the model characteristics described there generally apply to AM2, many different versions of this model have been integrated as part of the model development process. The different versions are denoted as AM2p# (where # represents the version number) and have resulted from changes in parameter values and, in some cases, more substantive changes to the model physics. This study exploits the available sequence of AM2 versions to explore differences in their feedback strengths while working within a consistent model and experimental design framework.

b. SST perturbation experiments

Given a direct radiative forcing G , the climate system restores radiative equilibrium by inducing a change in

surface temperature T_s ; that is, $G = \{[\Delta(F - Q)]/\Delta T_s\} \Delta T_s$, where F is the outgoing longwave radiation and Q is the absorbed shortwave radiation at the TOA, and $\gamma = \Delta T_s/[\Delta(F - Q)]$ defines the climate sensitivity parameter. Here, we follow the inverse approach for estimating climate sensitivity outlined by Cess et al. (1990). By imposing a prescribed change in SST, the climate sensitivity can be inferred from the model-simulated change in total-sky TOA fluxes. For each SST perturbation, the cloud radiative forcing (Ramanathan et al. 1989) can also be computed by differencing the clear-sky and total-sky radiative fluxes; that is, $\text{CRF} = (F_{\text{clr}} - F) - (Q_{\text{clr}} - Q)$. Positive values of CRF have a heating effect on the climate, and negative values have a cooling effect. The change in cloud radiative forcing between the two perturbed states (ΔCRF) thus provides a measure of the contribution of clouds in altering the climate sensitivity of the model.

To induce changes in the climate state, we follow the widely used ∓ 2 K SST experimental framework. In our experiments, SSTs are uniformly perturbed from their climatological values for each month of the year, while sea ice coverage is fixed to its seasonally varying climatological distribution. For both sets of perturbed SSTs, model simulations are performed over the full seasonal cycle for a period of 10 yr, with the last 9 yr being used for the analysis.

c. Partial radiative perturbations

In addition to the TOA fluxes, profiles of the temperature, water vapor, cloud properties, and the surface albedo are archived every 3 h from each experiment to permit offline calculation of the climate feedbacks using the PRP method (Wetherald and Manabe 1988; Mitchell and Ingram 1992; Le Treut et al. 1994; Zhang et al. 1994; Colman and McAvaney 1997). For these calculations, the same radiation code used in the model is run offline using archived variables from the SST perturbation experiments as input. The feedbacks associated with a particular mechanism are then estimated by changing only the input variables relevant to that mechanism and computing the resulting perturbation in the net radiation at the tropopause.

To compute cloud feedback, offline radiation calculations are performed using all input from the -2 K simulation with the exception of cloud amount and cloud water paths, which are taken from the $+2$ K simulation; that is, $\delta_c \bar{R} = \bar{R}(T, C + \Delta C, r, \alpha_s) - \bar{R}(T, C, r, \alpha_s)$. Here, $R = F - Q$ is the net upward radiation at the tropopause; T , C , and r are the profiles of temperature, cloud properties, and water vapor, respectively; α_s is the surface albedo; and the overbar indicates global averaging. The total radiative perturbation can be written in terms of the partial contributions from each feedback variable, $\delta \bar{R} = \delta_T \bar{R} + \delta_C \bar{R} + \delta_r \bar{R} + \delta_{\alpha_s} \bar{R}$, which may be further separated into longwave and shortwave components. A feedback parameter for each variable X

TABLE 1. Total-sky climate sensitivity parameter ($K m^2 W^{-1}$) for versions of AM2.

P5	P7	P9	P10	P12a	P12a (Jul)
0.68	0.66	0.70	0.79	0.57	0.61

can then be written as $\lambda_X = -(\partial\bar{R}/\partial X)(dX/dT_s)$ for $X = T, C, r$, and α_s such that $\Delta T_s = -G/\lambda$, where $\lambda = (\lambda_T + \lambda_C + \lambda_r + \lambda_\alpha)$, which may also be separated into its longwave and shortwave components.

3. Results

Table 1 compares the climate sensitivity parameter γ for versions of the GFDL AM2 in which SST perturbation experiments were available. For reference, the climate sensitivities of the 19 GCMs considered by Cess et al. (1990) range from approximately 0.4 to 1.2 $K m^2 W^{-1}$ (under perpetual July conditions). When these versions of AM2 were coupled to a mixed-layer ocean model, their sensitivities to a doubling of CO_2 ranged from 2.3 K (AM2p12a) to 4.6 K (AM2p10), indicating that the AM2 model versions occupy a sizable portion of the 1.5–4.5 K intermodel range noted by IPCC and that the SST perturbation method qualitatively captures the interversion differences in AM2 sensitivity.

The largest transition in model sensitivity occurs between AM2p10 and AM2p12a, which, in fact, represent the most and least sensitive versions of AM2. To help understand the cause of the sensitivity change, Fig. 1 shows the individual feedback components λ_X for these two versions computed from the archived output of the SST perturbation experiments using the PRP method. The sign convention is such that positive values of λ_X indicate a positive feedback and vice versa. As expected, the temperature (Planck) feedback (which includes contributions from both the surface warming and lapse-rate changes) represents the strongest negative feedback in both versions, while the water vapor feedback is by far the strongest positive feedback. The large difference in sensitivity between the two versions is attributable to differences in the cloud feedbacks, which are shown separately for the longwave and shortwave components. While the longwave cloud feedback is reduced by ~30% between AM2p10 and AM2p12a, the shortwave feedback changes sign, going from a weak positive feedback (AM2p10) to a relatively strong negative feedback (AM2p12a). It is this change that is responsible for the large reduction in sensitivity in AM2p12a. Note that the linear sum of the individual feedbacks (“Sum”) agrees well with the total radiative perturbation between the -2 and +2 experiments (“Total”), supporting the validity of the linear feedback approximation at the global scale (Colman et al. 1997).

To assess the consistency between the CRF and PRP methods, Fig. 2 compares the cloud feedback parameters from each version of the GFDL AM2. The PRP feed-

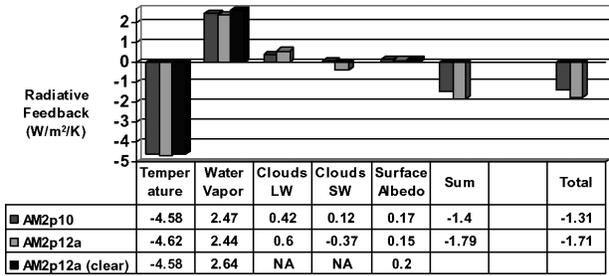


FIG. 1. Comparison of total-sky feedback strengths between AM2p10 and AM2p12a. Also shown are the clear-sky feedback strengths from AM2p12a for temperature, water vapor, and surface albedo.

back parameter is defined as in section 2c. The CRF parameter is defined as the change in global-mean CRF between the -2 and +2 SST experiments, divided by the change in global-mean surface temperature, that is, $\Delta CRF/\Delta T_s$. Both are shown separately for the longwave, shortwave, and net radiative fluxes.

Comparison of the shortwave cloud feedback (Fig. 2a) indicates that, while the two methods differ in the

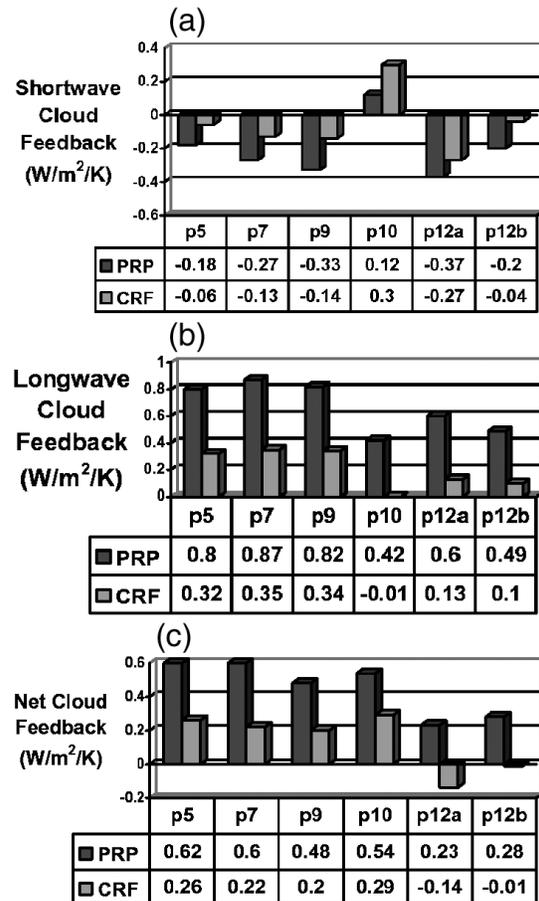


FIG. 2. Comparison of the (a) shortwave, (b) longwave, and (c) net cloud feedback parameters for the PRP and CRF methods.

absolute magnitude, as might be expected, their measure of version-to-version changes is relatively consistent. The change in sign of the cloud feedback from negative to positive (p9 to p10) and positive to negative (p10 to p12a) is captured by both methods. Likewise model versions with weakly negative shortwave cloud feedback (p5, p12b) are generally distinguished from those with stronger negative cloud feedback (p12a).

Comparison of the longwave cloud feedback parameters (Fig. 2b) is less encouraging. While relative consistency between the two methods still occurs, the implications of their differences become more apparent. Most striking are the results for AM2p10 in which the ΔCRF change is -0.01 , indicating effectively no change in longwave cloud forcing between the -2 and $+2$ SST experiments. In contrast, the PRP method still indicates a strong positive longwave feedback from clouds. Indeed, the feedback from AM2p10 is only $\sim 50\%$ weaker than that simulated in AM2p9, whereas the CRF method implies that the strong positive response in AM2p9 has completely disappeared in AM2p10.

The potential for misinterpreting ΔCRF becomes even more apparent for the net cloud feedback (Fig. 2c). In this case, the CRF method indicates a reduction in net cloud forcing for both AM2p12a and (to a lesser extent) AM2p12b, implying that the change in cloud properties between the -2 and $+2$ K SST perturbations is having a negative feedback on the climate. But, in fact, the PRP method reveals that both model versions retain a sizable positive net cloud feedback, albeit one that is reduced by $\sim 50\%$ relative to earlier versions.

The cause of this discrepancy stems from the effects of clouds in masking the noncloud feedbacks (i.e., temperature, water vapor, and surface albedo). Consider the net cloud feedback, which, in the PRP method, is defined as $\delta_c \bar{R} = \bar{R}(T, C', r, \alpha_s) - \bar{R}(T, C, r, \alpha_s)$, where C' represents the altered cloud properties in the perturbed climate; that is, $C' = C + \Delta C$, and so forth for other variables. In contrast, the change in net cloud forcing between climate states is

$$\Delta\overline{\text{CRF}}_{\text{net}} = [\bar{R}(T', C', r', \alpha'_s) - \bar{R}(T', 0, r', \alpha'_s)] - [\bar{R}(T, C, r, \alpha_s) - \bar{R}(T, 0, r, \alpha_s)].$$

For the case of no cloud feedback $\Delta C \equiv 0$, thus $C' = C$ and $\delta_c \bar{R} = 0$. The change in net cloud forcing becomes

$$\Delta\overline{\text{CRF}}_{\text{net}} = [\bar{R}(T', C, r', \alpha'_s) - \bar{R}(T, C, r, \alpha_s)] - [\bar{R}(T', 0, r', \alpha'_s) - \bar{R}(T, 0, r, \alpha_s)].$$

Assuming that ΔT , Δr , and $\Delta \alpha_s$ are nonzero, the only way that $\Delta\overline{\text{CRF}}_{\text{net}} = 0$ would be if the changes in total-sky flux (left-hand brackets) and clear-sky flux (right-hand brackets) due to noncloud feedbacks were equal. In other words, there would have to be no effect of cloud masking on the total-sky fluxes. As shown below, this is not the case.

The impact of cloud masking on the radiative feedback is quantified in Fig. 1 by comparing the strengths of the total-sky and clear-sky PRP feedbacks for AM2p12a. This comparison shows that the presence of clouds strengthens the Planck feedback (i.e., making it more negative) because the cloud opacity enhances the radiative impact of the upper-tropospheric warming, which is, on average, significantly larger than the surface warming. On the other hand, the masking effect of clouds weakens the water vapor feedback (i.e., making it less positive) by preferentially shielding regions of strong water vapor feedback, such as the deep Tropics (Held and Soden 2000). A qualitatively similar “cloud masking” effect occurs for the surface albedo feedback, which is also weaker in the presence of clouds, indicating that larger-than-average albedo reductions occur in regions typically obscured by clouds. The net result is a systematic overestimate of $\sim 0.3 \text{ W m}^{-2} \text{ K}^{-1}$ in the clear-sky calculation of feedbacks relative to that obtained under total-sky conditions.

Since the ΔCRF is derived by differencing clear-sky and total-sky radiative sensitivities, the overestimate of noncloud feedbacks leads to a systematic underestimate of the cloud feedback by a similar amount. Comparison of the PRP and ΔCRF metrics for net cloud feedback (Fig. 2c) shows that the magnitude of this underestimate in AM2 ranges from ~ 0.3 to $0.4 \text{ W m}^{-2} \text{ K}^{-1}$ (Fig. 2c), which is consistent with the discrepancies noted between the total-sky and clear-sky feedback strengths (Fig. 1).

While the $\sim 0.3 \text{ W m}^{-2} \text{ K}^{-1}$ cloud-masking bias is small relative to the magnitudes of the Planck and water vapor feedbacks, it is not negligible compared to the magnitude of cloud feedback and does alter the interpretation of the sign of cloud feedback in some versions of AM2 (e.g., p12a and p12b). More generally, accounting for the effects of cloud masking on the ΔCRF results of Cess et al. (1996) substantially alters the perceived distribution of cloud feedback in that study. For example, roughly half (8 out of 18) of the models in the Cess et al. study have $\Delta\text{CRF} < 0$. However, if we consider Fig. 3 of Cess et al. (1996) and assume that $G = 4 \text{ W m}^{-2}$ and $\Delta T_s = 4 \text{ K}$ for the $+2/-2 \text{ K}$ experiments, then the strongest negative ΔCRF of the 18 models is only about $-0.3 \text{ W m}^{-2} \text{ K}^{-1}$. Thus, adding the cloud-masking offset of $0.3 \text{ W m}^{-2} \text{ K}^{-1}$ to the ΔCRF results in the Cess et al. study would change the sign of virtually all the models with a negative ΔCRF ; that is, all of the models would have a positive cloud response, as opposed to being almost equally divided between positive and negative values. Such an interpretation is consistent with the survey of Colman (2003), in which virtually all models have a positive net cloud feedback when computed using the PRP method.

4. Discussion

To effectively use observed or model-simulated TOA fluxes to diagnose the effects of clouds on climate sen-

sitivity, one must have an accurate understanding of the distinction between a change in cloud radiative forcing and a cloud feedback. Because positive values of CRF indicate a heating effect of clouds on the climate system, it is quite natural to expect that a decrease in CRF in response to an increase in temperature implies a negative feedback by clouds. Indeed, both modeling and observational studies have assumed this to be true (e.g., Cess et al. 1996; Tsushima and Manabe 2001).

In this paper, feedback calculations from both the PRP and CRF methods were applied to the same set of GFDL AM2 simulations to illustrate differences between the two and to highlight the potential for Δ CRF metrics to be misinterpreted. The distinction between a change in cloud forcing and a cloud feedback is particularly important when attempting to assess the sign of cloud feedback. In particular, we demonstrate that reductions in cloud forcing can be associated with a positive cloud feedback. Thus, while almost half of the models in Cess et al. (1996) exhibit Δ CRF < 0 , it is likely that many actually have a positive cloud feedback.

These results should not be construed as a repudiation of the Cess et al. studies, the purpose of which was not to quantify precisely the strength of cloud feedback in models. Rather, we wish to emphasize the need to extend the Cess et al. comparisons using the PRP method in order to truly understand the range in sign and strength of cloud feedback, and to urge caution when interpreting either observed or model-simulated changes in cloud forcing.

Likewise, we are not suggesting that Δ CRF metrics of cloud response are not without merit. For example, the Δ CRF method provides a measure of cloud response that can be directly compared to observations. However, if one wishes to understand whether clouds amplify or dampen the climate system's response to an external perturbation, or to understand the range in sign and strength of this feedback in climate models, then a PRP analysis is necessary.

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