

CLOUD-CLIMATE FEEDBACK: HOW MUCH DO WE KNOW?

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This paper introduces the concept of cloud-climate feedback along with its role in the sensitivity of the climate system. It reviews available cloud feedback diagnostic methods and representative results in three-dimensional atmospheric general circulation models. A case study is presented to analyze physical processes responsible for cloud feedbacks that control the sensitivity of the NCAR Community Climate Model. The paper further discusses weaknesses of current numerical models and areas of required research on this subject.

1. Introduction

Potential global warming caused by increasing greenhouse gases in the atmosphere is currently a major environmental concern worldwide. Given the complexity of the climate system, three-dimensional numerical models are the only tools to reliably project future climate changes in response to human activities.

One of the unsettling issues from using these models is the large discrepancy in the magnitude of the simulated global warming in different models. In response to a doubling of CO₂ concentration in the atmosphere, coupled general circulation models (GCMs) projected warming that ranged from 1.9 to 5.2 °C in 1990 (Mitchell et al. 1990). In 1995, this range was 2.1 to 4.6 °C (Kattenberg et al. 1996). In the latest IPCC report, this range is 2.0 to 5.1 °C (Cubasch et al. 2001). This spread of model projections casts large uncertainties to regional climate changes and prompts indeterminable societal responses. There is therefore the need to understand and narrow this model difference.

Cloud-climate feedback emerged as the leading cause of model uncertainties. I use Figure 1 to schematically show the physical concept. Clouds, consisting of water or ice particles, affect radiative transfer. Clouds reflect solar (shortwave) radiation to space, thus serving as a cooling agent to the Earth atmosphere system. This is analogous to an umbrella in the summer to shield the Earth from sunlight. Clouds also act as a greenhouse agent to the infrared radiation (longwave). This role of clouds is analogous to a blanket that keeps the Earth warm. The question is: how do these umbrella and blanket effects of clouds vary as a result of climate change? In a climate warming scenario, if the cooling umbrella becomes larger while the greenhouse blanket does not change, the variation of clouds would offset the warming, and thus a negative feedback. If the greenhouse blanket is larger while the cooling

umbrella is fixed, variation of clouds would amplify the initial warming. In reality, clouds are an internal variable of the climate system. The size and thickness of both the umbrella and the blanket vary with the climate. The exact cooling and warming effects of clouds depend on the height, location, amount, and the microphysical and radiative properties of clouds, as well as their appearance of time with respect to the seasonal and diurnal cycles of the incoming solar radiation. Simple theoretical calculations, with hypothetical yet reasonable assumption on cloud variation, show that clouds can indeed either significantly reduce or amplify a global warming projection.

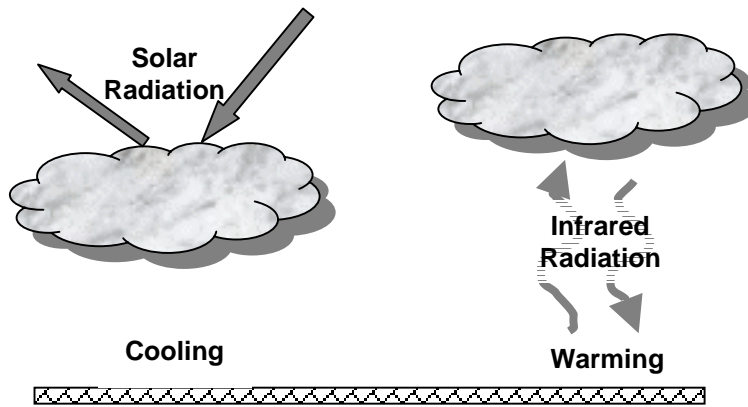


Figure 1. Schematic illustration of the solar and infrared effects of clouds on surface temperature.

Cloud-climate feedback has therefore risen to the list of highest priority in the U.S. Global Change Program. Several satellite and surface measurement programs were specifically established to narrow model uncertainties of cloud feedbacks. After about 15 years of intensive research, however, this issue is still evasive.

The objective of this paper is to appraise where we stand in the research of cloud-climate feedback, and to point out remaining challenges. The paper is organized as follow. Section 2 introduces the definition of cloud feedback and diagnostic methods in climate models. Section 3 reviews representative cloud feedback results that highlight differences among models. Section 4 reports a case analysis by using the latest version of the NCAR Community Atmospheric Model Version 2 (CAM2). Section 5 further evaluates simulated clouds in the CAM2 by using observations. The last section contains a summary and a discussion on challenges and related science issues.

2. Definition and Diagnostic Methods

2.1. Terminology

The quantitative definition of feedback was first used in electrical signal control systems (Bode 1945). Given forcing ΔQ to a system, if the system responds by one measure ΔT_0 , and

if this response does not impact the initial forcing, the system is said to be without feedback. This is schematically shown in Figure 2a. When the relationship between the response and the forcing is written as

$$\Delta T_0 = G_0 \Delta Q, \quad (1)$$

G_0 is defined as the zero-feedback gain. It describes the system response per unit forcing. The system does not have to be linear. Nevertheless, it is more convenient to consider both the forcing and response as small perturbations, thus the linear approximation holds, so that G_0 is independent of the magnitude of the forcing.

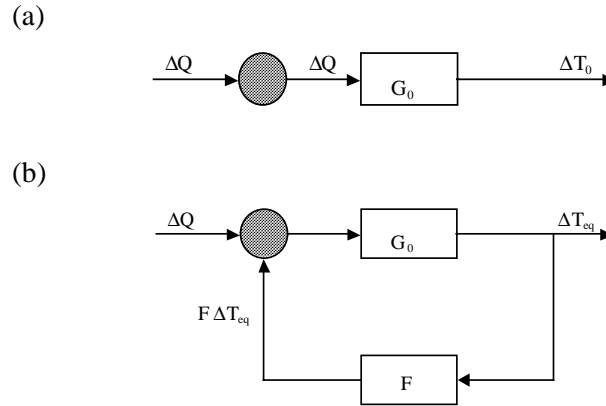


Figure 2. System response to a forcing: (a) without feedback, (b) with feedback.

If the system response also induces changes in the forcing, as schematically shown in Figure 2b, and if the induced forcing is written as proportional to the final response ΔT_{eq} by a factor of F , namely $F \Delta T_{eq}$, then F is called the feedback. Feedback therefore is *the induced forcing from a unit response*. The total forcing to the system at the final state is $\Delta Q + F \Delta T_{eq}$. As a result,

$$\Delta T_{eq} = G_0 (\Delta Q + F \Delta T_{eq}). \quad (2)$$

It follows from the above that

$$\Delta T_{eq} = \frac{G_0 \Delta Q}{1 - G_0 F}. \quad (3)$$

This equation can be also written in terms of the zero feedback response ΔT_0 as

$$\Delta T_{eq} = \frac{\Delta T_0}{1-f}, \quad (4)$$

where

$$f = G_0 F. \quad (5)$$

The dimensionless parameter f is called the feedback factor. It scales the feedback against the zero-feedback gain.

Since there could be more than one process that can induce changes in the forcing, Equation (2) can be expanded to:

$$\Delta T_{eq} = G_0 (\Delta Q + \sum_i F_i \Delta T_{eq}). \quad (6)$$

Equation (4) then becomes

$$\Delta T_{eq} = \frac{\Delta T_0}{1 - \sum_i f_i}, \quad (7)$$

where

$$f_i = G_0 F_i. \quad (8)$$

While feedbacks and feedback factors are additive, the system response to additional feedback is not. Theoretically, the denominator in (4) and (7) could be zero or negative. In the case of negative value, the final response of the system is opposite to what is initially forced. In the zero case, the system is unstable.

2.2. Application to climate models

If we write the net *upward* radiative flux at the top-of-the-atmosphere (TOA) as N , then

$$N = N(\text{Solar and GHG Forcing}, T_s, H_2O, \text{Clouds}, \text{Snow}, T_{air}, \dots) \quad (9)$$

where *GHG* represents greenhouse gases” such as CO_2 and methane etc., T_s is the surface temperature, other variables are as commonly used. With a small climate change from one equilibrium to another caused by a forcing, since $\Delta N = 0$, one has

$$-\Delta N_{forcing} = \frac{\partial N}{\partial T_s} \Delta T_s + \frac{\partial N}{\partial (H_2O)} \frac{d(H_2O)}{dT_s} \Delta T_s + \frac{\partial N}{\partial (Clouds)} \frac{d(Clouds)}{dT_s} \Delta T_s + \dots \quad (10)$$

In the above, $-\Delta N_{forcing}$ is the change of net *downward* radiation at TOA as a result of a direct forcing (since N is defined as the upward flux), such as from an increase in solar radiation or increase of CO_2 . The right hand side of the above equation describes the increase of upward radiation as a result of the induced climate change. Equation (10) is therefore a simple balance of forcing and response.

Without internal feedbacks --- no contributions from water vapor, clouds and other terms on the right hand side of (10) except for the surface temperature, Equation (10) becomes

$$-\Delta N_{forcing} = \frac{\partial N}{\partial T_s} \Delta T_s . \quad (11)$$

When the simple forcing and response format of Equation (1) is used, the above is written as

$$\Delta Q = G_0^{-1} \Delta T_s , \quad (12)$$

where the zero-feedback gain is

$$G_0 = \left(\frac{\partial N}{\partial T_s} \right)^{-1} . \quad (13)$$

G_0 can then be estimated with the Stefan-Boltzman law using an effective atmospheric emissivity ε through

$$\frac{\partial N}{\partial T_s} = \varepsilon 4\sigma T_s^3 = \frac{4 \times \text{OLR}}{T_s} .$$

where σ is the Stefan-Boltzman constant, and OLR is the outgoing longwave radiation at TOA. With 240 W m^{-2} for OLR, and 280 K for T_s , the zero feedback gain is estimated to be about $0.3 \text{ K/(W m}^{-2}\text{)}$. For a doubling of CO_2 , the greenhouse forcing is about 4 W m^{-2} , which corresponds to a direct warming of 1.2 K .

With feedbacks from water vapor, clouds and other processes, (10) can be written as

$$\Delta Q = \frac{1}{G_0} \Delta T_s + \frac{\partial N}{\partial(H_2O)} \frac{d(H_2O)}{dT_s} \Delta T_s + \frac{\partial N}{\partial(Clouds)} \frac{d(Clouds)}{dT_s} \Delta T_s + \dots \quad (14)$$

or

$$\Delta T_s = \frac{G_0 \Delta Q}{1 - G_0 F} , \quad (15)$$

where the total feedback is

$$F = -\frac{\partial N}{\partial(H_2O)} \frac{d(H_2O)}{dT_s} - \frac{\partial N}{\partial(Clouds)} \frac{d(Clouds)}{dT_s} - \dots = -\frac{\delta_{H_2O} N}{dT_s} - \frac{\delta_{Clouds} N}{dT_s} - \dots \quad (16)$$

The individual feedbacks are thus the partial differentiation of the net downward radiative flux at TOA with respect to the individual physical process (water vapor variation, cloud variation etc.) accompanying a unit change of surface temperature. For example, the first term on the right hand side of (16) is the water vapor feedback, and the second term is the cloud feedback.

This definition of feedbacks for climate models is also used in Wetherald and Manabe (1988) and Schlesinger (1988). The feedback has a unit of $\text{W m}^{-2} \text{K}^{-1}$.

Some caveats are pointed here. First, the original definition of feedback was introduced for a zero-dimensional system. For the three-dimensional climate system, it is customary to use the globally averaged net radiation at TOA as the forcing and the globally averaged surface temperature as the response. The partial differentiations, however, should be evaluated from spatially varying physical quantities. Second, since air temperature is strongly coupled to surface temperature, in the zero-feedback calculation, the atmospheric temperature change is often assumed to be the same as that of the surface. The remaining air temperature variation can be considered as a temperature lapse rate feedback. The decomposition of feedbacks is therefore subjective that should be guided by the gain of physical insights. Third, feedback analysis does not directly lead to improved models. It, however, helps to pinpoint why a model is sensitive or insensitive to a forcing. This is similar to a physician diagnosing a disease without directly curing it.

2.3. Diagnostic methods

Based on the above discussion, there are two ways to calculate feedbacks in climate models. One is to impose forcing and calculate the response in a model. Feedback processes are introduced into the model one at a time by arbitrarily holding other physical quantities fixed. The coefficient terms of ΔT_s in Equation (14) are then obtained by differencing results from two experiments. This approach is feasible for feedback studies using simple climate models since it requires multiple climate change integrations of the model.

The second method is to directly calculate the partial differentiations of radiative flux in Equation (16) by using a single climate change simulation. The partial differentiations with respect to various physical quantities are calculated offline. This method was used in Wetherald and Manabe (1988) and in Zhang et al. (1994).

An elegant implementation of the second method, specific for diagnosing model cloud feedbacks, was introduced by Cess and Potter (1988). Instead of obtaining a climate change by imposing a forcing and calculating the system response, they prescribed a simple climate change at the sea surface and calculated the induced radiation perturbation at the top of the

atmosphere. The method thus imposes ΔT_s and calculates the required forcing ΔQ in Equation (14). The partial differentiation of radiative fluxes with respect to clouds is calculated from the change of cloud-radiative forcing (CRF), which is a standard model diagnostics after ERBE (Ramanathan 1987). The cloud feedback in Equation (16) therefore becomes a simple diagnostic of

$$F_{clouds} = \frac{\Delta CRF}{\Delta T_s}. \quad (17)$$

In Cess and Potter (1988), a uniform plus or minus 2 degree SST perturbation was used. It is noted that cloud feedback can be different under different forcing conditions. Yet, the feedback analysis from this type of surrogate climate change has provided considerable insights about climate model sensitivities. Furthermore, as has been recently shown with several GCMs, cloud feedback diagnosed from this surrogate climate change is consistent with those diagnosed directly from global warming simulations.

3. Model Results

3.1. Results from early models

Early studies of cloud feedbacks were carried out through zero-dimensional energy balance models and one-dimensional radiative-convective models (RCM). Since clouds are not explicitly calculated in these models, empirical relationships between clouds and temperature had to be used. Representative works include Budyko (1969), Schneider (1972), and Cess (1975) among others. A review of these studies can be found in Schlesinger (1988). In the present paper, we restrict our discussion to cloud feedback analysis in GCMs.

Hansen et al. (1984) were the first to diagnose cloud feedback from a GCM. They combined the GISS GCM output from a CO₂ climate change simulation with a one dimensional RCM. The RCM was used to calculate the change of surface temperature in response to a doubling of CO₂, with variations of water vapor, clouds, and snow replaced by those from the GCM simulations. In the case of cloud feedback, Hansen et al. (1984) separated it into cloud amount feedback and cloud height feedback. The first was obtained by inserting a globally averaged total cloud change in the GCM into the whole column of the RCM. The cloud height feedback was calculated as a residual of the total cloud feedback and the cloud amount feedback. Hansen et al. (1984) reported that both feedbacks were positive as a result of the following process. In a warmer climate, the model had a reduction of clouds and a shift of clouds to higher altitude. Reduction in low and middle clouds has larger impact on solar radiation than on the infrared radiation, and thus the net cloud cooling becomes smaller, a positive cloud amount feedback. On the other hand, the raised cloud altitude enhances the cloud greenhouse effect of clouds, also producing a positive feedback. The diagnosed cloud feedbacks from the two processes are listed in Table 1. With a zero-

feedback gain of $0.3 \text{ K}/(\text{W m}^{-2})$ as discussed earlier, the cloud feedback of $0.73 \text{ (W m}^{-2} \text{ K}^{-1})$ translates to a feedback factor of $0.73 \times 0.3 = 0.22$, which alone amplifies the global warming by a factor of $1/(1 - f) = 1/0.78 = 1.28$.

Table 1. Cloud feedback results

	Cloud amount feedback ($\text{W/m}^2/\text{K}$)	Cloud height feedback ($\text{W/m}^2/\text{K}$)	Total cloud feedback ($\text{W/m}^2/\text{K}$)
Hansen et al. (1984)	0.33	0.40	0.73
Wetherald and Manabe (1988)	0.12	0.25	0.37

Wetherald and Manabe (1988) calculated the partial differentials of the TOA net downward radiation with respect to various physical quantities, by using the GFDL GCM, to derive the feedback components in Equation (16). The cloud amount feedback was obtained by scaling the model cloud radiative forcing with the variation of cloud amount. The cloud height feedback was derived as a residual similarly to Hansen et al. (1984). Modest positive cloud amount and cloud height feedbacks were reported. They are also listed in Table 1. These feedbacks were also attributed to reduction of middle and low clouds and increase of high clouds.

3.2. Model intercomparisons

The similarity of diagnosed positive cloud feedbacks between the early GISS GCM and the GFDL GCM was later found to be a coincidence rather than a true physical consistency. Cess et al. (1990) used the surrogate climate change of plus and minus 2 K over the oceans and diagnosed the cloud feedback through Equation (17) in 19 GCMs. Figure 3 shows the total cloud feedback from these models. It is seen that they range from negative to strongly positive. These feedback values, along with a zero-feedback gain of $0.3 \text{ K}/(\text{W m}^{-2})$, correspond to a feedback factor of -0.2 to 0.4 . This would imply a difference of a factor of two difference in the temperature response to a prescribed forcing.

The Cess et al. (1990) study was updated in Cess et al. (1996) and it was found that the models continue to show serious physical disagreements as measured by cloud feedbacks from infrared and solar components separately, even though the range of the net cloud feedbacks appeared to be smaller.

The next interesting set of GCM cloud feedbacks was reported by Senior and Mitchell (1993, 1996). They used the same version of the unified GCM at the UK Met Office (UKMO) with only small modifications to the cloud scheme. CO_2 climate change simulations were carried out and cloud feedbacks were diagnosed. Table 2 lists cloud feedbacks from three experiments. Experiment B is from the standard UKMO GCM using the cloud scheme as described in Smith (1990). Experiment A differs from the standard model in that cloud particle size is calculated based on the cloud liquid amount through an empirically observed relationship. Experiment C differs from B in that the assumed subgrid scale distribution of total water within a GCM grid is changed from a triangle distribution to

a top-hat distribution. It is seen that the model exhibited quite large changes in cloud feedbacks. The magnitudes of the simulated global warming are also listed in the table and they are very different. This sensitivity has been also demonstrated in other GCMs (Le Treut et al. 1994).

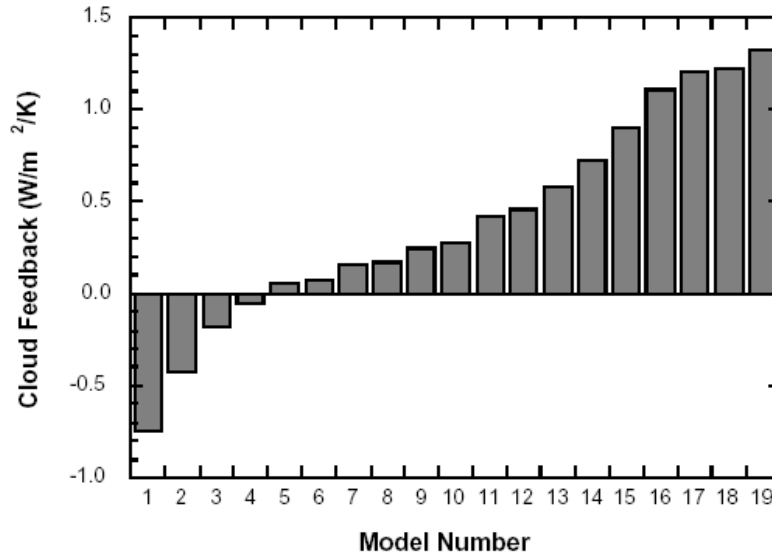


Figure 3. Cloud feedback ($\text{W m}^{-2} \text{K}^{-1}$) in 19 General Circulation Models. Adapted from Cess et al. (1990).

Table 2. Climate change simulation and cloud feedbacks in the UKMO GCM

Experiment ID	A	B	C
ΔT_s ($^{\circ}\text{C}$)	1.9	3.4	5.5
ΔCRF (W/m^2)	-1.04	0.93	3.64
Cloud Feedback ($\text{W/m}^2/\text{K}$)	-0.55	0.27	0.66

Figure 4 shows an updated model intercomparison of the change of CRF from doubling CO_2 simulations in 10 GCMs as reported in Stocker et al. (2001). Since the direct radiative perturbation from a doubling of CO_2 is about 4 W m^{-2} , cloud feedbacks in the figure add up to a range of 2.8 W m^{-2} to over 7.0 W m^{-2} of radiative imbalance in the models, which is sufficiently large to explain the spread in simulated global warming as reported in Cubasch et al. (2001). Figure 4 suggests that cloud feedback uncertainties in current climate models are still about as large as they were fifteen years ago.

At the writing of this paper, an international model intercomparison project is being initiated to carry out a systematic assessment of cloud feedbacks in climate models and why they differ from each other (McAveney and Le Treut, 2003, personnel communication).

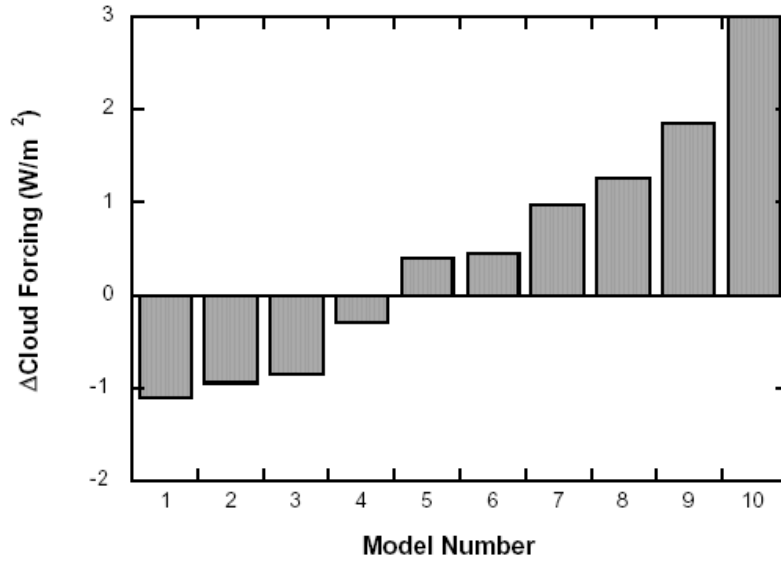


Figure 4. Variation of cloud radiative forcing (W m^{-2}) in doubling CO_2 simulations from 10 General Circulation Models. Adapted from Stocker et al. (2001).

4. Case Study Using the NCAR Community Atmospheric Model

The NCAR Community Climate Model (CCM) is one of the most widely used atmospheric GCMs for climate simulation studies. The model has evolved over the years with continuing modifications and enhancements in its physical components (Williamson et al. 1987; Hack et al. 1993; Kiehl et al. 1996; Collins et al., 2003). Figure 5 shows the evolution of diagnosed cloud feedbacks in CCM0, CCM1, CCM2, CCM3, and the latest Community Atmospheric Model Version 2 (CAM2). It is seen that cloud feedback in this model started from a strong positive feedback to a modest negative feedback since the introduction of the CCM2.

To understand the physical mechanism behind these diagnostics, a brief description of the essential changes to the model cloud parameterization is given here. CCM0 and CCM1 had several important commonalities in their cloud parameterizations. In these models, stratiform clouds were assigned a fixed cloud amount of 95% when there was large-scale condensation in a grid box. Convective clouds were assigned a 30% total amount and the whole convective column was assumed to have randomly overlapping clouds. CCM0 required eighty percent relative humidity for large-scale condensation to occur, while CCM1 required one hundred percent humidity. Cloud radiative properties were specified by using condensed water from large-scale saturation and a moist adiabatic adjustment convection scheme (Ramanathan et al. 1983). Increased cloud amount in the upper troposphere in a warmer climate is likely responsible for the strong positive feedback in these two versions of the model.

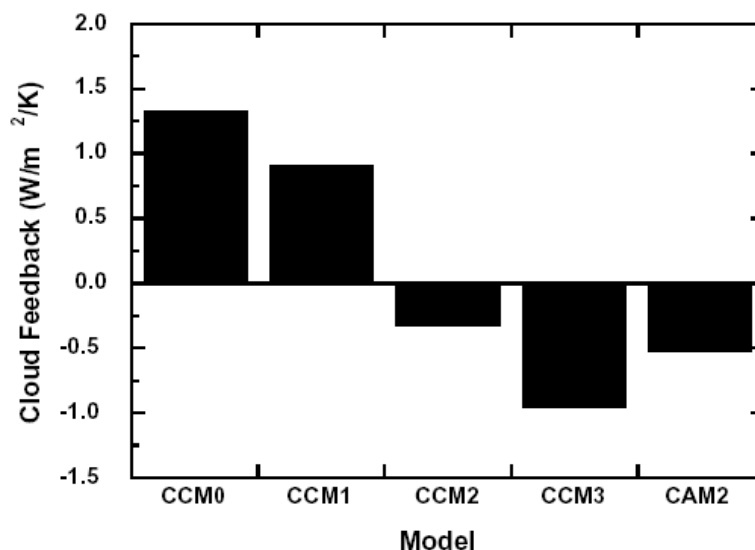


Figure 5. Evolution of cloud feedbacks ($\text{W m}^{-2} \text{K}^{-1}$) in different versions of the NCAR Community climate model. CCM0 and CCM1 results are derived from Cess et al. (1990). CCM2 result is from Zhang et al. (1994). Feedbacks from CCM0 to CCM3 are derived from perpetual July surrogate climate change. Feedback from CAM2 is derived from prescribing sea surface temperature variation from a doubling CO_2 experiment in the UKMO GCM.

CCM2 allowed the calculation of fractional cloudiness following Slingo (1987). Parameterized relationships between grid-scale relative humidity and cloud amount were used. Vertical velocity and stability were also used as input. Convective clouds were parameterized based on convective precipitation from the Hack (1993) mass flux scheme.

CCM3 used essentially the same cloud parameterization as in CCM2, except that convective cloud amount was parameterized based on the deep convective mass flux in the Zhang and McFarlane (1995) scheme. The Hack (1993) scheme is still used to simulate shallow convection, but it does not directly generate clouds.

The latest CAM2 contains three main changes related with clouds. First, boundary layer stratus clouds are calculated to be proportional to the atmospheric vertical stability between 700 mb and surface following the observations of Klein and Hartmann (1993). Second, CAM2 uses a microphysical prognostic cloud condensate scheme of Rasch and Kristjansson (1998) and a macrophysical formulation of Zhang et al. (2003). Third, falling rain is allowed to evaporate in the Zhang and McFarlane (1993) convection scheme.

We will contrast cloud feedbacks in the CCM2 and in the CAM2 to gain some insight in these models. Figure 6a shows the latitude-pressure distribution of the zonally averaged cloud amount in the CCM2 in a perpetual July simulation. It is characterized by two minimum regions of clouds in the subtropics that are associated with the descending branches of the Hadley circulations, and three maximum cloud regions associated with the Inter Tropical Convergence Zone (ITCZ) and middle latitude frontal clouds in the two hemispheres. Note that while these features of vertical cloud distributions qualitatively agree with our understanding of the general circulation of the atmosphere, there are no

observational measurements to directly validate them. Cloud climatologies from ground measurements only give an upward view of the total cloud amount, while those from satellites only give a downward view of the total cloud.

The variation of the cloud pattern in Figure 5a, along with associated changes in microphysical and radiative properties of clouds, is important to the cloud feedback. When the CCM2 is forced with a uniform perturbation of the SST by 4 K, Figure 6b shows the corresponding change of air temperature. One notable feature is that the upper troposphere warms up much more than the surface. This feature is shown in Zhang et al. (1994) to be dependent on the cumulus convection scheme. When the convection scheme is less rigorous, as measured by the percentage of convective precipitation in the total precipitation, there is less amplified warming in the upper troposphere.

Warmer temperature is typically associated with more moisture. It is generally understood that relative humidity varies little in a changed climate, since there is a cancellation between two large opposing changes in temperature and water vapor. However, the residual after the cancellation matters in the cloud variation and cloud feedback. Figure 6c shows the simulated change of clouds in the CCM2. There is a broad reduction of clouds in the middle and upper troposphere, which is closely related with the broad warming in Figure 6b. There is also an increase of cloudiness near the tropopause, associated with increased convection.

The reduction of clouds above 500 mb in the model turns out to dominate the model cloud feedback. Because of the relatively high altitude of cloud variations, the longwave effect overrides the shortwave effect, which permits more longwave radiation to escape to space, and thus a negative feedback in the model. Figure 7 shows the decomposed cloud feedbacks from infrared and solar radiation. The net negative cloud feedback is driven by the negative infrared cloud feedback.

The corresponding feedback components in the CAM2 are also plotted in Figure 7. While both CCM2 and CAM2 have negative net cloud feedbacks, it is seen that in CAM2 it is the negative solar component that dominates the cloud feedback, i.e., the enhanced reflection of solar radiation by clouds. This increase of solar reflection in CAM2 is a result of increased low-level stratus that can be traced to the modifications of cloud parameterizations. Figure 8 shows the latitudinal distribution of CRF variation for the southern summer season. The reduction of cloud forcing is primarily in the shortwave component at around 65°S and in the subtropics of the two hemispheres. Figure 9 shows the geographical distribution of the shortwave CRF in the control simulation and in the warmer simulation in the northern winter. The change of shortwave CRF is distinct around the sea ice line in the southern hemisphere. As the surface warms up, the atmosphere warms up globally. As a result, near the sea ice, the vertical stability of the lower atmosphere is increased, which in turn produces more low clouds in the CAM2 to reflect solar radiation.

The CAM2 modification of rain evaporation in the cumulus convection scheme also weakens the model convection, which is speculated to reduce the magnitude of temperature and cloud variations in the upper troposphere. This should contribute to the disappearance of negative longwave cloud feedback in CAM2.

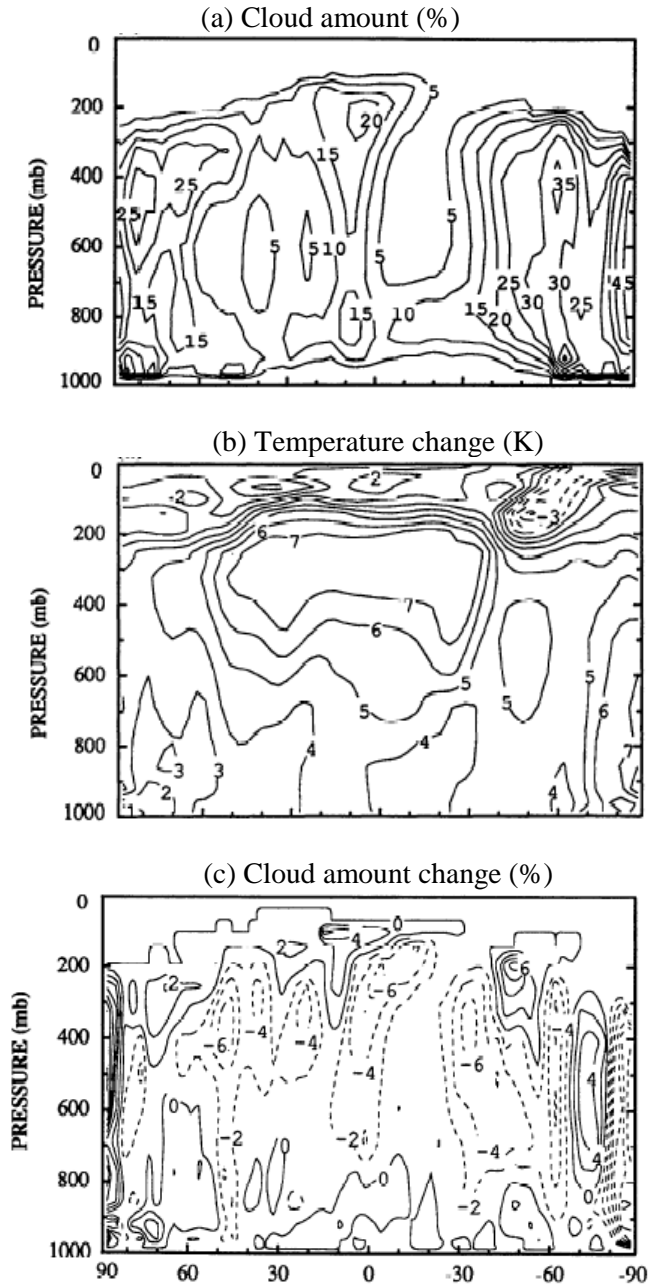


Figure 6. Zonal-pressure distributions of: (a) clouds, (b) temperature variation in perpetual July surrogate climate change, and (c) cloud variation.

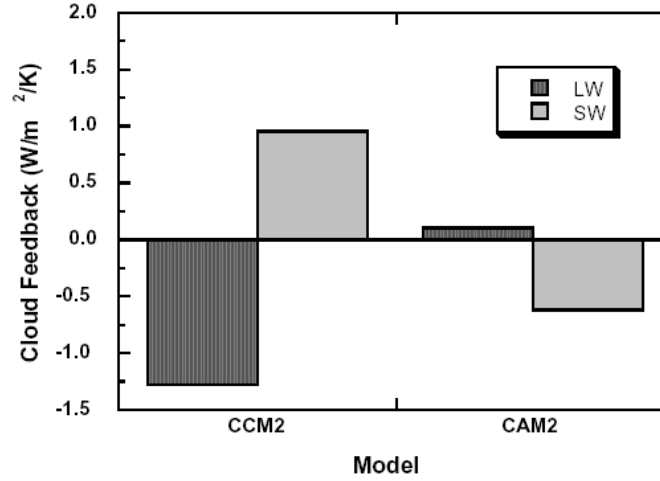


Figure 7. Feedbacks components separated into longwave (LW) and shortwave (SW) radiation in the CCM2 and in CAM2.

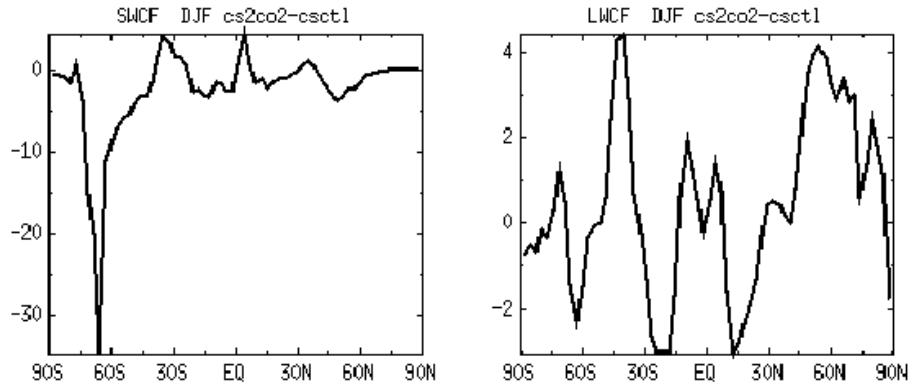


Figure 8. Change of the shortwave (left panel) and longwave (right panel) cloud forcing (W m^{-2}) in a climate change simulation for the DJF season.

Therefore, the negative cloud feedback in CCM2 is related with a dominant reduction of the cloud greenhouse effect associated with decreased cloudiness in the upper troposphere. The negative feedback in CAM2, however, is a result of the enhanced solar cooling effect from increased amount of low clouds.

5. Further Evaluation of Model Clouds

Given the above discussion, a reasonable question to ask is whether cloud feedbacks in current models have any fidelity to the real climate system. Since observations of cloud feedbacks are not directly available, a natural step is to evaluate model simulated clouds against available observations.

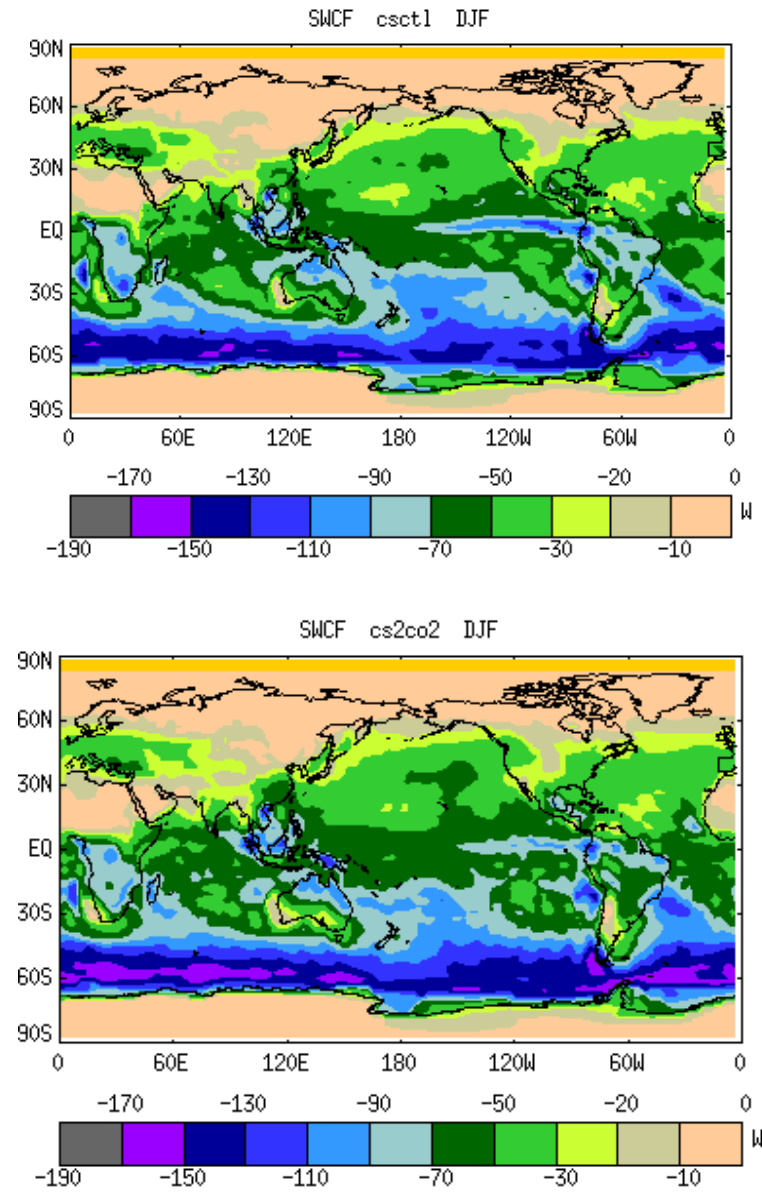


Figure 9. Geographical distribution of the January shortwave cloud forcing (W m^{-2}) in the control climate (upper panel) and warmer climate (lower panel).

A common practice in the climate modeling community is to use CRF measurements from ERBE as model validation datasets (e.g., Zhang et al. 2003). The same CRF at TOA, however, does not necessarily mean the same vertical structures of clouds, which is more directly related with cloud changes in response to a climate forcing. It is therefore desirable to use as many independent measurements as possible.

We present a comparison of simulated clouds in the CAM2 with measurements from the International Satellite Cloud Climatology Project (ISCCP) (Rossow and Schiffer 1991; Rossow et al. 1996). In ISCCP, geostationary satellite measurements of narrow band radiances are collected from around the world. A visible channel is used to derive the cloud optical thickness, and an infrared channel is used to derive the cloud top temperature and thus pressure. Clouds are then categorized according to their optical thickness and cloud top heights. To facilitate the comparison of model fields with ISCCP clouds, we implemented an ISCCP simulator in the CAM2, which was developed by Drs. Steve Klein at GFDL and Mark Webb at UKMO. The ISCCP simulator allows us to retrieve the same diagnostics from the CAM2 as what are available in ISCCP. The implementation method and discussions on the limitations of ISCCP can be found in Lin and Zhang (2003).

We only point out the main model deficiencies here and refer the reader to Lin and Zhang (2003) for more detailed comparison between CAM2 and ISCCP data. The three panels on the left in Figure 10 show the ISCCP low cloud amount in January 1983 categorized according to their optical thickness, with decreasing order from the top panel to the bottom panel. They are assigned the common names of stratus (optically thick), stratocumulus, and shallow cumulus (optically thin). The three panels on the right in Figure 10 show the corresponding diagnostics from the ISCCP simulator in the CAM2. Low clouds are defined as those with tops from the surface to 700 mb. The model overestimated optically thick low clouds, especially near 60°S, a region that has been highlighted in Figure 9 to describe the model cloud feedback. This bias is due to the model stratiform cloud scheme that tries to do the job of producing clouds associated with the boundary layer physics. The model also underestimated the optically thin and intermediate low clouds. This is because shallow convection does not directly produce clouds in the model.

Figure 11 shows the comparison of middle clouds in observations and in the model. They have cloud tops between 700 mb and 400 mb. The three types of middle clouds are given the names of nimbostratus (optically thick, top panels), altostratus, and altocumulus (optically thin, bottom panels). The model significantly underestimated middle clouds of thin and intermediate optical thickness. Again, the main cause is related with the arbitrary decoupling of model convection with low and middle clouds.

Figure 12 shows the three high-top clouds of deep convective clouds, cirrus, and thin cirrus, from the top panel to the bottom panel, in both ISCCP and in the CAM2. The model overestimated both optically thick and optically thin high clouds. The overestimation of high thin cirrus is caused by an inaccurate algorithmic relationship between the detrainment convective mass flux and cirrus anvils (Rasch and Kristjansson 1998). The overestimation of optically thick high clouds is likely due to biases in the model microphysical cloud scheme associated with the conversion rate of cloud droplets to precipitation.

These multiple opposing errors in high and middle clouds jointly produce a reasonable longwave cloud forcing at TOA. At the meanwhile, the opposing errors in optically thick and thin/intermediate clouds offset to produce a reasonable shortwave cloud radiative forcing at the TOA.

It is unlikely that a model can produce a reliable cloud variation and thus cloud feedback if its basic state is problematic. The overestimation of low stratus in the CAM2 has direct bearings on the negative feedback in the model. At the writing of this paper, the CAM2 is undergone significant revisions in its cloud scheme to improve the above-mentioned biases, so that it can be more faithfully used to project the sensitivity of the climate system for use in the next IPCC report scheduled for the year 2005.

We are therefore still far from getting a confident cloud feedback in climate models. Constructive process-oriented analysis is needed to understand the physical causes of model deficiencies. Figure 13 is used to illustrate this point. It shows two snapshots of one day apart of the observed surface pressure field, 500 mb height, and the infrared cloud image over North America on September 26 and September 27, 2002 respectively.

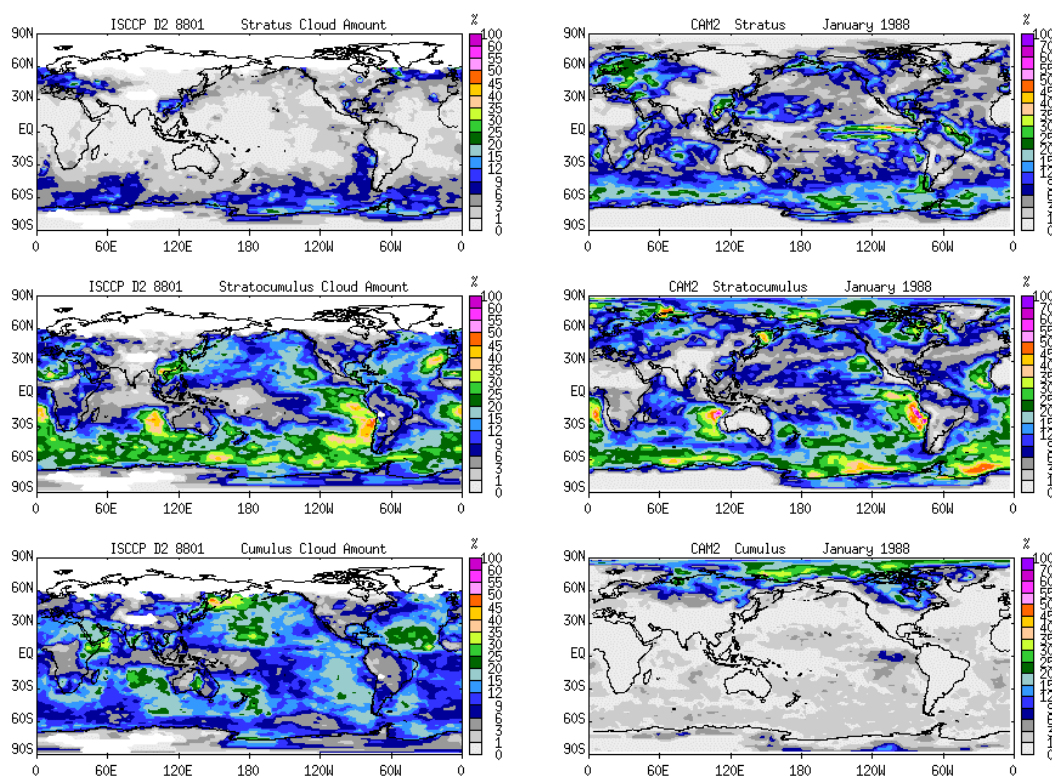


Figure 10. Low clouds in January 1988. Left panels: ISCCP observations with optical thickness decreasing from the top panel to the bottom panel. Right panels: CAM2 simulations.

Figure 14a shows the 12-hour CAM2 forecast corresponding to Figure 13a when it is initialized from the operational analysis. Plotted in the image is the model high cloud amount. The model is able to capture the cloud band associated with a hurricane. But the model overestimates high clouds in low latitudes, consistent with the ISCCP comparison in Figure 12. This confirms an algorithmic cause of error in the parameterization of high cloud

amount since the model thermodynamic fields are close to observations and yet the high clouds are not.

A completely different inference can be made, however, for the model cloud behavior in the 36 hour forecast shown in Figure 14b. When compared with Figures 13b, aside from the high cloud deficiencies at low latitudes, model clouds are very different from observations over the United States. This is because the hurricane in the model does not move northward as in observation. This failure is mainly the result of model resolution rather than initialization errors. Thus, the deficiency in simulated clouds is a reflection of errors in the model dynamics rather than in its cloud parameterizations. Any tuning of the cloud scheme to match simulated clouds with observed clouds could deteriorate the model rather than improve it. Correct simulation of clouds in this case thus requires a deeper effort in improving other aspects of the model.

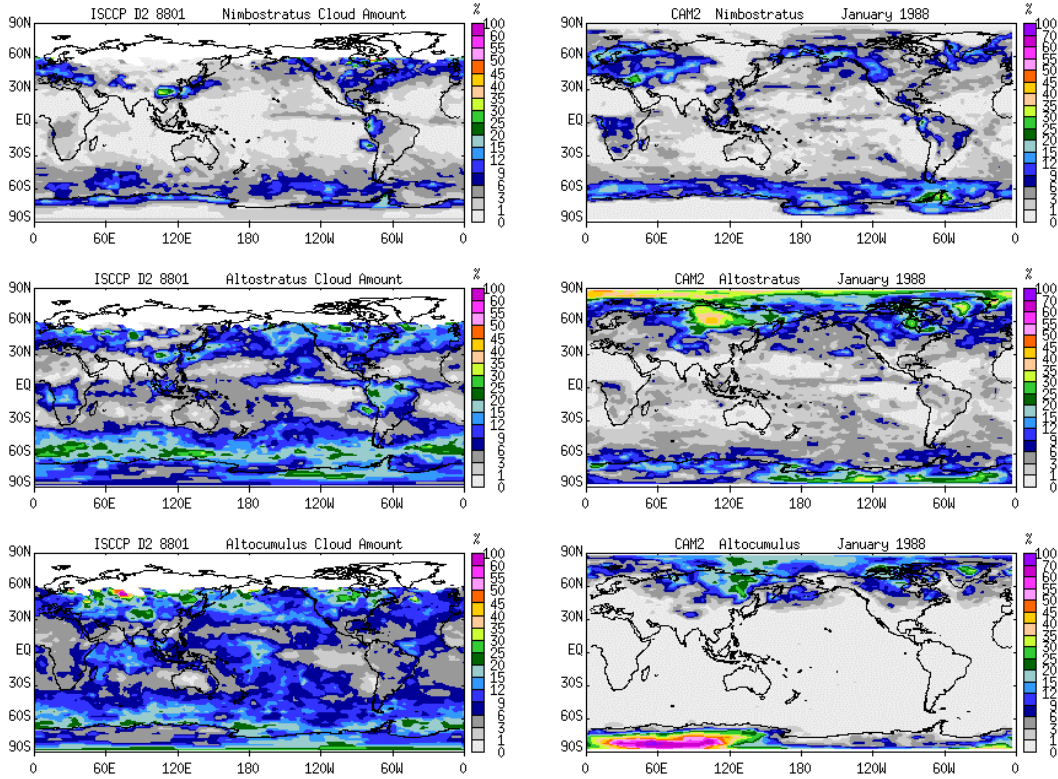


Figure 11. Same as Figure 10 except for middle clouds.

6. Summary and Discussions

Clouds are an integral part of the moist geophysical dynamics. They are strongly coupled with both grid scale and sub-grid scale processes. Cloud feedback is a result of aggregated change of cloud radiative forcing associated with chaotic transient atmospheric circulation.

This paper reviewed the concepts of cloud feedback and its role in defining the sensitivity of a climate model. It also discussed methods that have been used to diagnose the cloud feedbacks in climate models. I have attempted to present what cloud feedbacks are in current models, what processes caused the cloud feedbacks, and how model clouds compare with observations. From the discussions presented, it can be concluded that we are still far away from confidently simulating model clouds and their climate feedbacks.

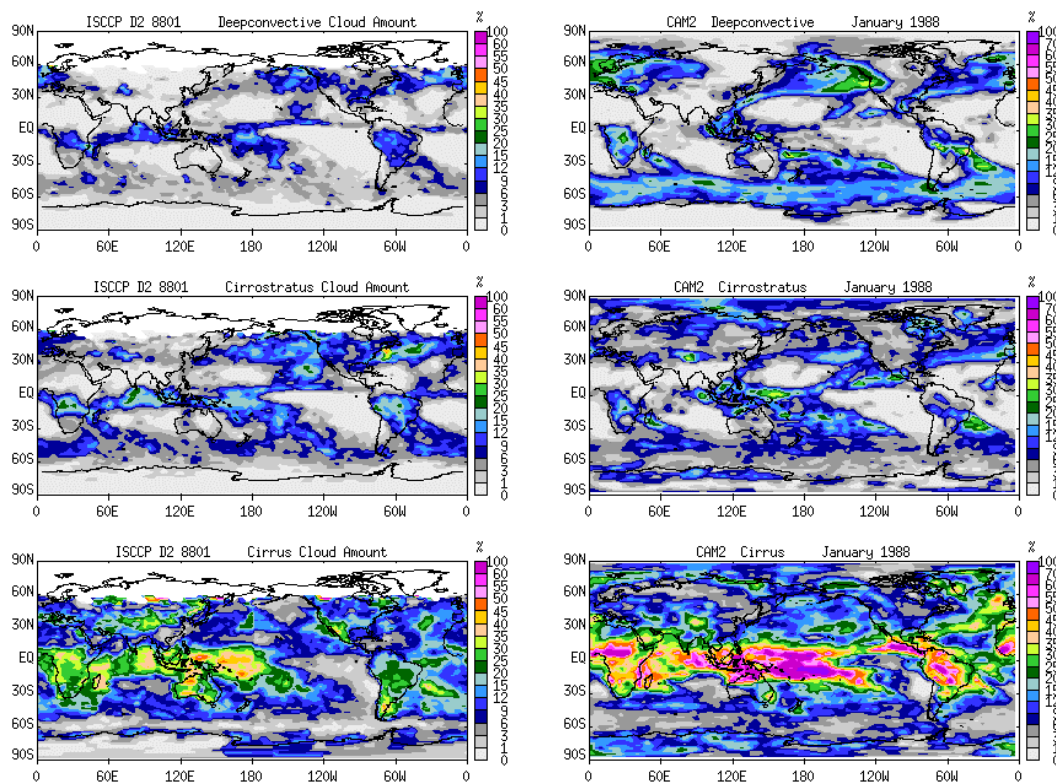


Figure 12. Same as Figure 10 except for high clouds.

The cloud feedback problem could become at least conceptually more tractable if we divide it into three different tasks. One is the microphysical cloud calculation with given dynamical circulation features. With detailed knowledge of aerosol distribution, the dynamical circulation provides the information on the generation of super saturation. This can then be used to estimate the number of cloud droplets nucleated, as in the simplified method of Ghan and Easter (1992) or more elaborate calculation using CCN spectral information as in Kogan (1991). The spectral size distribution of cloud droplets can then be calculated. This procedure itself incurs considerable demand on computational resource. This is one direction several modeling groups are currently pursuing.

The second aspect is the specification of the atmospheric dynamics on the subgrid scale. Clouds are mostly generated by subgrid scale processes, which have to be parameterized in climate models. To describe the variability of the atmospheric thermodynamic and dynamical structures within a grid, statistical description is needed. Yet, these statistical relationships should be based on realistic physical principles. At the present time, these subgrid models are either from empirical relationships or from highly intuitive conceptualizations. This deficiency in the subgrid scale dynamics has prompted Randall et al. (2003b) to use cloud resolving models inside a climate model to replace the parameterization package.

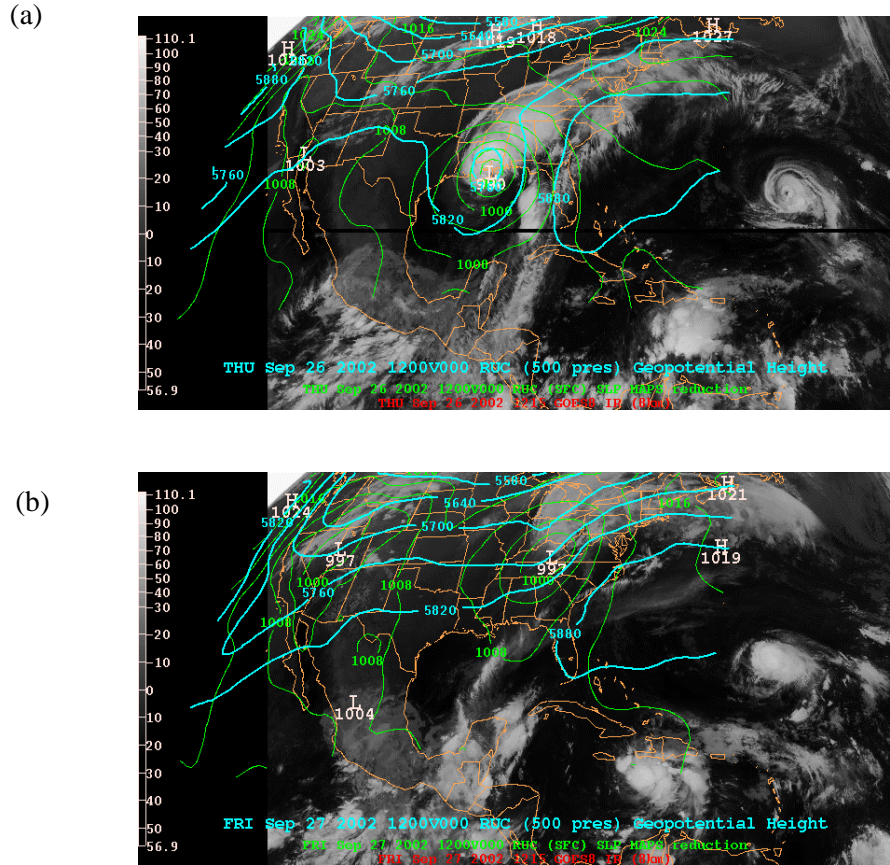
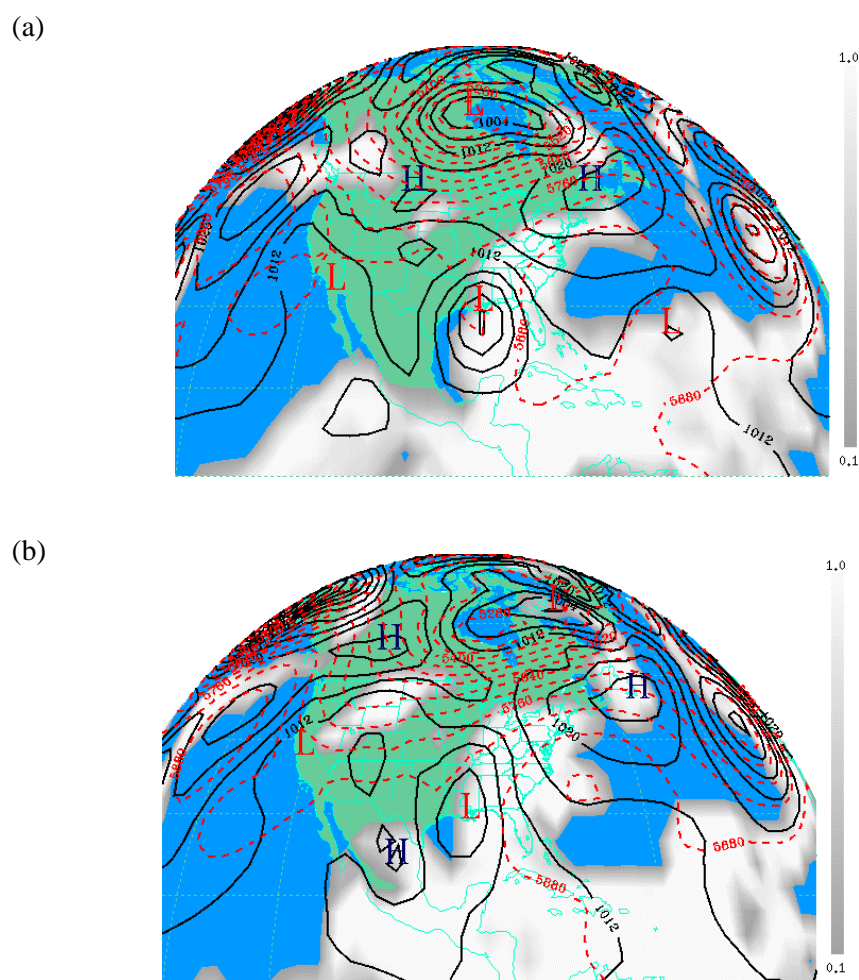


Figure 13. Observations of infrared clouds, 500mb height (cyan), and surface pressure (green). (a) 9/26/2002 12:00 GMT. (b) 9/27/2002 12:00 GMT.

The third issue is the abstraction of the coupling of subgrid scale dynamics with the subgrid scale cloud processes into a practical parameterization formulation. It is not clear whether this abstraction is possible. This issue is still valid even if spatial resolution of current models is reduced by an order of magnitude. On the practical side, very few existing convective schemes even include a component of the cloud microphysics.

The study of cloud-climate feedback is therefore rooted in the subgrid-scale dynamics and physics. This problem poses challenges and opportunities for constructive use of observations and calls for a breakthrough of parameterization methodology. Recent coordinated research activities in this regard include GCSS (GEWEX Cloud System Studies) (Randall et al., 2003a) and the DOE ARM (Atmospheric Radiation Measurement) program (Xie et al., 2002; Xu et al., 2002). Several subsequent papers in this volume describe specific examples of issues involved in the parameterization of subgrid scale physics and dynamics. The benefit of these activities will not just be to resolve the cloud-climate feedback uncertainty, but also to improve numerical modeling and numerical prediction of weather and climate in general.



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