The Sensitivity of the Tropical-Mean Radiation Budget

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(Manuscript received 15 July 2004, in final form 5 December 2004)

ABSTRACT

A key disagreement exists between global climate model (GCM) simulations and satellite observations of the decadal variability in the tropical-mean radiation budget. Measurements from the Earth Radiation Budget Experiment (ERBE) over the period 1984–2001 indicate a trend of increasing longwave emission and decreasing shortwave reflection that no GCM can currently reproduce. Motivated by these results, a series of model sensitivity experiments is performed to investigate hypotheses that have been advanced to explain this discrepancy. Specifically, the extent to which a strengthening of the Hadley circulation or a change in convective precipitation efficiency can alter the tropical-mean radiation budget is assessed. Results from both model sensitivity experiments and an empirical analysis of ERBE observations suggest that the tropical-mean radiation budget is remarkably insensitive to changes in the tropical circulation. The empirical estimate suggests that it would require at least a doubling in strength of the Hadley circulation in order to generate the observed decadal radiative flux changes. In contrast, rather small changes in a model's convective precipitation efficiency can generate changes comparable to those observed, provided that the precipitation efficiency lies near the upper end of its possible range. If, however, the precipitation efficiency of tropical convective systems is more moderate, the model experiments suggest that the climate would be rather insensitive to changes in its value. Further observations are necessary to constrain the potential effects of microphysics on the top-of-atmosphere radiation budget.

1. Introduction

It is often assumed that on large spatial and temporal scales, the balance between reflected shortwave (SW) and outgoing longwave radiation (OLR) is highly damped. Recent satellite measurements, however, have revealed unexpectedly large, decadal-scale variations in these quantities. Wielicki et al. (2002) compiled data from the last two decades of satellite measurements of radiative fluxes from both the Earth Radiation Budget Experiment (ERBE), and the Clouds and the Earth's Radiant Energy System (CERES) instruments. Averaging over the Tropics, this collection of measurements showed that the OLR has increased by $\sim$5 W m^{-2} over the period from the mid-1980s to the late 1990s and that the reflected SW has decreased by about 2 W m^{-2} over the same period. Both OLR and SW changes occurred primarily in cloudy skies. More recently, it was recognized that some of the trend is related to a slight drift in the altitude of the satellite (Wong et al. 2004). This can accurately be corrected by a simple scaling of the data by a known factor over time, and the resulting time series is shown in Fig. 1. While the magnitude of the OLR signal is slightly decreased relative to what was shown in Wielicki et al. (2002), the reflected SW signal is increased, which leads to a positive trend in the net radiative gain of the Tropics.

There are currently two different explanations that have been suggested for the observed trends: one involves changes in the tropical atmospheric circulation and the other cloud microphysics. Chen et al. (2002) analyzed upper-tropospheric humidity, cloud amount, surface air temperature, and National Centers for Environmental Prediction (NCEP) reanalysis vertical velocities. They found that the decadal trend in tropical

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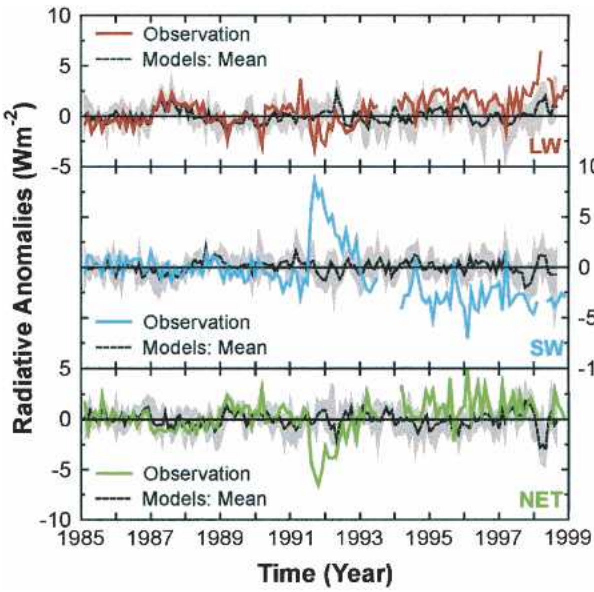

FIG. 1. Comparison of the observed OLR and reflected SW anomalies for the Tropics with climate model simulations. The red, blue, and green lines are the satellite records of OLR, reflected SW, and net TOA radiation, respectively, averaged from 20°S to 20°N. The model results are shown by the gray band and the black dashed line. The dashed line shows the mean of all five models, and the gray band shows the total range of model anomalies (max to min). This figure is a modified version of that shown in Wielicki et al. (2002), but using the radiative fluxes that were corrected based on Wong et al. (2004).

mean OLR is strongly correlated with moistening and increased cloud cover in convective regions of the Tropics and drying and decreased cloudiness in the subsiding regions. Those authors interpreted this as a signature of a strengthening of the tropical atmospheric circulation and suggested on the basis of the temporal correlation that an increase in circulation strength could explain the trend in tropical-mean radiative fluxes. Wielicki et al. (2002), on the other hand, proposed that the upward trend in OLR and reflected SW could be related to a decrease in tropical-mean cloudiness, though the authors did not suggest how or why this could occur. Wang et al. (2002) analyzed Stratospheric Aerosol and Gas Experiment II (SAGE II) measurements of uppermost opaque cloud occurrence frequency over the period 1985–98 and found that average cloud-top heights have lowered, which they pointed out could also explain the observed trends. Cess and Udelhofen (2003) provided further evidence that changes in cloud cover are responsible for the trend by computing variations in the atmospheric greenhouse parameter, $G = 1 - \mathrm{OLR}/\sigma T_s^4$. Increasing greenhouse gas concentrations increases the value of G, yet Cess and Udelhofen (2003) find that the observed value of G has decreased over the 1985–99 period. They suggest that this decrease in G can be explained by a decrease in the annual-mean tropical average cloud fraction, which they document with International Satellite Cloud Climatology Project (ISCCP) measurements.

The ability of global climate models (GCMs) to reproduce the observed variations in the top-of-atmosphere (TOA) radiation budget provides an important test of the models. Wielicki et al. (2002) compared the observed tropical-mean TOA fluxes with that predicted from several different models run with observed SSTs. In Fig. 1, the updated fluxes are also compared with the model simulations. While the magnitude of the observed OLR signal is slightly smaller, both the OLR and SW signals fall outside the model range. Cess and Udelhofen (2003) analyzed a simulation with the National Center for Atmospheric Research (NCAR) coupled model [Community Climate System Model (CCSM1)] in which the greenhouse gases and sulfate aerosols were gradually increased over the period 1870–1998. They compared the tropical mean OLR and SW over the last 29 yr of the run and found that not only are model-simulated changes small (<1 W m^{-2}), but the trend in OLR was actually of the opposite sign as the observations. Thus, whether they are given the observed SSTs or the known climate forcings, climate models are not currently able to reproduce the observed trends in tropical-mean radiative fluxes of the last two decades.

In this paper, we take a different approach to investigating the discrepancy between observed and model-simulated trends in TOA fluxes. Rather than imposing a set of observed forcings in a model (such as SST or atmospheric trace gases), we assess the viability of the two competing hypotheses that have been proposed to explain the observed changes: 1) an increase in the tropical circulation and 2) a change in cloud microphysics. We do not attempt to explain how such changes might have occurred or prove that they did occur. Instead we take a more tractable approach to this problem by asking if the changes did occur in a model, could the model then reproduce the observed changes in TOA fluxes. Specifically, we perform a series of model sensitivity experiments in which the strength of tropical circulation and convective precipitation efficiency are perturbed, and then we assess the ability of the perturbed models to simulate the observed changes in the tropical mean radiation budget.

The remainder of the paper is organized as follows. In sections 2 and 3 we investigate the sensitivity of the tropical-mean radiation budget to changes in the Hadley circulation using GCMs and satellite observations.

TABLE 1. The change in various quantities for NOFLUX minus the control run in four GCMs. Hadley cell changes are shown as a percent change of the max (Northern Hemisphere) and min (Southern Hemisphere) of the zonal mean meridional streamfunction. Radiative fluxes are shown in W m^{-2} and are averaged over the Tropics (30°S–30°N). Bold indicates total sky OLR and SW. Results are similar for averages over 20°S–20°N (as in Wielicki et al. 2002 and Fig. 1); however, here we choose 30°S–30°N because it covers the extent of the Hadley cell. OLR is defined as positive upward. Shortwave fluxes are shown as reflected shortwave, and so they are positive upward. Separation into clear- and cloudy-sky fluxes are not shown for the GISS model because they were not saved. Cloud changes are shown in change in percent. The "dynamic cloud radiative forcing" is estimated as in Bony et al. (2004) as described in the text.

	GISS	CCM	MCM	AM2
Horizontal resolution (latitude × longitude in degrees)	4 × 5	2.8 × 2.8	2.2 × 3.75	2 × 2.5
Number of vertical levels	9	18	17	17
ΔHadley max (NH) (percent)	+93%	+105%	+98%	+91%
ΔHadley min (SH) (percent)	+35%	+46%	+58%	+79%
ΔOLR	**+3.3**	**+3.8**	**+5.8**	**−2.5**
ΔOLR cloudy sky	N/A	+4.0	+1.7	+0.5
ΔOLR clear sky	N/A	−0.2	+4.1	−3.0
ΔSW (reflected)	**+0.7**	**+1.6**	**−6.9**	**+6.2**
ΔSW cloudy sky	N/A	+1.4	−8.3	+4.9
ΔSW clear sky	N/A	+0.2	+1.4	+1.3
ΔTotal cloud	−1.4	−1.5	−4.5	+2.7
ΔLow cloud	−0.9*	+1.1	−5.5	+7.4
ΔHigh cloud	−1.5**	−2.0	−0.6	−1.5
Dynamic cloud radiative forcing: ΔOLR	+0.6	+2.4	+0.5	+1.6
Dynamic cloud radiative forcing: reflected ΔSW	0.0	−1.6	−0.4	−0.7

* Layers 1, 2, and 3.
** Layers 6 and 7.

In section 4 we assess the sensitivity to changes in convective precipitation efficiency, and then in section 5 discuss the results and their implications.

2. Circulation controls on TOA fluxes

Chen et al. (2002) suggested that the decadal trend in tropical-mean radiative fluxes is related to a strengthening of the Hadley/Walker circulation. Their motivation was the strong temporal correlation between the observed radiative fluxes and various indicators of the circulation strength. The physical linkage between the strength of the circulation and the tropical-mean TOA fluxes, however, has not been demonstrated. Atmospheric GCMs are a viable tool with which to test this physical linkage. However, because the strength of the circulation is determined interactively in a GCM, it is difficult to vary in a controlled manner.

Clement and Seager (1999) found a convenient way to alter the strength of the Hadley circulation by changing the effective heat transport by the ocean. Atmospheric GCMs forced with a fixed SST climatology have an equilibrium net surface heating that can be interpreted in large part as the effect of dynamical transport of heat in the ocean. It is common practice to use this implied ocean heat transport (OHT) when the atmospheric GCM is coupled to a mixed layer ocean so that the resultant SST is close to climatology. While the implied OHT really contains both the effects of the ocean and model errors, it is similar in spatial structure to the best estimates of the actual ocean heat transport (Russell et al. 1985; Trenberth and Caron 2001). If a model is then run with the OHT set to zero, the atmospheric heat transport increases to compensate in part by increasing the strength of the Hadley circulation.

Here we take advantage of this model arrangement in order to test the response of the TOA fluxes to changes in the strength of the Hadley circulation. Recognizing that the radiation will depend significantly on the physical parameterizations in the model, we use four different atmospheric GCMs that contain a wide range of atmospheric physics. The models are 1) the Goddard Institute for Space Studies (GISS) Model II′ (Hansen et al. 1983), 2) the Community Climate Model (CCM) version 3.10 (Kiehl et al.1998), 3) the Geophysical Fluid Dynamics Laboratory (GFDL) Manabe Climate Model (Delworth et al. 2002), and 4) a new version of the GFDL model [Atmosphere Model 2 (AM2) version p3, hereafter referred to as AM2]. The version of the GFDL model that we use here is slightly different than that documented in GFDL Global Atmospheric Model Development Team (2004). Experiments performed here have been repeated with both versions of the model, and the results are consistent, but we only show results from AM2 because those experiments have been previously published and analyzed by Winton (2003).

Table 1 gives the spatial resolution of each model, and we will include some discussion of the details of the models that are relevant for understanding the TOA fluxes. However, we refer the reader to the references cited for complete details. All of these experiments have been previously published in studies of how the ocean heat transport affects the mean climate (Clement and Seager 1999; Seager et al. 2002; Winton 2003; Herwiejer et al. 2004). Here we use these experiments simply as analogs for strong and weak Hadley circulations, and the reader is referred to those studies for further discussion of the adjustment of the global climate.

Each model has a control run with a realistic SST climatology (by design), which is compared with the same model run with an ocean mixed layer in which the OHT is set to zero everywhere (hereafter referred to as NOFLUX). While this change in OHT is rather extreme, it allows for an experiment that can be made in a consistent way across models. Table 1 summarizes the results for the four models. In each model, the Hadley cell strength, as measured by the maximum (for the Northern Hemisphere) and minimum (for the Southern Hemisphere) of the zonal mean meridional streamfunction, increases dramatically. The Northern Hemisphere cells increase approximately by a factor of 2, while the Southern Hemisphere cells increase by less. The smaller response in the Southern Hemisphere is related to the fact that the zonal mean OHT is smaller there, though it has been noted that uncertainties in the estimates of OHT in the Southern Hemisphere are considerably larger (Trenberth and Caron 2001).

Figure 2 shows the zonal mean streamfunction for AM2 in both the control and NOFLUX runs. This pattern is consistent across all four models: The circulation intensifies, and the maximum of the streamfunction moves equatorward. In addition, there is a sizeable drying of the subtropics as shown in Fig. 2c. Herweijer et al. (2004) explain that this drying is due to a dynamical water vapor feedback. In the NOFLUX case, the convecting region of the atmosphere is confined to the equator. There the atmosphere is close to saturation so the moisture flux is closely balanced by precipitation. In contrast, for the control run, the convection is spread over a broader meridional extent and occurs in regions where the relative humidity is at more intermediate values. There vapor can accumulate without getting close to saturation and being removed by precipitation, resulting in a moister subtropical free troposphere.

These changes in atmospheric circulation and structure result in large redistributions of the TOA radiative fluxes (Fig. 3). In all models, there is a decrease in the OLR and an increase in the reflected SW on the equator. This is primarily due to an increase of high cloud cover there because of the equatorward shift of convection. In the subtropics, there is a decrease in high cloud cover with opposite effects on the TOA fluxes. This change is somewhat analogous to the observed trends in TOA fluxes shown in Chen et al. (2002) in that it is a change in the extremes. In the subtropics, where OLR is relatively high and reflected SW is low, there is an increase (decrease) in OLR (SW), and in the deep Tropics, a region of relatively low OLR and high reflected SW, the OLR is reduced and reflected SW increases. Allan and Slingo (2002) showed a similar spatial pattern with an increase in OLR of several watts per squared meters in the off-equatorial regions and a comparable decrease on the equator. We note, however, that while the radiative flux changes in the model may have a somewhat analogous pattern (at least in a zonal mean sense) to the observations, the amplitude of the changes in the model is at least an order of magnitude higher.

In spite of these rather extreme changes in circulation and regional TOA fluxes, the tropical mean OLR and reflected SW changes are small and vary in sign from model to model (Table 1). This scatter is due to the different behavior of clouds in the models. To show this, let us first consider the two models that have strong low-cloud feedbacks: Manabe Climate Model (MCM) and AM2. In both of these models, the change in reflected SW is compensated for by an OLR change that occurs primarily in the clear sky (Table 1). The SW signal is associated with a change in the low-cloud cover (Fig. 4). Low clouds have little effect on OLR because the cloud-top temperature is not significantly different from the surface temperature (Hartmann et al. 1992). Rather, the change in reflected shortwave is balanced in part by a change in temperature that affects the clear-sky OLR via Plank's law and in part by changes in the humidity distribution. In AM2, a stronger Hadley cell is associated with an increase in the low-cloud cover that occurs throughout most of the Tropics (Fig. 4). Because the reflected SW is increased, there is a decrease in temperature that results in a *decreased* clear-sky OLR in that model. The change in humidity (Fig. 2c) also effects the clear-sky OLR but in the opposite sense: As explained in Herweijer et al. (2004), the dynamically induced drying in the subtropics increases the OLR, but that does not overwhelm the effect of the temperature change in a tropical-mean sense. In contrast, in MCM the low cloud cover decreases under a stronger Hadley cell, and there is an associated increase in temperature and an increase in clear-sky OLR. The stronger Hadley cell still produces a strong subtropical drying in this model, so the temperature and humidity

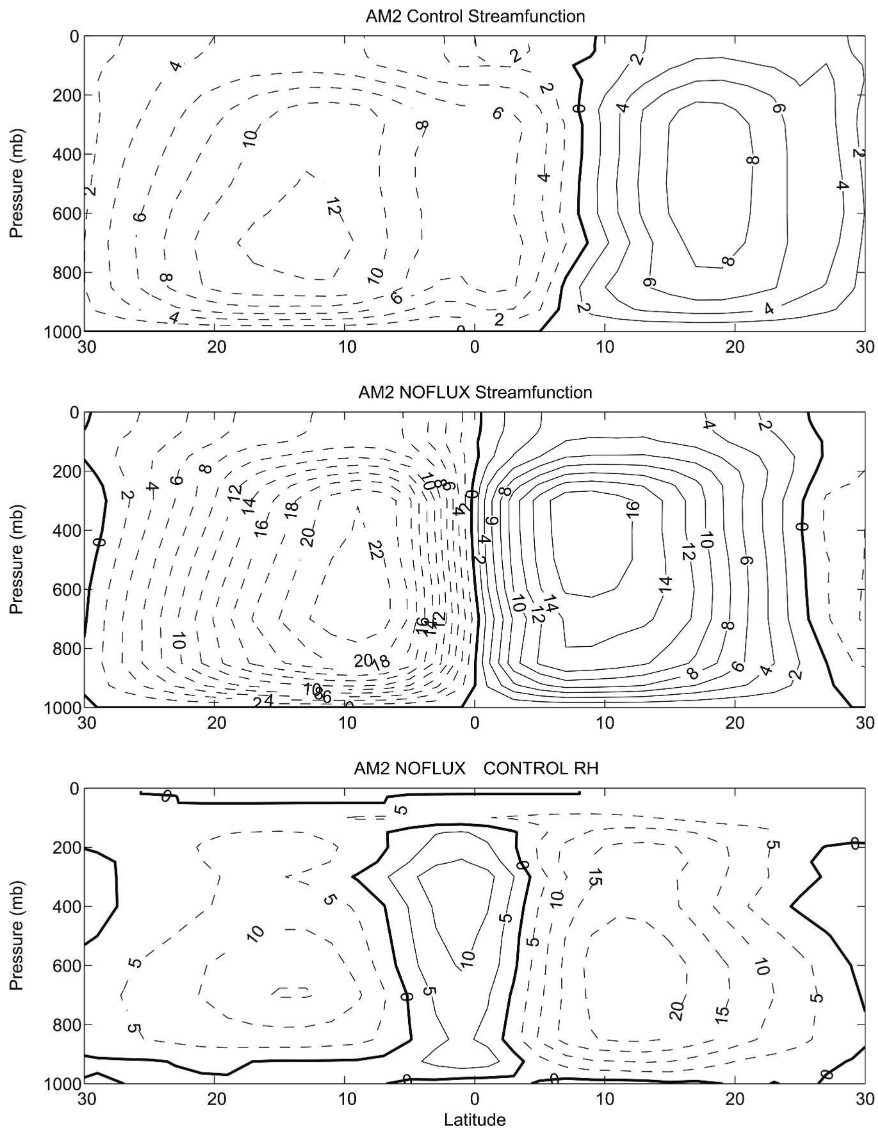

Fig. 2. Zonal mean meridional mass streamfunction (10^{10} kg s^{-1}) for (top) the control run of AM2 and (middle) the NOFLUX run of AM2. (bottom) The difference in zonal mean relative humidity (in percent) between NOFLUX and the control for AM2.

changes both increase the OLR, resulting in a larger change than in AM2.

While both of these models have a significant low-cloud feedback, only AM2 behaves in a way that is consistent with the observed behavior of tropical/subtropical marine stratus. In this model, when the lower-tropospheric static stability increases, the low-cloud fraction increases (Winton 2003) as in observations (Klein and Hartmann 1993). The free tropospheric potential temperature of the Tropics is tied to the surface moist static energy of the convecting regions (Betts and Ridgway 1989; Pierrehumbert 1995; Miller 1997; Clement and Seager 1999). In the NOFLUX run, the sea surface temperature of the convecting regions goes up because these are regions that are cooled by export of heat by Ekman divergence in the ocean. As the surface temperature increases, the moist static energy increases by more because of the effect of additional moisture in the boundary layer (Betts and Ridgway 1989). Hence, the lower-tropospheric stability increases in most of the Tropics, and the low-cloud fraction increases in AM2. In a similar manner, the decrease in low-cloud cover in the eastern ocean basins is due to the strong warming that occurs in the absence of the cooling effect of ocean upwelling. However, this region has little effect on the tropical mean. We emphasize, however, that while AM2 may have realistic low-cloud behavior, an increase in the circulation

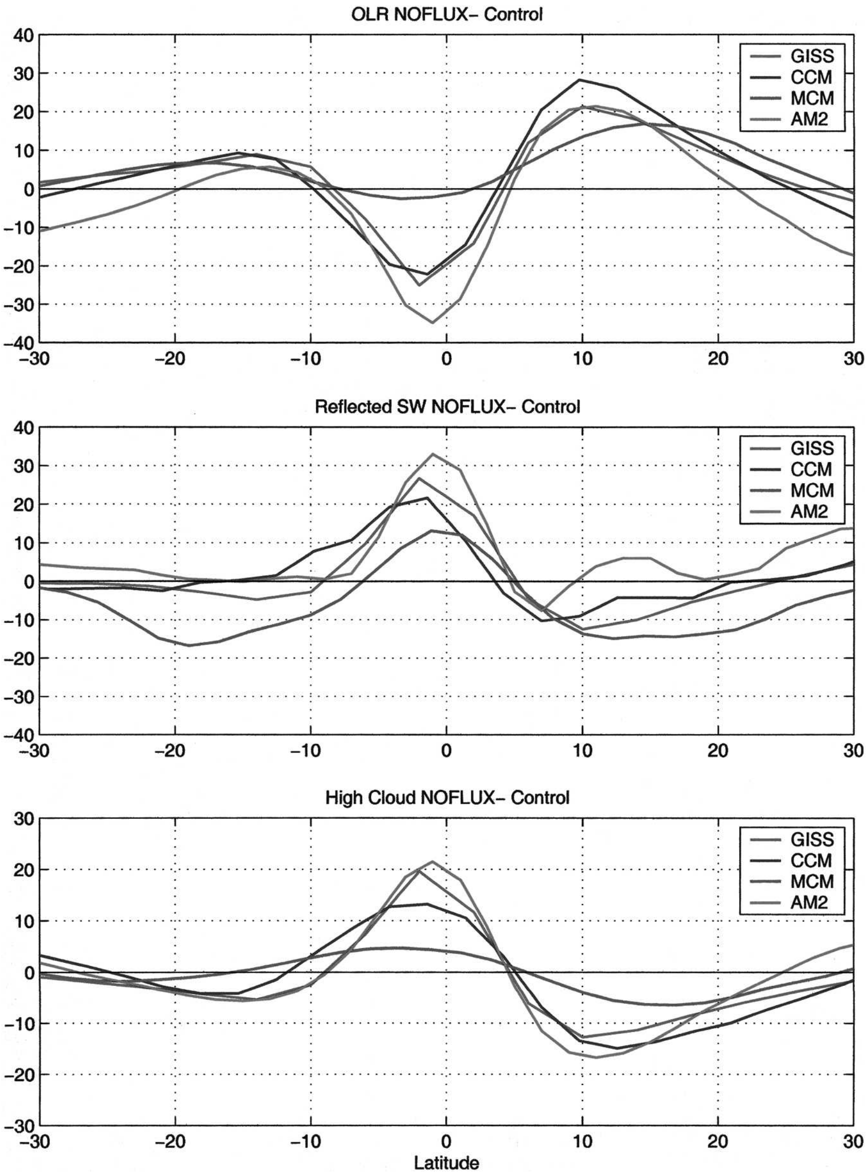

FIG. 3. Zonal mean change between the NOFLUX and the control run in (top) OLR in W m^{-2}, (middle) reflected SW in W m^{-2}, and (bottom) high-cloud cover in percent for all four models.

strength leads to an *increase* in reflected SW and a *decrease* in the OLR, which is opposite in sign to the observations shown in Fig. 1.

Low-cloud cover changes in the GISS and CCM models are small both regionally (Fig. 4) and in the tropical mean (Table 1). These models significantly underestimate the climatological low-cloud fields (Del Genio et al. 1996; Kiehl et al. 1998), so it is not surprising that the sensitivity of these clouds is also weak. In these models, it appears that the tropical-mean radiative flux change is controlled by the high clouds. There is a small net decrease in tropical-mean high cloud because the area where the cloud decreases (subsiding regions) is greater than where it increases (convecting regions), as is apparent in Fig. 3. For CCM this is consistent with the change in OLR, which is mainly a cloudy-sky signal. Clear and cloudy-sky fluxes were not saved in the GISS experiment, but the change in tropical-mean OLR, SW, and cloud cover are comparable with CCM, suggesting that the high clouds may be responsible for the change in total-sky OLR in GISS as well.

In summary, the effect of an increase in Hadley cell strength is to redistribute clouds and radiative fluxes,

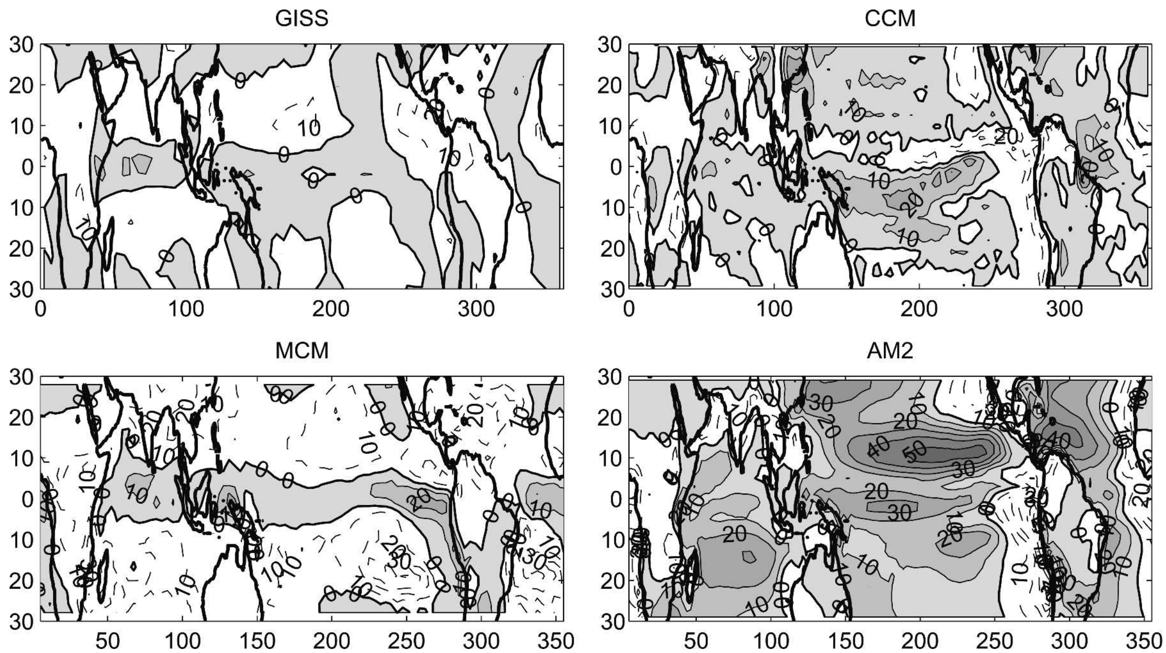

FIG. 4. Change in low-cloud cover (percent) between NOFLUX and the control run for all four models.

which results in sizeable regional changes but relatively small tropical-mean changes: it takes roughly a doubling of the strength of the Hadley cell to produce changes similar in magnitude to the trend observed by ERBE, a result that is similar in all models. The quantitative change in the tropical-mean radiative fluxes is determined by the response in the tropical-mean cloud fields, which is highly model dependent. In the models with large low-cloud response (MCM and AM2), the OLR change is in the clear sky, which is not consistent with the observations. Moreover, in the model with realistic low-cloud behavior, AM2, the sign of the response of OLR and SW to increased circulation strength is opposite to the observations (Fig. 1). In GISS and CCM, a stronger Hadley cell results in an increase in the tropical-mean OLR and reflected SW that is apparently related to high cloud cover, but, again, it takes an extreme alteration to the Hadley cell to produce such changes.

3. Empirical estimation of the effect of circulation changes on TOA fluxes

Cloud feedbacks in GCMs are known to be highly uncertain, and the experiments performed here are no exception: the change in tropical-mean cloud cover and resultant TOA fluxes vary not only in magnitude, but in sign from model to model for a very similar change in the circulation. Rather than rely on the model to predict the response of cloud forcing to a change in circulation, here we take advantage of the fact that the circulation changes in the models are robust enough to make an empirical estimate of the effect of an increase in Hadley cell strength on tropical-mean TOA fluxes. The work of Bony et al. (2004) provides a convenient framework in which to do this. They separate cloud feedbacks into a dynamic and a thermodynamic component by defining circulation regimes according to the vertical motions. A probability distribution function, P_ω, of the midtropospheric vertical velocity is defined such that

$$\int_{-\infty}^{+\infty} P_\omega \, d\omega = 1.$$

The distribution taken from AM2 is shown in Fig. 5a (and is similar for all four models). The peak at values of ω of about 20 hPa day^{-1} represents the large area of the Tropics where there is moderate subsidence. The tropical average cloud radiative forcing, $\overline{C}$, can be written as

$$\overline{C} = \int_{-\infty}^{+\infty} P_\omega C_\omega \, d\omega,$$

where C_ω is the cloud radiative forcing for a particular value of ω. Perturbations about the mean climate in the cloud radiative forcing can then be separated into different terms:

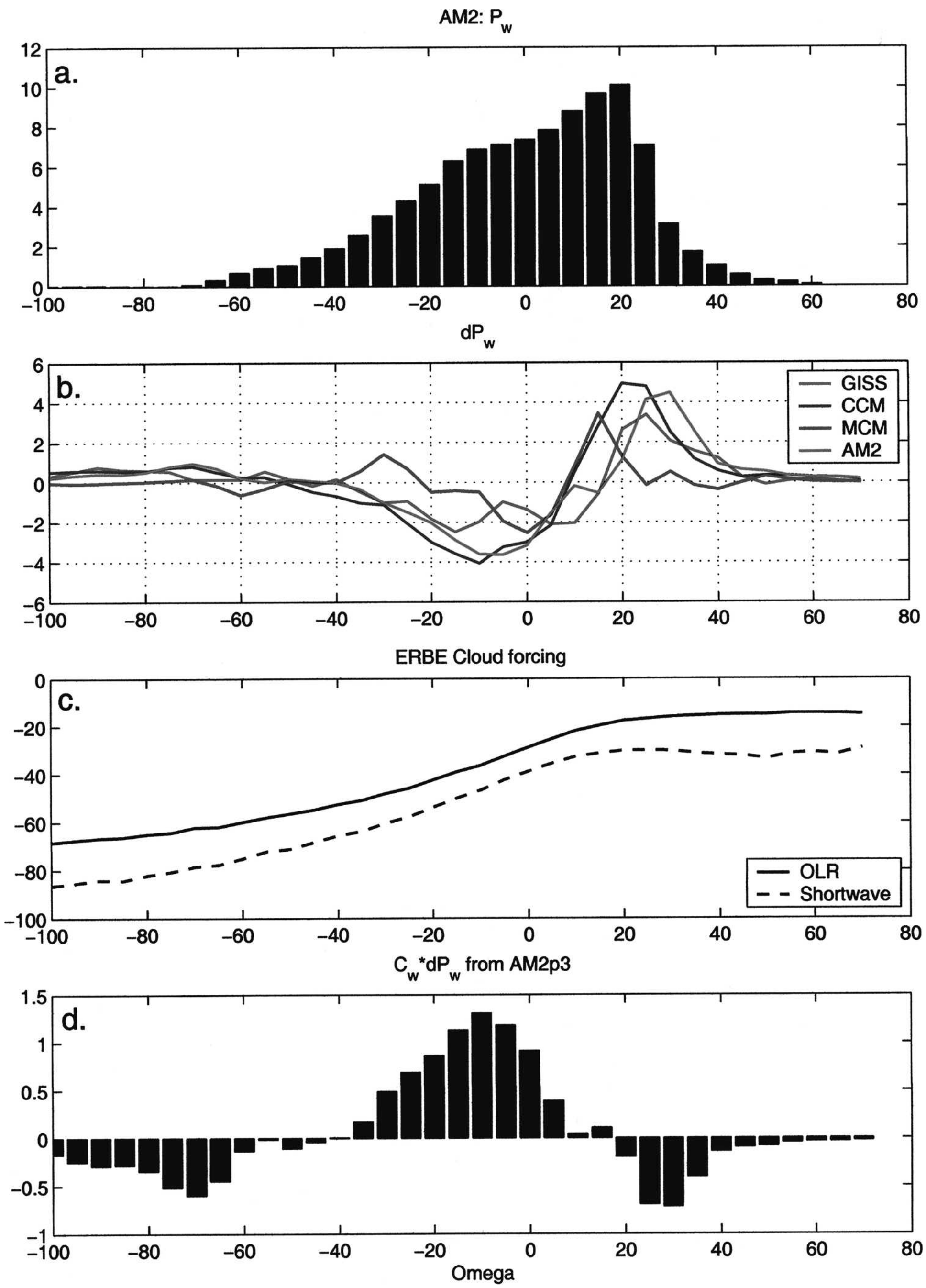

FIG. 5. (a) The probability distribution function, P_ω, of the annual-mean 500-hPa vertical velocity (hPa day^{-1}) for the control run for AM2. Values are binned by 10 hPa day^{-1} and normalized as described in the text. (b) The change in PDF, δP_ω, of the 500-hPa vertical velocity (hPa day^{-1}) between the NOF-LUX and the control run for all four models. (c) Shortwave and longwave cloud forcing, C_ω, in W m^{-2} as calculated in Bony et al. (2004). Values are binned using the NCEP reanalysis values for 500-hPa vertical velocity. Longwave cloud forcing is shown as OLR, so it is positive upward. The presence of clouds lowers the OLR, so values are negative. Shortwave cloud forcing is defined as absorbed SW, which is positive downward. The presence of clouds lowers the absorbed SW, so the values are also negative.

$$\delta c = \int_{-\infty}^{+\infty} C_\omega \delta P_\omega \, d\omega + \int_{-\infty}^{+\infty} P_\omega \delta C_\omega \, d\omega + \int_{-\infty}^{+\infty} \delta P_\omega \delta C_\omega \, d\omega.$$

The first term on the right-hand side is the cloud radiative forcing due to circulation changes, what Bony et al. (2004) call the "dynamic" component. The second is what they refer to as the "thermodynamic" component, and the third is a cross term that is small.

Here we are interested in the dynamic component. That is, how does a change in the Hadley cell strength affect the tropical-mean cloud forcing? The models provide a robust estimate of δP_ω for a strengthening of the Hadley cell, shown in Fig. 5b. There is a reduction in the moderate values of ascent (values between about -40 and 0), and the frequency of strong subsidence and strong ascent increase, as would be expected with an intensification of the circulation. Rather than use the model to estimate C_ω, an empirical estimate can be made using satellite observations. Bony et al. used ERBE data to calculate the shortwave and longwave components of C_ω, which are shown in Fig. 5c. Comparing the change in circulation (Fig. 5b) with the observed cloud forcing (Fig. 5c), it is clear that the change in circulation can produce an increase in the extremes (both high and low values) of OLR and SW cloud forcing. As discussed earlier, this result is consistent with the analysis of Chen et al. (2002). However, integration over all values of ω shows, again, that the large change in dynamics has relatively little effect on the tropical mean (Table 1). This empirical estimate is even smaller than that produced by the models. The analysis clarifies why this is the case, as demonstrated in Fig. 5d (for the longwave only, though the SW is analogous). There is a large increase in the subsidence where the longwave cloud forcing is relatively small (boundary layer clouds in subsiding regions have little OLR signal) and a small increase in the occurrence of large values of ascent where the longwave cloud forcing is large (deep clouds have a big OLR signal). Both of these decrease the OLR by a comparable amount. The decrease is compensated for by the reduction in modest values of ascent where the OLR signal is in the middle range. This clearing of the skies in regions of moderate ascent dominates and produces a small tropical-mean increase in OLR for all models.

To summarize, our results show that changes in the strength of the Hadley cell are fairly ineffective at altering the tropical-mean radiation balance. Whether we estimate this effect directly from the models or by combining model dynamics with observed cloud forcing, it takes a rather dramatic change in the strength of the Hadley cell to generate ERBE-like changes in the tropical-mean TOA fluxes. The empirical estimate, because it does not rely on uncertain cloud feedbacks in the models, is more robust and implies that even in the most sensitive model, it would take at least a doubling in the strength of the circulation to explain the observed decadal trend in tropical-mean radiative fluxes.

4. Microphysical controls on TOA fluxes

In this section, we perform another series of GCM sensitivity experiments to explore the viability of this second hypothesis—that changes in cloud microphysics could produce the observed trends in the tropical radiation budget. Previous studies have shown that the radiation budget in simple models is highly dependent upon the assumptions regarding the cloud microphysical processes (Renno et al. 1994; Emanuel and Pierrehumbert 1996). Chief among these is the convective precipitation efficiency, whose simplified representation in GCMs has raised concerns about their adequacy for climate change studies. Here we explore the sensitivity of the tropical-mean radiation budget in GCMs to changes in convective precipitation efficiency. As with the previous hypothesis, our intention here is not to explain the underlying cause of a supposed change in microphysical quantities. Rather we simply wish to determine whether or not a model with altered cloud microphysics is capable of reproducing the observed changes in the tropical radiation budget. We focus specifically on the role of convective precipitation efficiency in regulating the distribution of upper-level cirrus clouds because (i) previous studies have demonstrated a large sensitivity to this parameter in simple models (e.g., Renno et al. 1994) and (ii) cirrus clouds have roughly equal but opposing effects on the OLR and SW in the Tropics (Kiehl et al. 1998), which is consistent with the observed trend from the *Earth Radiation Budget Satellite* (*ERBS*).

Cirrus anvils are known to play a key role in regulating the radiative energy budget and climate of the Tropics (Ramanathan et al. 1983; Ramaswamy and Ramanathan 1989; Ramanathan and Collins 1991; Donner et al. 1997; Zender and Kiehl 1997). The spatial coverage, lifetime, and water content of cirrus anvils are largely determined by complex microphysical processes acting within convective updrafts that determine, among other things, the fraction of condensate that falls as precipitation, termed the "precipitation efficiency." However, because of the coarse resolution of GCMs, the representation of these microphysical processes in current cloud/convective parameterization schemes is a key source of uncertainty (Zender and Kiehl 1997), as are any cloud feedbacks that might be potentially associated with them (Renno et al. 1994).

Despite large uncertainties, changes in precipitation efficiency have also been invoked as a potential feedback in climate change. Lindzen et al. (2001) have argued for the presence of a negative feedback in the climate system by which the precipitation efficiency increases as the underlying surface warms, which leads to a reduction in high-cloud cover. We note that the interpretation of the observations used to support this hypothesis as well as the radiative characteristics of the clouds used to infer the climate feedback from the proposed changes in precipitation efficiently have been repeatedly called into question (Chambers et al. 2002; Fu et al. 2002; Hartmann and Michelsen 2002; Lin et al. 2002; Del Genio and Kovari 2002). Nevertheless, there is some observational evidence to suggest that precipitation efficiency in clouds can vary and is dependent on surface temperature. For example, recent measurements from TRMM (Lau and Wu 2003) suggest that the precipitation efficiency in low clouds increases as the surface temperature increases, although they could find no evidence of a similar behavior in deep convective clouds. On the other hand, Del Genio and Kovari (2002) found that the efficiency of precipitation in tropical convective systems, defined as the ratio of surface precipitation rate to precipitable water vapor, does increase with surface temperature.

It is important to reiterate, however, that our intention here is not to identify the mechanisms that could lead to a change in the convective precipitation efficiency or to verify that such changes have, in fact, occurred over the past two decades. Rather we quantify the sensitivity to this parameter and determine whether or not a model with altered cloud microphysics is capable of reproducing the observed changes in the tropical radiation budget.

GCM sensitivity studies

While the sensitivity of the tropical radiation budget to changes in precipitation efficiency has been clearly demonstrated in simple models (Renno et al. 1994; Emanuel and Pierrehumbert 1996), few studies have investigated such sensitivity in GCMs (Lohmann and Roeckner 1995; Jakob 2002). To address this issue, we performed a series of simulations with the GFDL AM2 in which the convective precipitation efficiency, an externally specified parameter in the model's convection scheme, was varied through a range of values. For each value of specified precipitation efficiency (PE) the model was integrated for a period of 5 yr using climatological SSTs.

Moist convection in the GFDL AM2 is parameterized using the Relaxed Arakawa–Schubert (RAS) formulation of Moorthi and Suarez (1992). In this scheme, convection is represented by a spectrum of entraining plumes that produce precipitation. Closure is determined by relaxing the cloud work function for each cloud in the spectrum back to a critical value over a fixed time scale. The cloud work function is defined by the integral over the cloud depth of the product of the mass flux and the buoyancy (which is proportional to the difference between the cloud virtual temperature and that of the grid-scale environment at the same height). The convective precipitation efficiency, defined as the fraction of water condensed in the cumulus updrafts that becomes precipitation, is specified to be 0.99 for deep convection and 0.5 for shallow convection for AM2. Deep convection is defined as updrafts that detrain at pressure levels above 500 hPa whereas shallow convection is defined as updrafts that detrain beneath 800 hPa. For pressures between 500 and 800 hPa, the precipitation efficiency is linearly interpolated in pressure between the values for deep and shallow convection. Complete details regarding the implementation of RAS in the GFDL AM2 can be found in GFDL Global Atmospheric Model Development Team (2004).

Figure 6 illustrates the difference in annual-mean OLR and reflected SW between two simulations of the GFDL AM2. The first uses a control value of 0.99 for the deep convective precipitation efficiency (indicating that 99% of the total convective condensate is converted into rainfall). In the second simulation, the PE is increased by 0.5% from 0.99 to 0.995. All other parameter settings, including shallow convective precipitation efficiency, are the same in both simulations. The simulations with the slightly higher value of convective precipitation efficiency detrain less cloud water in the upper troposphere, resulting in fewer high clouds. The reduction in high-cloud cover results in an increase in OLR (Fig. 6a) and a decrease in reflected SW radiation (Fig. 6b), both of which are largest over the Tropics. The opposing nature of the changes in OLR and reflected SW largely offset each other such that the net change in radiation at the TOA is relatively small—a characteristic that is consistent with the ERBE observations. The changes are most pronounced over the Tropics, with little change occurring over midlatitudes, again qualitatively consistent with the published analyses of the ERBE data (Wielicki et al. 2002; Cess and Udelhofen 2003). We note that the changes shown in Fig. 6 shift the mean fluxes but have less impact on the extremes in OLR and reflected SW.

The change in tropical-mean OLR and reflected SW

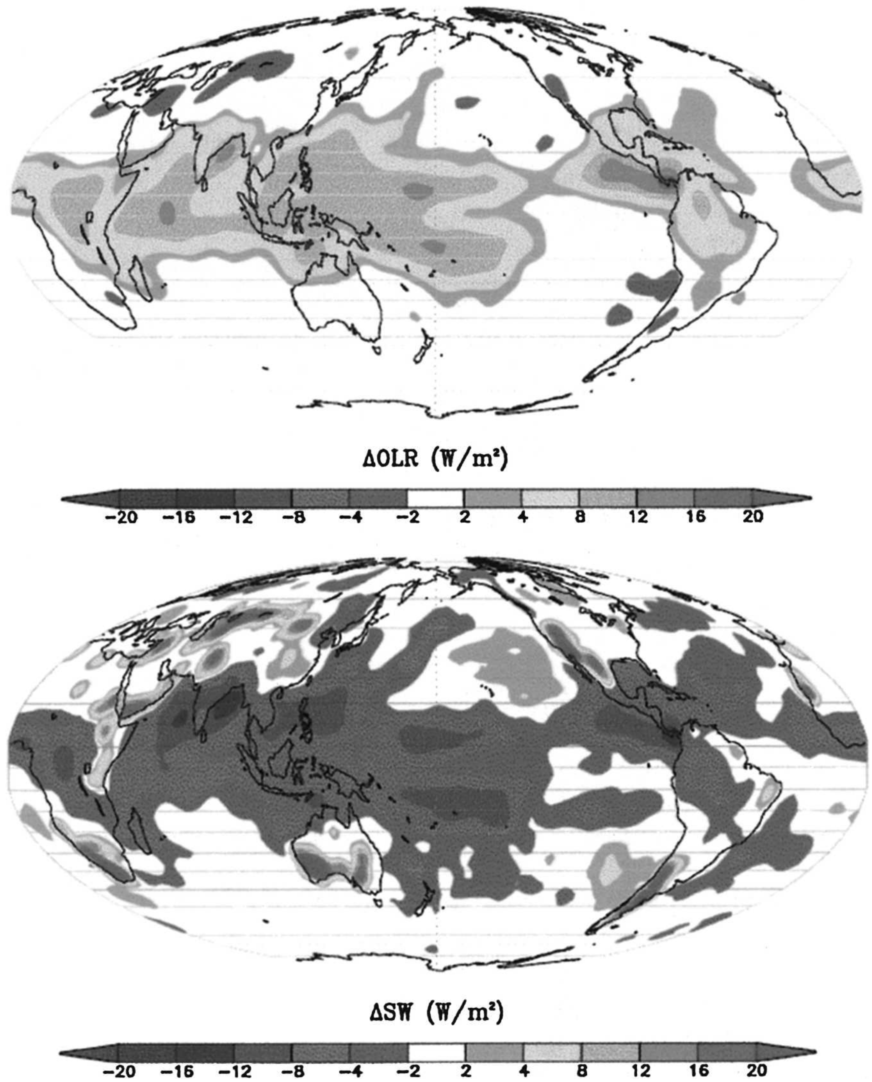

FIG. 6. The change in annual-mean (top) TOA OLR and (bottom) reflected SW radiation simulated by the GFDL GCM due to changes in the convective precipitation efficiency (difference between 0.995 and 0.99).

shown in Fig. 6 are 3.7 and -2.7 W m^{-2}, respectively, which is comparable in magnitude to the changes observed by ERBE between the 1980s and 1990s. This suggests that the model convective precipitation efficiency does have a strong influence on the tropical-mean radiation budget and that a relatively small change in its value would be sufficient for this model to generate an ERBE-like decadal trend. To further explore the sensitivity of the model's radiation budget to this parameter, Fig. 7 shows the change in tropical-mean OLR and reflected SW from 5-yr simulations in which the precipitation efficiency was varied over a range of values. All changes are computed with respect to the model's control value of convective precipitation efficiency (PE = 0.99). Two points can be drawn from this analysis. First, the sensitivity of the radiation budget to a specified change in PE is largest for the highest values of PE. In fact, to a good approximation, the changes in tropical-mean radiation scale in proportion to the fractional change in PE; that is, $\Delta\text{OLR} \sim \Delta\ln\text{PE}$. This behavior arises because the cloud water detrained from the model is proportional to $1 - \text{PE}$. Thus while increasing the PE from 0.99 to 0.995 constitutes only a ~0.5% increase in precipitation efficiency, it represents nearly a 50% reduction in the amount of cloud water available for cirrus formation. This result has potentially important implications. If, for example, the control version of the model used a much lower value of

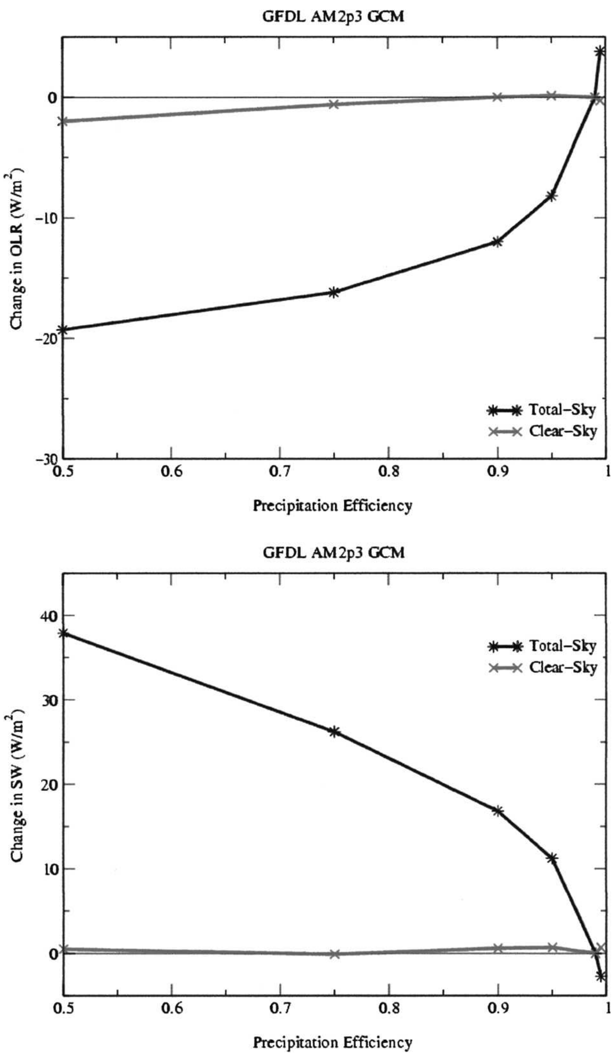

FIG. 7. Changes in tropical-mean (30°N–30°S) (top) OLR and (bottom) reflected SW as a function of the convective precipitation efficiency

precipitation efficiency (e.g., PE ~ 0.5), it would exhibit very little response to a 0.5% increase in precipitation efficiency. It should be noted that the value of PE used in the version of AM2 documented in GFDL Global Atmospheric Model Development Team (2004) is 0.975. The slightly different value from that used in AM2 reflects the fact the PE is used as a tuning parameter in the model to help achieve a net radiative energy balance at the TOA.

The second point to be drawn from this analysis is that while the total-sky radiation fluxes (black lines) are strongly sensitive to change in PE at the high end of the parameter range, the clear-sky radiative fluxes (red lines) are very insensitive to changes in PE across the entire parameter range considered here. The lack of change in clear-sky fluxes is another feature that is consistent with the observations of the decadal-scale trend described in Wielicki et al. (2002). The fact that even a ~50% decrease in PE produces such a small change in the clear-sky OLR reflects the fact that detrained cloud water plays a largely insignificant role in the model in determining the vapor budget of the tropical free troposphere and is consistent with observational analyses (Salathe and Hartmann 1996; Dessler and Sherwood 2000). This suggests that the distribution of upper-tropospheric water vapor can be explained primarily through the large-scale advection of saturated air; thus, reevaporation of cloud condensate is not required to explain the observed humidity distribution.

Unfortunately, there are relatively few observational estimates of convective precipitation efficiency. The uncertainty behind the appropriate range of values for precipitation efficiency reflects both the difficulty of measuring the concentrations of detrained ice relative to precipitating ice in the upper troposphere, as well as our lack of knowledge regarding the underlying microphysical processes that determine this ratio. However, precipitation retrievals from the Tropical Rainfall Measuring Mission (TRMM) Microwave Imager (TMI) using the 2A-12 algorithm (Kummerow et al. 1998) are an exception. This retrieval matches the observed microwave radiances to those simulated from a database of hydrometeor profiles generated by cloud-resolving models (CRMs). As a result, the algorithm yields observationally constrained profiles of cloud water, cloud ice, precipitating water, and precipitating ice. An analysis of the TRMM retrievals by Del Genio and Kovari (2002) indicates that the ratio of precipitating ice to cloud ice in tropical convective storms is ~0.7 for oceanic storms with lighting, ~0.9 for oceanic storms without lightning, and ~0.98 for land storms with or without lightning (see Table 3 of Del Genio and Kovari 2002). Note that oceanic storms containing lightning are relatively infrequent due their weak updraft strengths (Williams and Stanfill 2002).

These numbers should be viewed with caution since they rely heavily on the CRM simulations for which the condensate distributions are known to be sensitive to the particular microphysical parameterization used in the CRM (Xu et al. 2002). Nevertheless, they do illustrate that values of convective precipitation efficiency in excess of 0.9 for the tropical upper troposphere are not inconsistent with the available microwave observations from TRMM. However, the role of precipitation efficiency in modulating the tropical-mean radiation budget in the real climate system is likely to remain uncertain until future observations, such as those from Cloud Aerosol Lidar and Infrared Pathfinder Satellite Observation (CALIPSO) and CloudSat (Stephens et al.

2002), can better constrain the amount and distribution of cloud ice in the upper troposphere. While it may never be feasible to measure precipitation efficiency with an accuracy of a few percent, these results do suggest that if observations can at least exclude the likelihood that the value of convective PE lies at the high end of the spectrum (i.e., >0.95), it would have important implications regarding the sensitivity of the climate system to possible changes in PE.

5. Discussion and implications

Motivated by discrepancies between satellite-observed and model-simulated decadal variations in the OLR and SW over the Tropics, this study investigates the sensitivity of the tropical mean radiation budget using a combination of model simulations and empirical analyses. Experiments with GCMs were designed to test two existing hypotheses: that the decadal variations can be caused by an increase in atmospheric circulation or by a change in convective precipitation efficiency. The results show markedly different sensitivities for these two processes.

In four different GCMs, when the Hadley circulation is increased dramatically there is only a modest change in tropical-mean TOA fluxes. The quantitative change in OLR and SW depends on cloud feedbacks that are highly model dependent: not only the magnitude, but the sign of the response varies from model to model. Acknowledging that the parameterization of clouds is a weak point in GCMs, we turn to the observations to provide an empirically based estimate of the effect of a change in Hadley cell strength on tropical-mean TOA fluxes. By combining model-simulated changes in tropical circulation with ERBE estimates of the dependence of cloud forcing upon vertical velocity, we find that the tropical mean TOA fluxes are remarkably insensitive to changes in the strength of the circulation. This is because a change in circulation primarily redistributes the clouds and radiative fluxes with very little residual effect on the tropical mean. According to this empirical analysis, altering the tropical-mean radiative fluxes by $\sim$3 W m^{-2} would require at least a doubling in the strength of the tropical circulation over this period.

On the other hand, a second set of experiments show that tropical-mean TOA fluxes can be quite sensitive to a key microphysical parameter, the precipitation efficiency. The model shows a strong sensitivity of tropical-mean OLR to small changes in precipitation efficiency at the high end (>0.95) of the parameter range. This region of high sensitivity falls within the current observational estimates from TRMM. Thus, while our ability to measure the appropriate value of precipitation efficiency from observations is limited (even in a tropical-mean sense), these results suggest that better estimates of this quantity are necessary to understand controls on fluctuations in the tropical mean radiation budget. This will likely require improved knowledge of the concentrations of detrained cloud water relative to precipitating water. Observations from active spaceborne sensors like CloudSat and CALIPSO will hopefully provide the measurements necessary to help constrain this problem.

In this study, we have not addressed the mechanisms that can lead to a change in the tropical circulation or in convective precipitation efficiency, nor have we attempted to evaluate whether such changes did, in fact, occur during the ERBE period of record. Rather, we focus on the discrepancy between models and data and address whether a model with increased tropical circulation or altered cloud microphysics is capable of reproducing the observed changes in the tropical radiation fluxes. The results raise questions about the plausibility of both of these scenarios. First, because of the weak sensitivity of tropical-mean radiative fluxes on circulation strength, it is unlikely that if a model were to simulate an increased circulation over the satellite period that that alone would produce a change in the radiative fluxes as large as has been observed. Second, if cloud microphysics were allowed to change, even slightly, in the model considered here, this process could generate a change comparable to that observed. This result is subject to the caveat that the precipitation efficiency must be at the high end of the parameter range to produce such sensitivity. While the current version of the model is in this high end, further observational work is necessary to determine whether this is appropriate.

Finally, it is entirely plausible that circulation may be linked to precipitation efficiency or may occur in combination with other cloud microphysical processes that change over time. For example, while an increase in the strength of the Hadley cell has little impact on the tropical-mean fluxes, it does alter the pattern of OLR and SW in a way that resembles the observations (Allan and Slingo 2002; Chen et al. 2002). On the other hand, increasing the convective precipitation efficiency in AM2 can alter the tropical-mean fluxes, but the change is fairly spatially uniform unlike the observations. So it may be that some combination of these processes is necessary to explain all the features of the observations.

Acknowledgments. The authors thank Sandrine Bony, Mike Winton, and Takmeng Wong for their generous contributions. The comments of Leo Donner, Steve Klein, Bob Cess, and an anonymous reviewer are also greatly appreciated. This work was supported by NASA Grant NNG04GM67G. BS received additional support from the NOAA–CLIVAR program.

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