Tropical Water Vapor and Cloud Feedbacks in Climate Models: A Further Assessment Using Coupled Simulations

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ABSTRACT:

An earlier evaluation of tropical cloud and water vapor feedbacks using AMIP runs of some leading climate models reveals two common biases: (1) an underestimate of the strength of the negative cloud albedo feedback and (2) an overestimate of the positive feedback from the greenhouse effect of water vapor. Extending the same analysis to the fully coupled simulations of these models, we find that these two common biases persist. Relative to the earlier estimates, the overestimate of the positive feedback from water vapor is alleviated somewhat for most of the models. Improvements in the simulation of the cloud albedo feedback are only found in the models whose AMIP runs suggest a positive or nearly positive cloud albedo feedback. The strength of the negative cloud albedo feedback in all other models is found to be substantially weaker than that suggested in the previous analysis. Now all models are found to have a weaker negative feedback from the net surface heating over the ocean than that indicated in observations. The weakening in the cloud albedo feedback is linked to a reduced response of deep convection over the equatorial Pacific which is in turn linked to the excessive cold-tongue in the mean climate of these models. The results underscore the nonlinear relationship between deep convection and SST and suggest that the bias in the mean tropical SST can in turn influence the regulatory effect from deep convection.

1. Introduction:

Water vapor provides most of the greenhouse effect of the Earth's atmosphere. Clouds are a major contributor to the planetary albedo (Kiehl and Treneberth 1997). A small change in these radiative effects of water vapor and clouds can either offset or greatly amplify the perturbation to the Earth's radiation balance from anthropogenic effects (Houghton et al. 2001). Therefore, it is imperative for climate models on which our economical polices are increasingly relying on to narrow the uncertainties in their simulations of the feedbacks from water vapor and clouds. Toward that objective, we have to critically evaluate how well the existing leading climate models simulate the feedbacks from water vapor and clouds.

Two methods have been employed to shed insight onto the question how well climate models simulate the feedbacks from water vapor and clouds. The first one is to check the differences in the feedbacks of water vapor and clouds among different models. A pioneering study using this method was carried out by Cess et al. (1990, 1996). Their analysis revealed that the cloud feedbacks differ greatly among models while the globally averaged feedback from water vapor in the models follows that of a constant relative humidity model. A later study by Soden and Held (2006) reached the same conclusion for the IPCC AR4 models (Meehl et al. 2007). These results underscore the uncertainties in the cloud feedbacks in the climate models, but do not address the question which model has the right cloud feedbacks. Another limitation of these results is that consistency in the simulation of water vapor feedback does not rule out the possibility that all the models have a biased water vapor feedback.

The second method is to compare the response of water vapor and clouds to SST changes over the time scales for which observational data are available. A frequently used natural signal in the SST is El Nino warming (Sun and Held 1996, Soden 1997, Held and Soden 2000, Sun et al. 2006). By comparing the observed changes in the water vapor and clouds with those from the model with the observed SST boundary conditions (AMIP simulations), these studies suggest that the sign of the water vapor feedback in the GCMs is probably correct on the time-scale of ENSO, even after averaged over the entire tropics. The study of Sun et al. (2006) shows, however, that models tend to overestimate the positive feedback of water vapor over the immediate region of the El Nino warming. The study of Zhang and Sun (2007) further shows that at least for the NCAR models, the overestimate of the positive feedback of water vapor during El Nino warming is due to an excessive response of upper tropospheric water vapor to the surface warming. A more serious concern raised by the study of Sun et al. (2006) is the finding of a common bias in the simulation of the cloud albedo feedback in the leading climate models: with the exception of the GFDL model, all the models they analyzed in that study underestimate the response of cloud albedo to the surface warming. Nonetheless, the finding that at least the GFDL model may have a cloud albedo feedback as strong as the observed strikes an optimistic tone.

The study of Sun et al. (2006) used the AMIP simulations. Using AMIP simulations of these models to estimate feedbacks has an inherent limitation: the feedbacks are the feedbacks operating in the immediate neighborhood of the observed climatology. As the mean climate is free to drift to the state that is in turn determined by the feedbacks, the feedbacks may change in the process of integration of the coupled model. In other words, if the coupled system—the models of it in particular—is not strictly a linear feedback system, the feedbacks estimated about the equilibrium state of the coupled runs could be significantly different from those estimated from the corresponding AMIP runs if the SST of the equilibrium state of the coupled run differs significantly from the observed. The coupled models do have a significantly different climatological SST from the observed—they all have an excessive cold-tongue in the equatorial central Pacific (Sun et al. 2006). The purpose of this paper is to further assess the feedbacks from water vapor and clouds over the tropical Pacific region by directly using the outputs from fully coupled runs.

In addition to heeding to the general need to assess the fidelity of the water vapor and cloud feedbacks in climate models, we are also motivated to further examine the hypothesis that a weaker negative feedback from the surface heating over the ocean contributes to the development of the excessive cold-tongue—a common syndrome in coupled climate models without flux adjustment (Sun et al. 2003, Sun et al. 2006).

2. Methodology

As in Sun et al. (2006), we will use the response of tropical convection to ENSO forcing to obtain the feedbacks of water vapor and clouds associated with tropical convection. The group of coupled models we have selected for the analysis consists of the models whose AMIP runs we analyzed in the study of Sun et al. 2006. In that study, we examined nine AGCMs. Seven of the nine AGCMS has a corresponding fully coupled GCM whose control runs are available for our analysis. These models are: NCAR CCSM1 (Boville and Gent 1998), the NCAR CCSM2 (Kiehl and Gent 2004), the NCAR CCSM3 at respectively T42 and T85 resolution (Collins et al. 2006, www.ccsm.ucar.edu/experiments/ccsm3.0/), the HadCM3 (Collins et al. 2001), the French IPSL CM4 model (Marti et al. 2005), and the GFDL CM2.0 (Delworth et al. 2006, http://data1.gfdl.noaa.gov/nomads/forms/decen/CM2.X/). As in Sun et al. (2006), we also assess the corresponding feedback from the atmospheric transport and deduce the feedback from the net surface heating from the energy balance equation of the atmosphere. Unless explicitly stated, calculations for all models use a 50 year-long segment of the control runs of the models.

3. Results:

Applying the same linear regression technique of Sun et al. (2006) to the simulations of tropical inter-annual variations by those coupled models, we obtain Table 1. The observation results listed in the table are those reported in Sun et al. (2006).

The problems uncovered in the previous analysis also show up in this extended analysis. First, models tend to underestimate the strength of the negative feedback from cloud albedo. The model that now has the strongest negative cloud albedo feedback is the IPSL/CM4, but the feedback is only of 70% of the observed value. Substantial weakening in the simulated strength of the cloud albedo feedback occurs in all the four models that were identified as better models in the previous analysis using their AMIP runs (the NCAR Model at T85, the UK model, the French Model, and the GFDL model; See Table I of Sun et al. 2006). This reduction in the strength of the cloud albedo feedback is particularly notable for GFDL model—the value changed from -12.58 $Wm^{-2}K^{-1}$ in the previous analysis to -7.64 $Wm^{-2} K^{-1}$ in the present analysis.

Fig. 1 provides a basin-wide view of the response of C_s to El Nino warming in these coupled models. The spatial pattern of the response of C_s in the coupled models resembles that obtained from their corresponding AMIP runs (see Fig. 5 in Sun et al. 2006), but the maximum response of C_s is located more westward in the coupled models by about 20^o. With the exception of NCAR CCSM2 and CCSM3 at T42, the maximum response of C_s in the coupled simulations is also weaker than that in the corresponding AMIP runs. In the AMIP runs, the maximum response of Cs in

NCAR CCSM3 at T85, UKMO/HadCM3, IPSL/CM4, and GFDL/CM2.0 has a value that exceeds -30 $Wm^{-2}K^{-1}$. In their corresponding coupled runs, however, the maximum response of C_s in these four better models is significantly reduced. The reduction in the maximum response of C_s in NCAR CCSM3 at T85, IPSL/CM4, and GFDL/CM2.0 is about10 $Wm^{-2}K^{-1}$. The reduction in the maximum response of C_s in UKMO/HadCM3 is even higher (~15 $Wm^{-2}K^{-1}$).

There are exceptions to this general weakening in the response of C_s --NCAR CCSM2 and NCAR CCSM3 at T42. NCAR CCSM2 has no negative response in its AMIP run, but now develops a weak but detectable negative response in the region immediately west to the dateline (150^{o}E-170^{o}E). The improvements in NCAR CCSM3 at T42 are even more substantial. The response of C_s in this model is comparable to that in GFDL/CM2.0. Such a "self-correction" of the cloud albedo feedback clearly indicates the importance of nonlinearity in the coupling between the atmosphere and ocean. While it is encouraging to see that ocean-atmosphere coupling allows a "self-correction" to take place, it is disappointing to see that this "self-correction" is limited to the two models whose cloud albedo feedback assessed at the immediate neighborhood of the observed SST has the largest error. As indicated by the results from their AMIP runs, the negative feedback from cloud albedo barely exists in these two models in the immediate neighborhood of the observed SST (Table 1 in Sun et al. 2006). All other models that have been judged from their respective AMIP runs to a significantly negative feedback of cloud albedo at the observed SST are found to have an even weaker cloud albedo feedback at their respective equilibrium SST. (In other words, the negative feedback from cloud albedo assessed from their coupled runs is weaker than that estimated from the AMIP runs).

The general weakening of the response of C_s in the coupled models (relative to the values estimated from the AMIP runs) is linked to the weakened precipitation response. Fig. 2 shows the precipitation response to El Nino warming in the coupled models. Contrasting Fig. 2 with Fig. 6 of Sun et al. (2006), one finds that the reduction in the maximum response of the precipitation in NCAR CCSM3 at T85, UKMO/HadCM3, IPSL/CM4, and GFDL/CM2.0 all exceeds 30%. The location of the maximum response of precipitation in these coupled models also shifts westward relative to that in the AMIP runs. The general reduction in the precipitation and the westward shift of the response suggests that the excessive cold-tongue in these models plays a role in further weakening the response of C_s to SST changes in that region. (Interestingly, the two models (NCAR CCSM2 and CCSM3 at T42) that has an improved cloud albedo feedback are the only two models that do not have a significant weakening in their maximum response in the precipitation). It has been noted before that the precipitation depends nonlinearly on the SST (Lighthill et al. 1994). As the models have a colder SST relative to their respective maximum SST over the equatorial Pacific, there is just not an adequate level of deep convection to begin with over the central Pacific to respond SST changes. To examine this nonlinearity more quantitatively, we plotted the scatter diagram of precipitation and SST over the cold-tongue region (Fig3a). The figure indeed shows that in the observations, the precipitation increases with SST increases at a faster rate when SST is warmer. Such nonlinearity also exists in the models, but is weaker than in the observations. The corresponding figure for the surface level solar radiation shows that the relationship between the surface solar radiation and SST mirrors the relationship between precipitation and SST (Fig.3b). The surface solar radiation decreases at a faster rate with SST when SST is warmer. All models underestimate the rate of decrease in the solar radiation with increasing SST. Fig. 3a and Fig. 3b together reinforce the impression that the weakened cloud albedo feedback over the equatorial Pacific in the coupled models is linked to the weakened deep convection over that region.

The new estimates also confirm another common bias existing in the climate models: the overestimate of the positive feedback of water vapor. Comparing Table 1 in the present paper with

Table 1 in Sun et al. (2006) reveals that estimating from their AMIP runs of the climate model can result in a stronger positive feedback of water vapor. For example, the new estimate of the water vapor feedback in the UKMO/HadCM3 is now significantly closer to the observed value than that from its AMIP runs. Still, all the models have a stronger water vapor feedback than what indicated in ERBE. The overestimate ranges from about 25% in NCAR CCSM2 to about 45% in NCAR CCSM3 and IPSL/CM4. Discrepancy of this magnitude is hard to be accounted for by errors in the ERBE observations. In any case, the existence of a significant spread in this discrepancy among these models suggests that any agreement in the global averaged values among these models in this regard must be linked to error cancellations among different regions in these models. Fig. 4 shows the spatial pattern of the response of G_a. Clearly, the models do not just differ from the observations over the immediate region of surface warming due to El Nino. They also differ in the surrounding regions—the far western Pacific and the subtropical regions. The spatial pattern of G_a also suggests a westward shift of convection in the coupled models relative to their AMIP runs (see Fig. 2 in Sun et al. 2006).

With the exception of the NCAR CCSM2 and CCSM3, the new estimates using the coupled simulations also yield a lower value for the positive feedback from the greenhouse effect of clouds. The decrease in the strength of the positive feedback from the greenhouse effect of clouds is consistent with the decrease in the strength of the negative feedback from the cloud albedo. The exceptional behavior in the NCAR two models in this regard is also consistent with the exceptional behavior in these two models in simulating the cloud albedo feedback (recall that the feedback from cloud albedo in these two models is estimated to be more negative than that estimated from AMIP runs). The contribution to the increase in the value of $\frac{\partial}{\partial T} C_l$ as shown in Table 1 in these two NCAR models mainly comes from the western edge of the equatorial cold-tongue region defined here (150$^{\mathrm{o}}$E-250$^{\mathrm{o}}$E). Fig. 5 shows spatial pattern of the response of the greenhouse effect of clouds (C_l)

in these two models. Judging from Fig. 5 and the corresponding pattern of the precipitation response (Fig. 2), it is likely that the large positive initial bias in the cloud albedo feedback (the bias in the immediate neighborhood of the observed SST) in these two models enhances the deep convection in the region about 150 ^{o}E-170 ^{o}E (where SST is already warm) to a degree that the weakening effect from the excessive cold-tongue is offset.

As indicated in the estimates using the AMIP runs, not all the models overestimate the feedback from the total greenhouse effect of water vapor and clouds (G_a+ C_l). Two of the models are actually found to underestimate the combined feedback from G_a and C_l (The NCAR CCSM2 and the UKMO/HadCM3). Clearly, the relationship between G_a and C_l is not the same in all the models.

The new estimates of the atmospheric transport result in no new results for the two European models. The change in this feedback in the GFDL CM2 is also small. All NCAR models, however, have a much closer result to the observational estimate in this aspect. With the exception of the UKMO/HAdCM3, the feedback of the atmospheric transport is well simulated by the coupled models.

Seen in the net atmospheric feedback ($\frac{\partial F_a}{\partial T}$) or the feedback from the net surface heating ($\frac{\partial F_s}{\partial T}$), the discrepancy with the observations in the new estimates increases in the models that were identified as better models in this regard in the previous analysis (GFDL CM2, IPSL/CM4, HadAM3, and NCAR CAM3-T85), but decreases in the models that were identified as the worse models in this regard (NCAR CAM1, NCAR CAM2, and NCAR CAM3). No models in the new estimates have a regulatory effect that is comparable to the observations. The best model identified in the previous analysis—the GFDL/CM2, however, remains the one that has the strongest the regulatory effect, though the new estimate suggests that this regulatory effect from deep convection in this model is still too weak compared to observations (about $-10 Wm^{-2}K^{-1}$ in the model versus about $-15\ Wm^{-2}K^{-1}$)

4. Conclusion.

The extended calculation using coupled runs confirms the earlier inference from the AMIP runs that underestimating the negative feedback from cloud albedo and overestimating the positive feedback from the greenhouse effect of water vapor over the tropical Pacific is a prevalent problem of climate models. The estimates from the coupled simulations of both the cloud albedo feedback and the water vapor feedback differ from the estimates from the corresponding AMIP simulations. The changes in the cloud albedo feedback are particularly significant. The previous analysis of Sun et al. (2006) has suggested that the GFDL CM2 may have a cloud albedo feedback that is as strong as observations. The new estimate puts this suggestion in doubt as the new estimate is significantly weaker than the previous estimate. All models are found to have a weaker negative feedback from the net surface heating than that from observations, indicating that deep convection over the equatorial Pacific in the models has a weaker regulatory effect over the SST in that region.

The results underscore the nonlinear relationship between deep convection and SST over the equatorial Pacific. This nonlinearity may lie behind the difficulty for coupled GCMs to simulate the equilibrium SST distribution in that region. Once the delicate balance among the various forces that maintain the zonal position and intensity of the equatorial Pacific cold-tongue is tilted and an initial colder SST anomaly (relative to the equilibrium state) is created in the equatorial central Pacific as a consequence, the colder SST then further pushes the deep convection westward which further weakens or even eliminates entirely the regulatory effect from deep convection over the SST. This nonlinear temperature dependence of the damping effect from deep convection over the SST deviations though surface heating also explains the prevalence of a cold-bias (relative to the warm-pool SST), but not a warm bias in the equatorial Pacific. The former quickly reinforces itself through its weakening of the regulatory effect from deep convection, while the latter enhances the regulatory effect of deep convection which in turn limits its further growth.

It should be emphasized that the feedbacks assessed are the feedbacks over the equatorial Pacific on the time-scale of ENSO. Though the systematic biases revealed in this analysis underscore the need to improve the simulation of the feedbacks in climate models, they do not necessarily imply that the sensitivity of the mean tropical climate to anthropogenic forcing is overestimated.

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FIGURE CAPTIONS

Figure 1: Response of the short-wave forcing of clouds (Cs) to El Nino warming. Shown are coefficients obtained by linearly regressing Cs at each grid point on the SST averaged over the equatorial Pacific (5^oS-5^oN, 150^oE-250^oE).

Figure 2: Same as for Fig.1, but for the precipitation. The result for the observation is from Sun et al. (2006).

Figure 3: Scatter diagrams showing the relationship between the net precipitation and the SST (a) and the relationship between the surface solar radiative heating and the SST (b) . Interannual anomalies of these quantities averaged over the equatorial Pacific (5^oS-5^oN, 150^oE-250^oE) and for the period July 1983—June 2001 are used for these figures. The surface solar radiation data are from ISCCP (Rossow and Schiffer 1999). The precipitation data are from Xie and Arkin (1996). The SST data are for from Rayner et al. (1996).

Figure 4: Same as for Fig. 1, but the greenhouse effect of water vapor (G_a).

Figure 5: Same as for Fig. 1, but the greenhouse effect of clouds (C_l).

TABLE CAPTIONS:

Table 1: Tropical cloud and water vapor feedbacks from observations and coupled climate models. $\frac{\partial}{\partial T}G_a$ is the water vapor feedback, $\frac{\partial}{\partial T}C_l$ is the feedback from the long-wave forcing of clouds (the greenhouse effect of clouds), and $\frac{\partial}{\partial T}C_s$ is the feedback from the short-wave forcing of clouds (cloud albedo feedback). The feedback from the atmospheric transport ($\frac{\partial}{\partial T}D_a$), the net atmospheric feedback ($\frac{\partial F_a}{\partial T} = \frac{\partial G_a}{\partial T} + \frac{\partial C_l}{\partial T} + \frac{\partial C_s}{\partial T} + \frac{\partial D_a}{\partial T}$), and the feedback from net surface heat flux into the ocean ($\frac{\partial}{\partial T}F_s$) are also listed. The values for these feedbacks are obtained through a linear regression using the inter-annual variations of the SST and the corresponding fluxes over the equatorial Pacific (5^oS-5^oN, 150^oE-250^oE) from a 50 year long simulations by the models. The observational results are those obtained in Sun et al. (2006). The model listed are: NCAR CCSM1 (Boville and Gent 1998), the NCAR CCSM2 (Kiehl and Gent 2004), the NCAR CCSM3 at T42 and T85 resolution (Collins et al. 2006), the French IPSL-CM4 (Marti et al. 2005), and the GFDL CM2 (Delworth et al. 2006).

Atmospheric Feedbacks in Models and Observations

Name of Process	Feedback ($Wm^{-2}K^{-1}$)							
	Observation	NCAR/CCSM1	NCAR/CCSM2	NCAR/CCSM3	CCSM3 (T85)	UKMO/HadCM3	IPSL/CM4	GFDL/CM2.0
$\frac{\partial(G_a)}{\partial T}$	6.72 ± 0.27	8.47 ± 0.13	8.25 ± 0.08	9.35 ± 0.09	9.05 ± 0.10	8.89 ± 0.11	9.17 ± 0.06	8.82 ± 0.08
$\frac{\partial(C_L)}{\partial T}$	12.21 ± 1.03	10.89 ± 0.34	7.21 ± 0.15	9.68 ± 0.20	11.35 ± 0.34	6.64 ± 0.18	11.89 ± 0.23	10.22 ± 0.29
$\frac{\partial(G_a+C_L)}{\partial T}$	18.93 ± 1.17	19.36 ± 0.45	15.46 ± 0.21	19.03 ± 0.26	20.41 ± 0.41	15.54 ± 0.29	21.06 ± 0.26	19.04 ± 0.34
$\frac{\partial(C_S)}{\partial T}$	-10.93 ± 1.37	-2.57 ± 0.21	-0.31 ± 0.17	-5.39 ± 0.38	-5.35 ± 0.41	-5.26 ± 0.28	-7.73 ± 0.27	-6.14 ± 0.39
$\frac{\partial(D_a)}{\partial T}$	-16.69 ± 1.51	-16.66 ± 0.52	-15.30 ± 0.28	-17.08 ± 0.34	-16.40 ± 0.45	-12.53 ± 0.34	-17.26 ± 0.29	-17.17 ± 0.41
$\frac{\partial(F_A)^*}{\partial T}$	-8.69 ± 1.76	0.13 ± 0.47	-0.15 ± 0.26	-3.44 ± 0.45	-1.34 ± 0.48	-2.25 ± 0.34	-3.93 ± 0.30	-4.28 ± 0.49
$\frac{\partial(F_S)}{\partial T}$	-14.89 ± 1.83	-5.86 ± 0.47	-6.15 ± 0.26	-9.53 ± 0.45	-7.38 ± 0.49	-8.13 ± 0.34	-9.96 ± 0.30	-10.25 ± 0.50

The net atmospheric feedback $\frac{\partial(F_A)}{\partial T} = \frac{\partial(G_a)}{\partial T} + \frac{\partial(C_L)}{\partial T} + \frac{\partial(C_S)}{\partial T} + \frac{\partial(D_a)}{\partial T}$

Table 1: Table 1: Tropical cloud and water vapor feedbacks from observations and coupled climate models. $\frac{\partial}{\partial T}G_a$ is the water vapor feedback, $\frac{\partial}{\partial T}C_l$ is the feedback from the long-wave forcing of clouds (the greenhouse effect of clouds), and $\frac{\partial}{\partial T}C_s$ is the feedback from the short-wave forcing of clouds (cloud albedo feedback). The feedback from the atmospheric transport ($\frac{\partial}{\partial T}D_a$), the net atmospheric feedback ($\frac{\partial F_a}{\partial T} = \frac{\partial G_a}{\partial T} + \frac{\partial C_l}{\partial T} + \frac{\partial C_s}{\partial T} + \frac{\partial D_a}{\partial T}$), and the feedback from net surface heat flux into the ocean ($\frac{\partial}{\partial T}F_s$) are also listed. The values for these feedbacks are obtained through a linear regression using the inter-annual variations of the SST and the corresponding fluxes over the equatorial Pacific (5°S-5°N, 150°E-250°E) from a 50 year long simulations by the models. The observational results are those obtained in Sun et al. (2006). The model listed are: NCAR CCSM1 (Boville and Gent 1998), the NCAR CCSM2 (Kiehl and Gent 2004), the NCAR CCSM3 at T42 and T85 resolution (Collins et al. 2006), the French IPSL-CM4 (Marti et al. 2005), and the GFDL CM2 (Delworth et al. 2006).

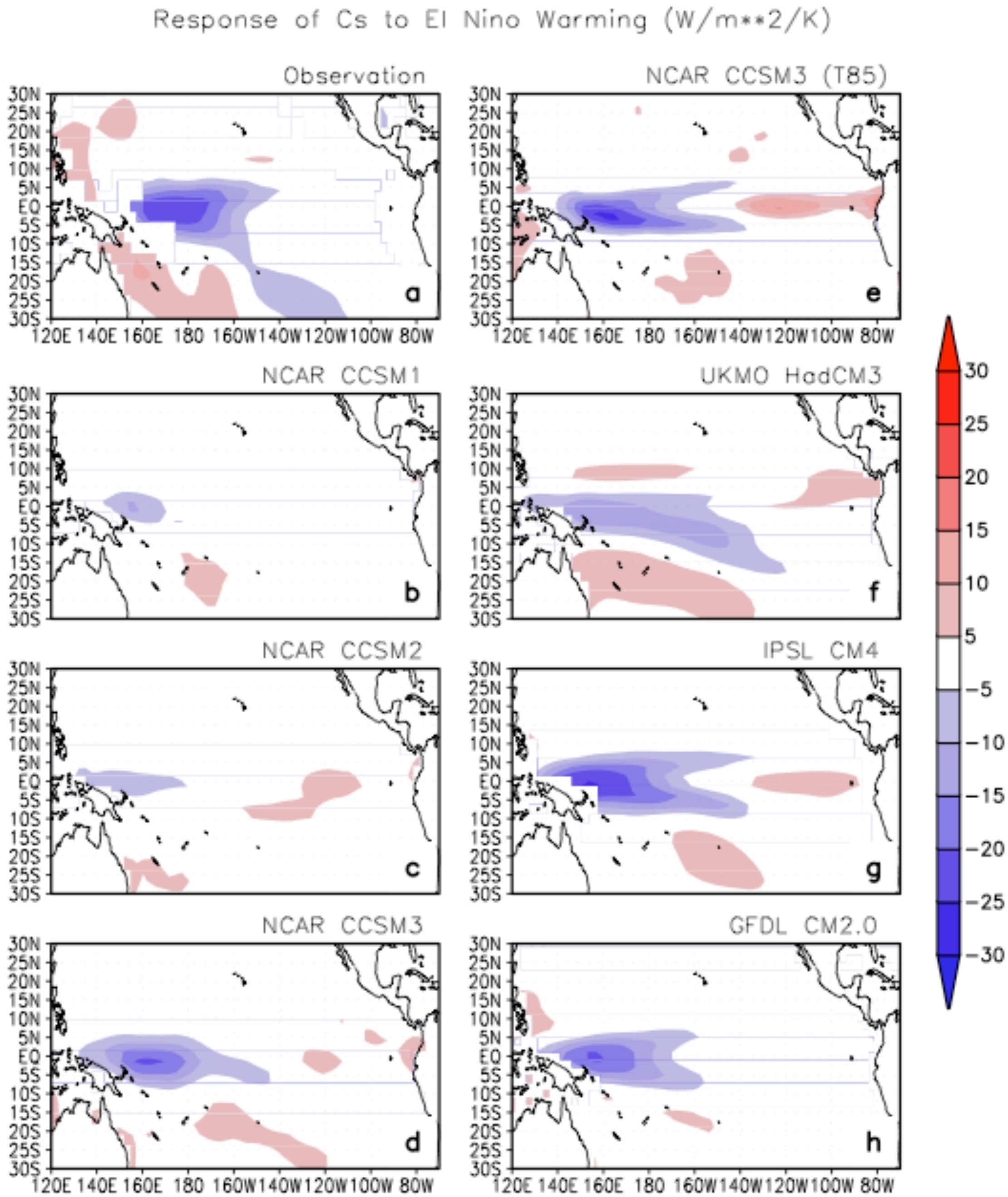


Figure 1: Response of the short-wave forcing of clouds (Cs) to El Nino warming. Shown are coefficients obtained by linearly regressing Cs at each grid point on the SST averaged over the equatorial Pacific (5°S-5°N, 150°E-250°E).

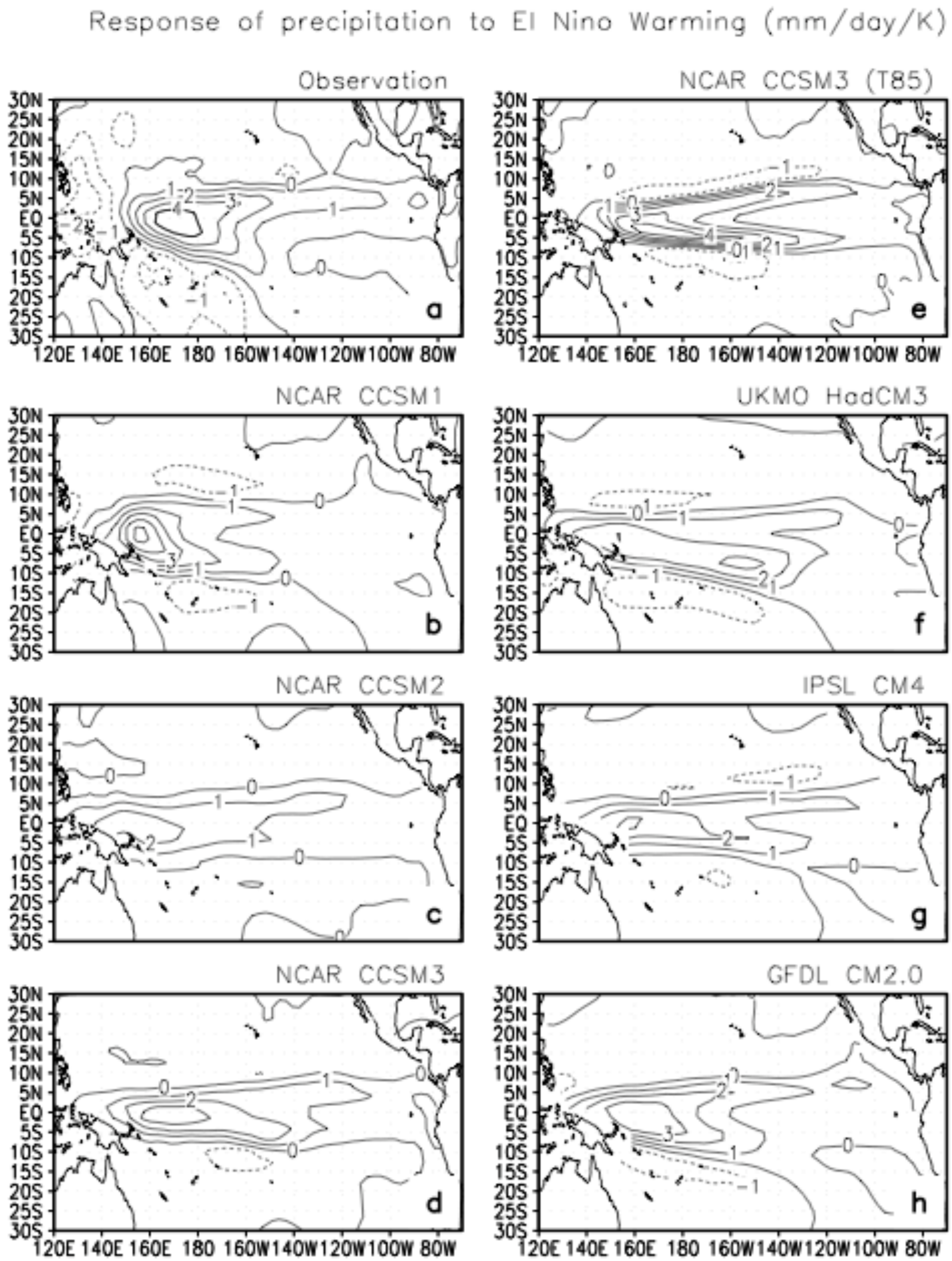
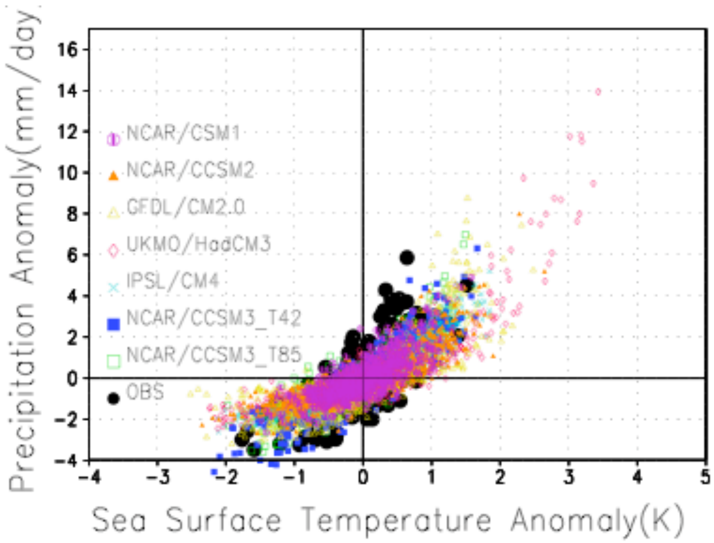


Figure 2: Same as for Fig.1, but for the precipitation. The result for the observation is from Sun et al. (2006).

Fig.3

(a)



(b)

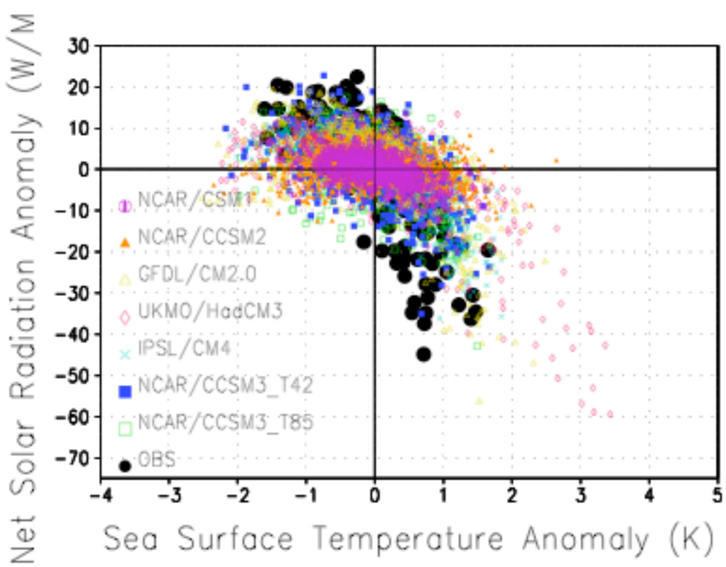


Figure 3: Scatter diagrams showing the relationship between the net precipitation and the SST (a) and the relationship between the surface solar radiative heating and the SST (b) . Interannual anomalies of these quantities averaged over the equatorial Pacific (5^{o}S-5^{o}N, 150^{o}E-250^{o}E) and for the period July 1983—June 2001 are used for these figures. The surface solar radiation data are from ISCCP (Rossow and Schiffer 1999). The precipitation data are from Xie and Arkin (1996). The SST data are for from Rayner et al. (1996).

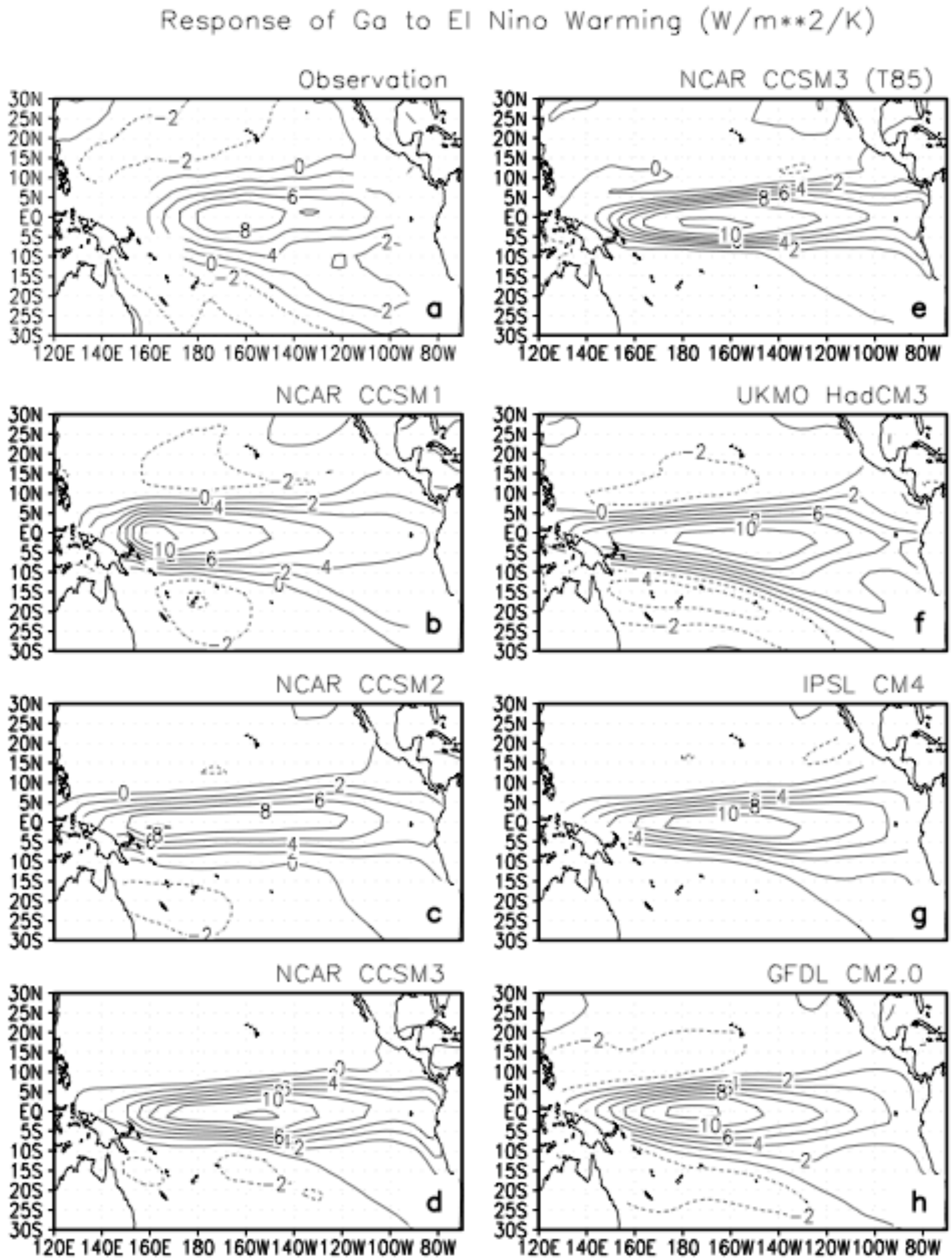


Figure 4: Same as for Fig. 1, but the greenhouse effect of water vapor (G_a).

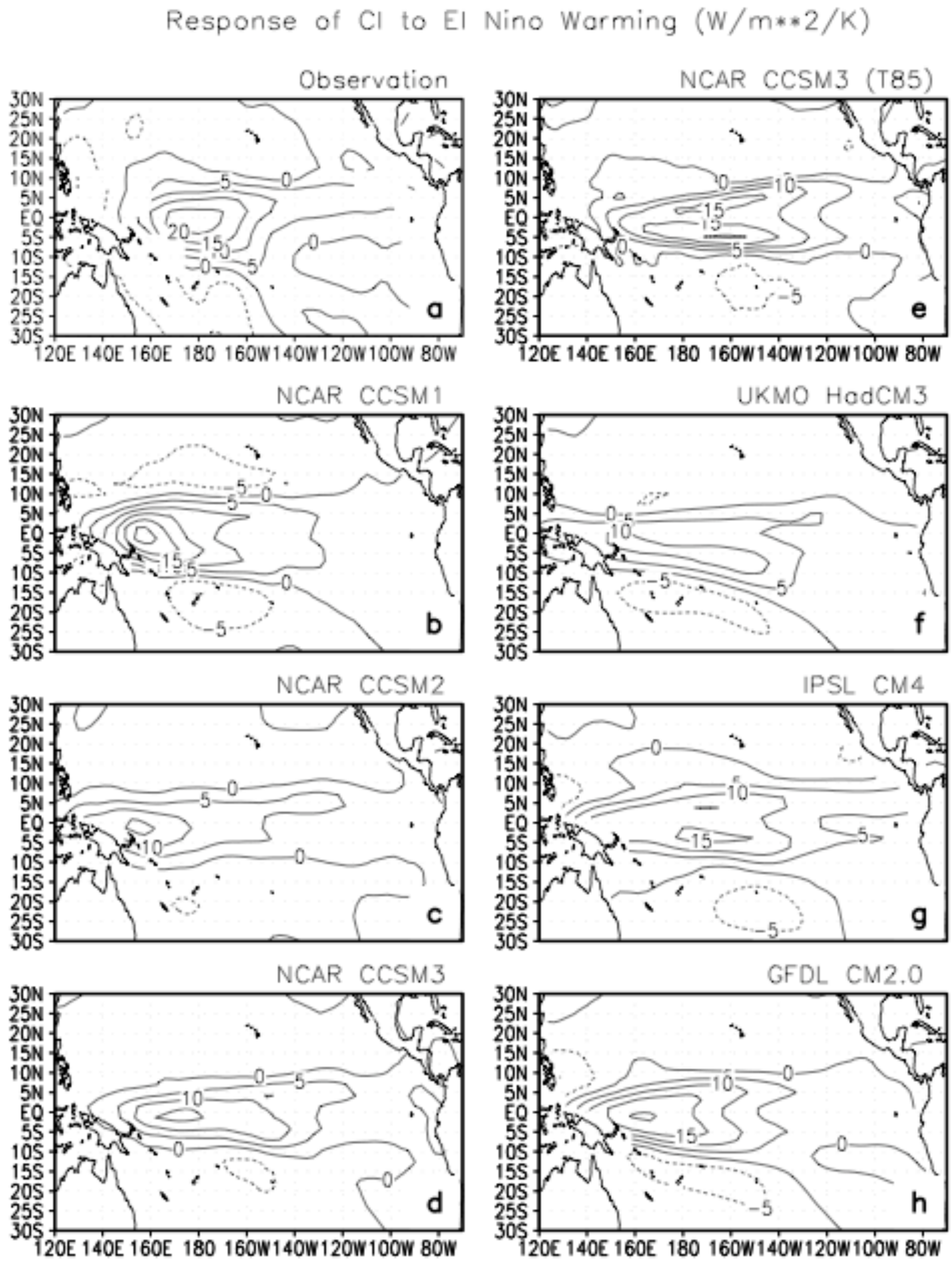


Figure 5: Same as for Fig. 1, but the greenhouse effect of clouds (C_l).