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Tropical cirrus and water vapor: an effective Earth infrared iris feedback?

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Abstract. We revisit a model of feedback processes proposed by Lindzen et al. (2001), in which an assumed 22% reduction in the area of tropical high clouds per degree increase in sea surface temperature produces negative feedbacks associated with upper tropospheric water vapor and cloud radiative effects. We argue that the water vapor feedback is overestimated in Lindzen et al. (2001) by at least 60%, and that the high cloud feedback is small. Although not mentioned by Lindzen et al. (2001), tropical low clouds make a significant contribution to their negative feedback, which is also overestimated. Using more realistic parameters in the model of Lindzen et al. (2001), we obtain a feedback factor in the range of -0.15 to -0.51 , compared to their larger negative feedback factor of -0.45 to -1.03 . It is noted that our feedback factor could still be overestimated due to the assumption of constant low cloud cover in the simple radiative-convective model.

1 Introduction

Motivated by the observational evidence that in the Tropics the boundary between regions of high and low free-tropospheric relative humidity is sharp, and that areas of high free-tropospheric relative humidity tend to be localized near upper-level cirrus, Lindzen et al. (2001) (hereafter LCH) analyzed cloud data for the eastern part of the western Pacific from the Japanese GMS-5. They concluded that the area of upper-level cloud coverage normalized by the area of cumulus coverage decreases about 22% per degree Celsius increase in the sea surface temperature (SST) of the cloudy region. They interpreted this as the effects of ocean surface temperature on cirrus detrainment from cumulus convection. This observational result, while open to question (Hartmann and Michelsen, 2001), is not the focus of the present paper.

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Instead, we will explore the climatic implications of the decrease of high cloud area with increasing SST by employing the same model framework as in LCH.

LCH suggested that decreasing areas of upper-level cloud and high moisture with increasing SST would produce a rise in the outgoing longwave radiation (OLR) and thus a negative feedback in the climate system. Such a feedback was referred to as an adaptive infrared iris. LCH examined the iris feedback using a two-dimensional radiative-convective model. Their calculations showed that changes of the relative areas of tropical moist and dry air in response to changes in SST could lead to a feedback factor of as much as about -1.03 . Such a large negative feedback would more than cancel all the positive feedbacks in the more sensitive current climate models.

There are three contributions to the iris effect suggested by LCH: that due to water vapor, that due to high clouds and that due to low clouds. (This last is not acknowledged by LCH.) Here the cloud and water vapor feedbacks are inextricably tied to each other. Based on the calculations by LCH, the contribution to the iris feedback factor due to water vapor is negative and lies between 0 and -0.58 depending on how well the moist area follows the cirrus cloud area. According to their calculations, even if regions of high humidity were not coupled to cloudiness, the feedback factor due to clouds alone would still amount to about -0.45 , which would cancel model water vapor feedback in almost all climate models. The strong negative feedbacks that result from the assumed negative correlation between SST and high cloud area are based on two major arguments in LCH. First, they assert that the effect of tropical high clouds on the OLR is much larger than their effect on reflected solar radiation. Second, they find that the greenhouse effect of water vapor in the tropical moist regions is much greater than that in tropical dry regions. It should be noted that the strong negative feedback presented in LCH is not only due to water vapor and high clouds, but also due to low clouds, which was not acknowl-

edged in LCH. LCH assume that low clouds are everywhere in the Tropics with a constant cloud cover and high cloud albedo, and thus a strongly negative cloud radiative forcing. Therefore, a decrease in cirrus cloud cover would increase the area with strong negative cloud forcing, which would always produce a cooling (negative feedback).

In this paper, we will examine the major arguments used in LCH to derive the negative feedbacks related to high clouds, low clouds, and water vapor. We discuss the impact of these arguments on their main conclusions on climate sensitivity. On the basis of calculations and observations we find that the contrast in clear-sky OLR between moist and dry regions is overestimated by at least 60% in LCH. Observations suggest that the tropical high cloud region has near zero net radiative forcing, and that the net radiative effect of low clouds is smaller than that assumed by LCH. Therefore we find a much reduced magnitude for each of the contributions to the hypothetical iris feedback. Assuming a 22% decrease in high cloud coverage per degree of SST increase, and using the model framework of LCH, we obtain a feedback factor of -0.15 to -0.51 , compared to the range of -0.45 to -1.03 obtained by LCH.

2 The feedback factor and LCH's 3.5-box model

Following LCH, the net response, ΔT , of the climate system to an external forcing, ΔQ , can be expressed in the equation

$$\Delta T = G_0(\Delta Q + F\Delta T)$$

where G_0 is the no-feedback gain of the climate system and F is the additional forcing per degree increase in the net response due to the feedback process. Solving for ΔT , we have

$$\Delta T = G_0\Delta Q/(1 - G_0F) = \Delta T_0/(1 - f)$$

where $\Delta T_0 = G_0\Delta Q$ is the response of the climate system to the forcing in the absence of feedbacks and $f = G_0F$ is referred to as the feedback factor. In the iris mechanism, F is the product of two parts. The first part is the change in tropical high cloud cover with SST, $\Delta A/\Delta T_s$, which is assumed to be 22% reduction of the high cloud fraction per degree Kelvin increase of SST. The other part is the change in the TOA (top of the atmosphere) radiation balance with a change in high cloud cover, $\Delta R_{\text{net}}/\Delta A$, which has contributions from the direct radiative effect of the clouds and their assumed effect on the area in which the upper troposphere is moist. Therefore,

$$F = \frac{\Delta A}{\Delta T_s} \frac{\Delta R_{\text{net}}}{\Delta A}$$

LCH constructed a 3.5-box model with radiative-convective equilibrium to calculate G_0 and $\Delta R_{\text{net}}/\Delta A$. In this model, the world is divided into three regions: the moist Tropics (25%), the dry Tropics (25%), and the extratropics (50%).

The SSTs for the tropical and extratropical regions are assumed to be 10 degrees higher and lower, respectively, than the global mean SST that has a current value of 288 K. The moist region of the Tropics is further divided into a cloudy-moist region (11%) covered by upper level cirrus, and a clear-moist region (14%) clear of such cirrus. The low cloud cover is assumed to be 25% everywhere in the Tropics; that is, 25% of the surface in the Tropics is covered by low clouds. The reflectivities of tropical high and low clouds are assumed to be 0.24 and 0.42, respectively. The tropical clear sky reflectivity is 0.13. The emission temperatures used are the tropical SST minus 27.6 K, 37 K, and 76 K for the tropical dry, clear-moist, and cloudy-moist regions, respectively. These choices give OLRs of 303 Wm^{-2} , 263 Wm^{-2} , and 138 Wm^{-2} , respectively. The complete list of choices of parameters for the 3.5-box model are given in Table 1 of LCH. They are consistent with the Earth Radiation Budget Experiment (ERBE) observations of the reflectivities and emission temperatures averaged for the tropical and extratropical regions and the globe.

To evaluate the magnitude of the iris feedback, LCH varied the areas of the tropical regions, decreasing the tropical cloudy- and clear-moist regions and increasing the tropical dry region accordingly. Here it is instructive to evaluate the relative contributions of water vapor, low clouds, and high clouds to the TOA radiative energy budget change in response to changes of high cloud area using the LCH parameter values. Consider a decrease of high cloud cover by 2.2% from 11% to 8.8% as an example and assume that the tropical moist region follows the area of cloudy moist air with the same relative decrease. Based on the 3.5-box model with the parameters in LCH, the changes of the radiative energy budget at the TOA are -2 Wm^{-2} due to water vapor, -0.84 Wm^{-2} due to high clouds, and -0.7 Wm^{-2} due to low clouds. In the calculations, we assume overlap of cirrus and low clouds for the high cloud region, following LCH. The total radiative energy budget change (-3.54 Wm^{-2}) obtained here is responsible for a feedback factor of -1.03 by using the relation between the high cloud cover and SST from LCH. Note that the water vapor effect corresponding to the change of moist area can be from 0 to -2 Wm^{-2} , depending on how closely the area of the moist region follows the area of cloudy moist air. Therefore, the feedback factor in LCH ranges from about -0.45 to -1.03 .

Since the values of the parameters used in the 3.5-box model in LCH are not completely constrained by observations, in what follows we will examine the dependence of $\Delta R_{\text{net}}/\Delta A$ on the choices of model parameters. In particular, we will use additional information derived from modeling and observations to further constrain the contrast in OLR between dry and moist regions and the net radiative effects of high and low clouds.

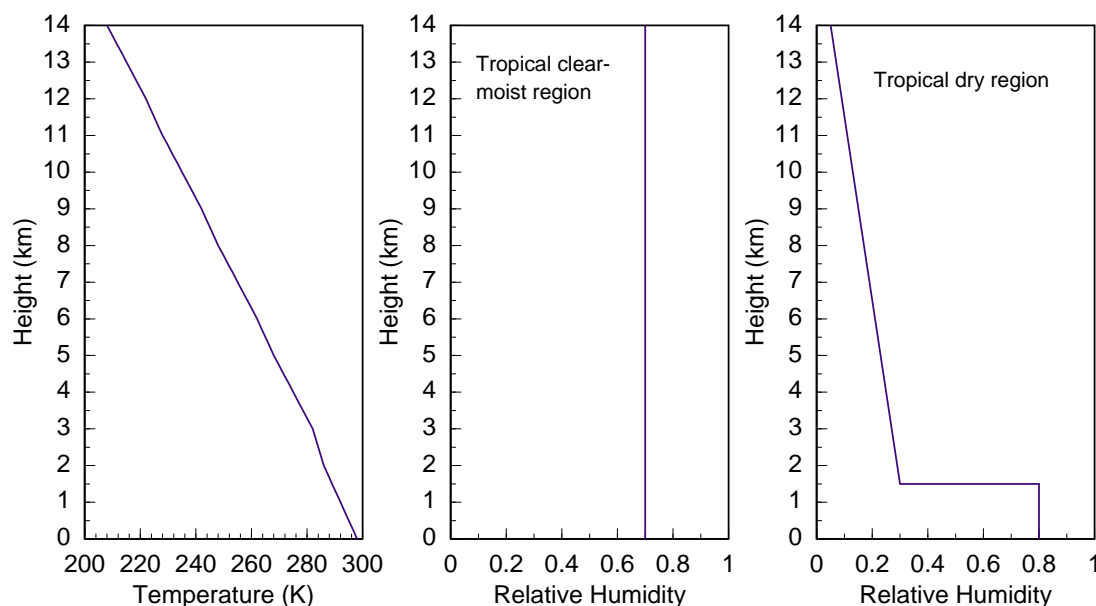


Fig. 1. Tropical atmospheric profiles used in clear-sky OLR simulations: (a) temperature; (b) relative humidity for tropical clear-moist region; and (c) relative humidity for tropical dry region.

3 Radiative effects of tropical water vapor in the iris feedback

The water vapor feedback related to changing the area of the tropical moist region is proportional to the difference in the OLR between the tropical dry and clear-moist regions. The OLRs used in LCH for these two regions are 303 and 263 Wm^{-2} , which result in a difference of 40 Wm^{-2} . The OLR for the cloudy-moist region is set to 138 Wm^{-2} to match the ERBE full sky OLR for the Tropics.

In order to examine the OLR in the tropical regions, we use a radiation model developed by Fu and Liou (1993). In this model, the radiative transfer scheme used is the delta-four-stream approximation. Gaseous absorption due to H_2O , CO_2 , O_3 , CH_4 , and N_2O is incorporated into the scattering model by employing the correlated k-distribution method (Fu and Liou, 1992). The H_2O continuum absorption is included by using CKD2.2 (Charlock et al., 1999).

We use the tropical standard temperature profile (McClatchey et al., 1971) with modification to be consistent with the sea surface temperature of 298 K. The relative humidity for the moist region is assumed to be 70% from the surface to 14 km, which is in rough agreement with observations (Brown and Zhang, 1997; LCH). For the tropical dry region, we use a relative humidity of 80% from the surface to 1.5 km, and decreasing linearly from 30% to 5% from 1.5 to 14 km (Larson et al., 1999; Manabe and Wetherald, 1967). Above 14 km, the tropical standard water vapor profile is used. Figure 1 shows our assumed temperature and relative humidity profiles below 14 km.

The OLRs from the radiation model are 293 Wm^{-2} and

268 Wm^{-2} , respectively, for the tropical dry and clear-moist regions. To match the ERBE full sky OLR for the Tropics, we require an OLR of 154 Wm^{-2} for the cloudy-moist region, as opposed to 138 Wm^{-2} in LCH. The OLR difference between the dry and clear-moist regions is then 25 Wm^{-2} , which is 15 Wm^{-2} lower than that derived from LCH. Our sensitivity study shows that in order to produce the OLR of 303 Wm^{-2} for the clear-dry region as used in LCH, the free-tropospheric relative humidity must be less than $\sim 10\%$. For their OLR of 263 Wm^{-2} in the tropical clear moist region the relative humidity must be as high as 80%.

As another way to estimate the OLR differences between clear moist and dry regions, we use the ECMWF water vapor data. The mean relative humidity profiles over tropical ocean are obtained for different regions that are defined based on their monthly-mean precipitation (Bretherton, private communication, 2001). Here an interval of 1 mm day^{-1} from 0 to 16 mm day^{-1} is considered. The monthly-mean precipitation of 3 mm day^{-1} roughly divides the tropics into 50% drier and 50% moister regions. Using averaged relative humidity profiles for these two regions (see Fig. 2), we obtain an OLR difference of 15 Wm^{-2} . This calculation might underestimate the OLR difference since the monthly-mean ECMWF water vapor data might smear out the sharp transition between the moist and dry regions in the instantaneous fields. Therefore, we also use the relative humidity profiles for the areas with the monthly-mean precipitation smaller (larger) than 1 mm day^{-1} (15 mm day^{-1}), corresponding to the extreme dry (moist) region (see Fig. 2). In this case, we derive an OLR difference of 23 Wm^{-2} , which is 2 Wm^{-2} smaller than the value derived using the profiles in Fig. 1. Thus the

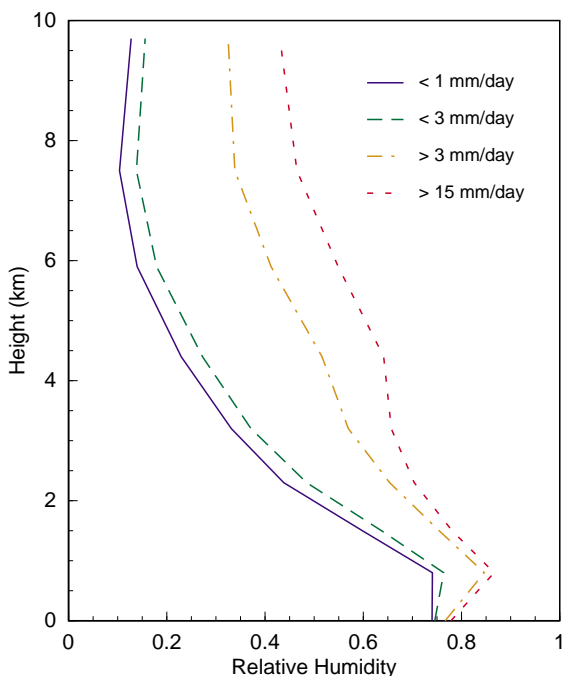


Fig. 2. Tropical ocean ECMWF relative humidity profiles averaged over regions with different monthly-mean Xie-Arkin precipitation categories. The water vapor profiles were based on all monthly means for the 15 years (1979–1993) of the ECMWF reanalysis ERA15 (C. Bretherton, private communication, 2001). The monthly-mean surface precipitation smaller and larger than 3 mm day^{-1} roughly divided the tropics into 50% drier and 50% moister regions.

OLR difference between clear moist and dry regions used in the 3.5-box model can not be larger than $\sim 25 \text{ Wm}^{-2}$.

ERBE observations suggest a clear-sky OLR difference of about 20 Wm^{-2} between moist and dry regions (e.g. Harrison et al., 1990). For example, for the Western Pacific convective region (0–15 N, 120–150 E) and the Central Pacific “dry zone” (0–15 S, 130–150 W), we have clear-sky OLRs of 280 Wm^{-2} and 301 Wm^{-2} , respectively (Hartmann et al., 2001). Note that the SSTs are $\sim 302 \text{ K}$ for the convective region and $\sim 300 \text{ K}$ for the dry zone. In order to obtain the OLR differences due to water vapor only, we convert the observed clear-sky OLRs to those with the same temperature profile and a SST of 298 K , which gives 295 Wm^{-2} for the dry zone and 273 Wm^{-2} for the moist region. The conversion here is based on radiation model simulations by assuming that the relative humidity profiles do not change with the changes of the temperature profiles. Therefore, the OLR difference due to water vapor between moist and dry region based on ERBE observations is also smaller than 25 Wm^{-2} . Here we argue that the effect of water vapor on the area feedback is overestimated in LCH by at least 60%.

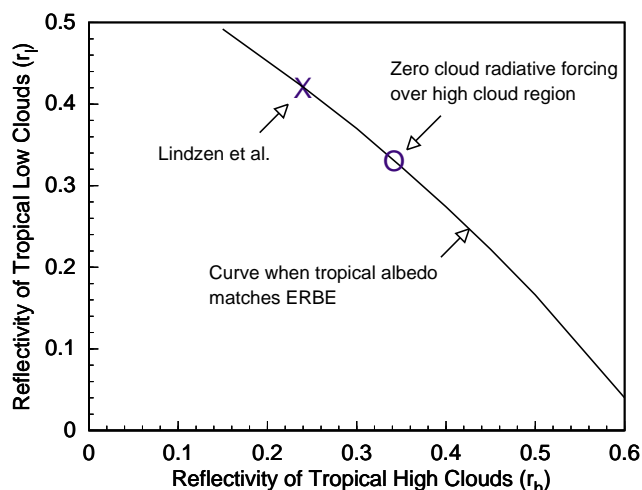


Fig. 3. Reflectivity of tropical low clouds versus reflectivity of tropical high clouds, required to match the tropical ERBE reflectivity in the 3.5-box model by LCH. The symbol “x” represents the values used in LCH while the symbol “O” represents the values when the TOA cloud radiative forcing in the cloudy-moist region is zero.

4 Radiative effects of tropical clouds in the iris feedback

LCH used the values of a tropical reflectivity of 0.24 and a tropical clear sky reflectivity of 0.13 based on ERBE observations to constrain their model parameters related to solar radiation in the Tropics. Their choices, however, are not the only ones consistent with the ERBE data. Figure 3 shows r_l , the reflectivity of tropical low clouds, as a function of r_h , the reflectivity of high tropical clouds, required to match the ERBE tropical reflectivity. Other parameters in the calculations follow LCH. We can see that r_l decreases as r_h increases in order to maintain the constant mean reflectivity for the tropics. Figure 3 indicates that consistency with ERBE reflectivities is possible for a wide range of r_h and r_l . The symbol “x” represents the choice of tropical cloud reflectivities used in LCH. The cloud radiative effect on the proposed area feedback is very sensitive to the cloud reflectivities chosen in the model. Note that the relation between r_l and r_h shown in Fig. 3 does not depend on the emission temperatures. The question that arises now is, what are the r_h and r_l we should use? It is clear that more observational information is needed to choose appropriate reflectivities for the 3.5-box model. We turn for this information to observed cloud radiative forcing (CRF) at the TOA.

The CRF at the TOA is a useful quantity for understanding the role of clouds in the climate system. It is defined as the difference in the TOA net radiative fluxes (absorbed solar radiation minus outgoing longwave radiation) between all-sky and clear-sky conditions. CRF gives a measure of the effect of clouds on the radiative energy budget at the TOA. Based on LCH’s 3.5-box model with their parameters, the CRF in the tropical cloudy-moist region after considering the over-

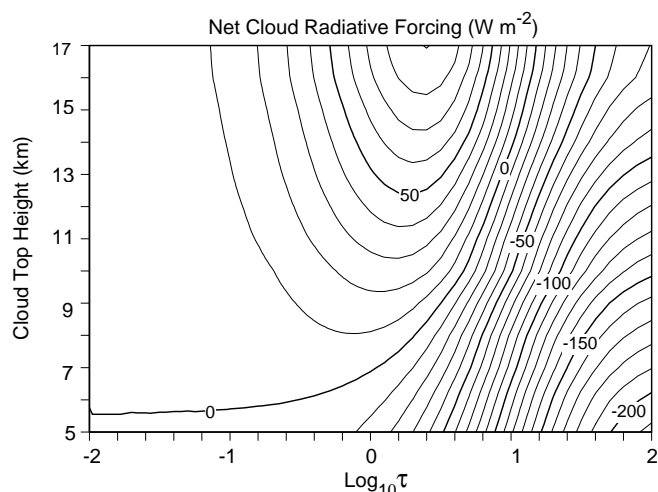


Fig. 4. Calculated Daily-averaged cloud radiative forcing (Wm^{-2}) for overcast high clouds as functions of cloud top height and cloud visible optical depth. The solar radiation is considered at equinox conditions at the equator.

lap of cirrus and low clouds is $+38 \text{ Wm}^{-2}$. By neglecting the greenhouse effect of low clouds, the CRF in the tropical clear-moist region due to LCH's low clouds is -32 Wm^{-2} . The positive cloud radiative forcing in cloudy-moist region and negative cloud forcing in the clear-moist region in LCH contribute powerfully to their negative feedback as the high cloud cover decreases with the increase of SST. Note that the calculated radiative forcing of high clouds depends on the emission temperatures assumed for the tropical regions through the longwave radiation. In the following, we will use observed CRF for the high cloud region and the OLRs we calculated in Sect. 3 to constrain the reflectivities of tropical high and low clouds.

The symbol “o” in Fig. 3 represents the reflectivities when the cloud radiative forcing for the tropical high cloud region (i.e. cloudy moist region) is zero. Here the high cloud reflectivity is increased to $r_h = 0.342$ and the low cloud reflectivity is reduced to $r_l = 0.331$. The cloud radiative forcing of the tropical cloudy moist region is positive (negative), when r_h is smaller (larger) than 0.342. Therefore the role of tropical high clouds in the area feedback can be either positive or negative, depending on the r_h used. We can constrain the value of r_h by known limits on the cloud radiative forcing in high cloud regions.

Since the low cloud forcing is always negative, low clouds always produce negative feedback associated with a decrease of high cloud cover. But this negative feedback effect decreases as the r_h increases, if we constrain the tropical mean reflectivity to match ERBE observations. We will discuss the low cloud effects more later.

4.1 Tropical high cloud effects

The shortwave effect of clouds cools the earth-atmosphere system by reflecting the solar radiation back to space, while their longwave effect warms the system by trapping the infrared radiation emitted from the warmer underlying atmosphere and surface. For tropical high clouds, their longwave effect can dominate their shortwave effect or vice versa, which would yield a positive or negative cloud radiative forcing.

Figure 4 shows the daily-mean cloud radiative forcing for tropical overcast cirrus as functions of cloud top height and cloud optical depth, calculated using the radiation model that was developed by Fu and Liou (1993) and modified by Fu (1996) and Fu et al. (1998). The temperature, water vapor, and ozone profiles are those of the tropical standard atmosphere (McClatchey et al., 1971). The mean effective size of ice particles in the cirrus is assumed to be $50 \mu\text{m}$ (Fu, 1996) and the ocean surface albedo is 0.05. The solar radiation is set at equinox conditions at the equator.

Figure 4 indicates that the radiative forcing due to cirrus clouds can range from positive to negative, depending on the cloud optical depth and cloud top height. For given cloud top height, the cloud radiative forcing is usually positive for very small optical depths and increases with the optical depth. As the optical depth continues to increase, the forcing reaches a maximum and then decreases to zero and becomes negative for large optical depths. Figure 4 also shows that for given cloud optical depth, the cloud forcing increases with the cloud top height. It is worth mentioning here that the widespread subvisible thin cirrus clouds in the tropics have an optical depth less than 0.03, which have a net cloud radiation forcing smaller than 4.5 Wm^{-2} from our calculation.

Observations based on ERBE data indicate that longwave and shortwave cloud forcing for tropical convective clouds nearly cancel (Ramanathan et al., 1989; Kiehl and Ramanathan, 1990); i.e. the total cloud radiative forcing is near zero. By analyzing ERBE and ISCCP data in a large region from 10° S to 10° N and 140° E to 90° W , Kiehl (1994) found that the longwave cloud forcing and shortwave cloud forcing are for the most part dependent on the high cloud amount. Kiehl (1994) explained the cancellation of the longwave and shortwave cloud forcing at the top of the atmosphere as a result of macro- and micro- physical properties of the prevalent anvil clouds.

Recently, Hartmann et al. (2001) used ERBE and ISCCP data in conjunction with the radiative transfer model (Fu and Liou, 1993; Fu, 1996; Fu et al., 1998) to estimate the effect of various cloud types on the TOA radiative budget in tropical convective region. The area considered is from $0\text{--}15^\circ \text{ N}$, $120\text{--}150^\circ \text{ E}$ during July and August 1985 and 1986, which can be considered a typical tropical moist region. Figure 5a shows cloud cover versus cloud top heights and visible optical depths for this region. The sum of numbers in the boxes for high clouds, which have cloud top heights higher than

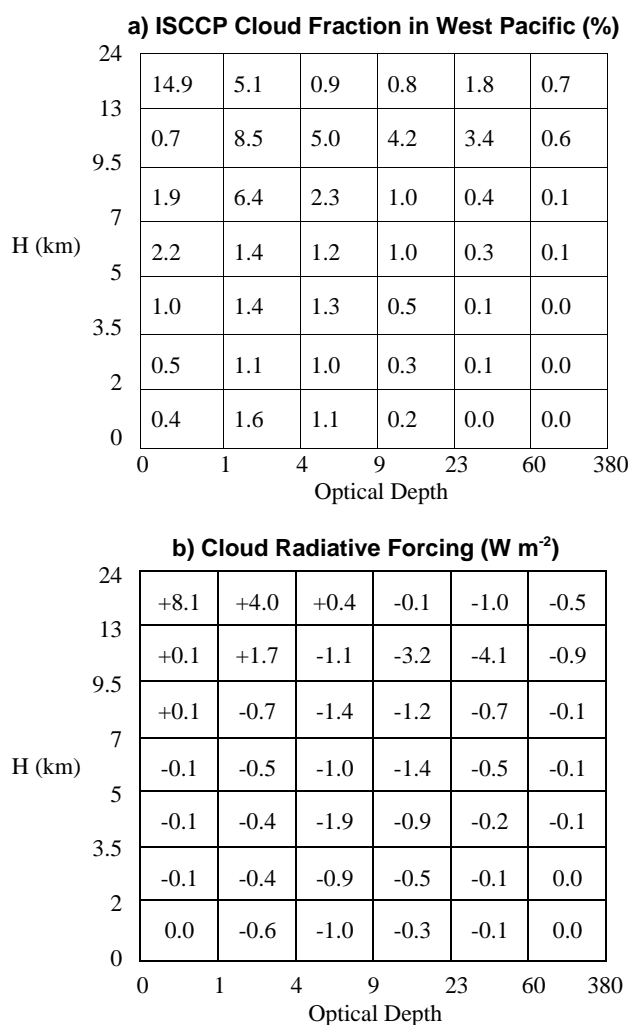


Fig. 5. (a) ISCCP cloud cover versus cloud top height and visible optical depth and (b) calculated TOA cloud radiative forcing for each cloud category, using cloud cover from the Fig. 5a, for the region 0–15 N, 120–150 E during July and August 1985 and 1986.

7 km, gives a total high cloud cover of about 0.6. Figure 5b shows modeled cloud radiative forcing for each of the ISCCP cloud categories, using the cloud cover values in Fig. 5a. The total cloud radiative forcing for this area is -11 Wm^{-2} from the radiation model with the ISCCP data as input and -7 Wm^{-2} from ERBE. Note that although the overcast cloud forcing for each individual cloud category can be strongly positive or negative (Fig. 4), the ensemble of cloud types that occurs in association with deep convection in the tropics arranges itself so that the individual positive and negative contributions cancel each other when averaged over the convective cloud system (Hartmann et al., 2001). From Fig. 5b, the total cloud radiative forcing due to high clouds, which is sum of numbers in relevant boxes, is -0.6 Wm^{-2} . Since the high cloud cover is 0.6, we infer that overcast cloud radiative forcing for the high cloud region is only -1 Wm^{-2} .

Therefore, we choose $r_h = 0.342$ and $r_l = 0.331$ for the 3.5-box model so that the cloud radiative forcing for the cloudy-moist region is zero, as observed. Since the feedback due to the direct radiative effect of high clouds related to area changes is proportional to the cloud radiative forcing for the high cloud region, we can argue that the direct radiative effect of the tropical high clouds cannot constitute an effective “infrared iris”.

4.2 Tropical low cloud effects

Given constant low cloud cover, tropical low clouds produce a negative feedback related to changing the relative area of the high cloud region. This feedback is proportional to the low cloud radiative forcing. Using a r_l of 0.331 with a cloud cover of 25%, and assuming no greenhouse effect of the low clouds, the cloud radiative forcing for the tropical low clouds based on the 3.5-box model becomes -24 Wm^{-2} . Since this is smaller than the low cloud radiative forcing of -32 Wm^{-2} in LCH, we argue that the negative feedback due to low clouds is also exaggerated in LCH.

Low cloud forcing in the trade cumulus regions adjacent to deep convection over the warm water may be even much smaller in magnitude than 24 Wm^{-2} , since ERBE estimates suggest a value less than 10 Wm^{-2} (Hartmann et al., 2001). The most reflective low clouds in the tropics are over the cooler SST in areas of upwelling, far removed from deep convection. Thus the assumption of constant low cloud cover in the simple 3.5-box model might significantly exaggerate the negative feedback related to boundary-layer clouds. Finally we note that Bajuk and Leovy (1998) suggested a negative correlation between the ocean temperature and the low cloud amount, which might introduce a positive feedback in the climate system. Further research is needed to examine the variation of tropical low cloud properties and its implication in climate sensitivity.

5 Conclusions and discussions

Using a simple 3.5-box radiative-convective model, LCH obtained a feedback factor between -0.45 and -1.03 related to the changes of tropical high cloud area with changes of SST. This feedback factor can be attributed to the radiative effects of water vapor (0 to -0.58), high clouds (-0.25), and low clouds (-0.20). In this paper, we argue that the contribution of tropical high clouds to the feedback process would be small since the radiative forcing over the tropical high cloud region is near zero and not strongly positive as LCH assume. It is also shown that the water vapor and low cloud effects are overestimated by at least 60% and 33%, respectively, in LCH. Using the model of LCH with our revised parameters we obtain a feedback factor ranging from -0.15 to -0.51 due to the radiative effects of water vapor (0 to -0.36) and low clouds (-0.15). It is noted that our feedback factor could

still be overestimated due to the assumption of constant low cloud cover in the simple radiative-convective model.

In this paper we have assumed that the assumption of 22% high cloud cover decrease for a 1°C increase in SST from LCH is valid. However, Hartmann and Michelsen (2001) pointed out that the negative correlation between cloud-weighted SST and high cloud fraction derived by LCH from observations within the domain considered by them is not really evidence that tropical anvil cloud area decreases with the increase of SST. Further research is needed to investigate the relation between high cloud cover and SST. It should be noted that for ANY rate of changes of cloudy- and clear-moist areas with changing SST, the feedback due to such a change will be significantly smaller than that suggested by LCH.

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