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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 of sea surface temperature increase 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 should be small. Although not mentioned by Lindzen et al, 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., 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.

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. 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 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 about -1.03. Such a large negative feedback would more than cancel all the positive feedbacks in the more sensitive current climate models. 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.

In the iris effect suggested by LCH, the cloud and water vapor feedbacks are inextricably tied to each other. Based on the calculations by LCH, the feedback factor due to water vapor is negative and lies between 0 and -0.58 depending on how closely the moist area follows the cirrus cloud area. The feedback factor due to clouds is -0.45. 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, the effect of tropical high clouds on the OLR is much larger than their effect on reflected solar radiation. Second, 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 acknowledged in LCH. LCH assume that low clouds are everywhere in tropics with a constant cloud cover, a 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 clouds and water vapor and discuss how the issues related to these arguments affect their main conclusions on climate sensitivity. We find on the basis of calculations and observations that the contrast in clear-sky OLR between moist and dry regions is overestimated by at least 60% in LCH. Their strong negative feedback due to cloud changes is related as much to the low clouds as to the high clouds, a fact not mentioned by LCH. Observations suggest that the net radiative effect of high clouds is near zero and the net radiative effect of low clouds is smaller than assumed by LCH. These two changes give a much reduced magnitude for the negative feedback associated with the radiative effect of clouds. 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

$$\Delta T = G_0(\Delta Q + F\Delta T) \tag{1}$$

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

$$\Delta T = G_0\Delta Q/(1 - G_0F) = \Delta T_0/(1 - f) \tag{2}$$

where $\Delta T_0 \equiv G_0\Delta Q$ is the response of the climate system to the forcing in the absence of feedbacks and $f \equiv G_0F$ is referred to as the feedback factor. In the iris mechanism, F is the product of two parts. The first part is the reduction of the high cloud fraction per degree Kelvin of SST, $\Delta A/\Delta T_{\mathrm{s}}$, which is assumed to be 22%. The other part is the change in the TOA radiation balance with a change in high cloud cover,$\Delta R_{\mathrm{net}}/\Delta A$, 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, $F = (\Delta A/\Delta T_{\mathrm{s}})\Delta R_{\mathrm{net}}/\Delta A$.

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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 W m^{-2}, 263 W m^{-2} , and 138 W m^{-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 top of the atmosphere (TOA) radiative energy budget change in response to changes of high cloud area. Consider a decrease of high cloud cover by 20% 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 W m^{-2} due to water vapor, -0.84 W m^{-2} due to high clouds, and -0.7 W m^{-2} due to low clouds. In the calculations, we assume overlap of high and low clouds for the high cloud region, following LCH. The total radiative energy budget change (-3.54 W m^{-2}) obtained here is responsible for a feedback factor of

-1.03 derived 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 W m^{-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 choices of all 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.

3. Radiative effects of tropical water vapor in the iris feedback

The response of the climate system to changing the area of the clear-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 W m^{-2}, which result in a difference of 40 W m^{-2}. The OLR for the cloudy-moist region is set to 138 W m^{-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 W m^{-2} and 268 W m^{-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 W m^{-2} for the cloudy-moist region. The OLR difference between the dry and clear-moist regions is then 25 W m^{-2}, which is 15 W m^{-2} lower than that derived from LCH. Our sensitivity study shows that in order to produce the OLR of 303 W m^{-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 W m^{-2} in the tropical moist region the relative humidity must be as high as 80%.

The OLR differences between moist and dry regions are also examined using 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 precipitation (Bretherton, private communication, 2001). Here an interval of 1 mm day^{-1} from 0 to 16 mm day^{-1} is considered. Using the relative humidity profile for the areas with the monthly precipitation smaller (larger) than 1 mm day^{-1} (15 mm day^{-1}), corresponding to the extreme dry (moist) region, we derive an OLR difference of 23 W m^{-2}, which is 2 W m^{-2} smaller than the value given by our calculation. Thus the OLR difference between moist and dry regions used in the 3.5-box model should not be larger than $\sim$25 W m^{-2}.

ERBE observations suggest a clear-sky OLR difference of about 20 W m^{-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 W m^{-2} and 301 W m^{-2}, respectively (Hartmann et al., 2001). Note that the SSTs are $\sim$302 K for the convective region and $\sim$300 K for the dry zone. In order to compare the OLR differences obtained from the model and observations, we convert the observed clear-sky OLRs to those with a SST of 298 K, which gives 295 W m^{-2} for the dry zone and 273 W m^{-2} for the moist re-

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gion. 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 regions based on ERBE observations is also smaller than 25 W m^{-2}. Here we argue that the effect of water vapor on the area feedback is overestimated in LCH by at least 60%.

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 2 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 mean reflectivity for the Tropics. Figure 2 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 reflecitvities 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. 2 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.

Cloud radiative forcing at the TOA is a useful quantity for understanding the role of clouds in the climate system. The cloud radiative forcing is defined as the difference in the net radiative fluxes at the TOA between cloudy-sky and clear-sky conditions. (The net radiation here is the difference between absorbed solar radiation and outgoing longwave radiation at the TOA.) Cloud radiative forcing 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 cloud radiative forcing for the tropical overcast high cloud region is +38 W m^{-2}. The cloud radiative forcing for LCH's tropical low clouds with a cloud cover of 25% is -32 W m^{-2}. The positive radiative forcing of high clouds and negative forcing of low clouds assumed in LCH contribute powerfully to the negative feedback that they obtain. 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 discuss the reflectivities of tropical high and low clouds by using the emission temperatures based on our new clear-sky OLRs.

The symbol "o" in Fig. 2 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 $r_\mathrm{h} = 0.342$ and the low cloud reflectivity is reduced to $r_\mathrm{l} = 0.331$. 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 and negative cloud radiative forcing, respectively.

Figure 3 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 μm (Fu, 1996) and the ocean surface albedo is 0.05. The solar radiation is set at equinox conditions at the equator.

Figure 3 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 3 also shows that for given cloud optical depth, the cloud forcing increases with the cloud top height.

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 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 0–15° N, 120–150° E during July and August 1985 and 1986, which can be considered a typical tropical moist region. Figure 4a 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 7 km, gives a total high cloud cover of about 0.6. Figure 4b shows modeled cloud radiative forcing for each of the ISCCP cloud categories, using the cloud cover values in Fig. 4a. The total cloud radiative forcing for this area is -11 W m^{-2} from the radiation model with the ISCCP data as input and -7 W m^{-2} from ERBE. Note that although the overcast cloud forcing for each individual cloud category can be strongly positive or negative (Fig. 3), 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. 4b, the total cloud radiative forcing due to high clouds, which is sum of numbers in relevant boxes, is -0.6 W m^{-2}. Since the high cloud cover is 0.6, we infer that overcast cloud radiative forcing for the high cloud region is only -1 W m^{-2}.

Therefore, we choose $r_{\mathrm{h}} = 0.342$ and $r_{\mathrm{l}} = 0.331$ 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 W m^{-2}. Since this is smaller than the low cloud radiative forcing of -32 W m^{-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 W m^{-2}, since ERBE estimates suggest a value less than 10 W m^{-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. This calls into question the assumption of constant low cloud cover in the simple 3.5-box model.

5. Conclusions and Discussion

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 rather than −0.45 to −1.03.

In this paper we have assumed that the result 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 average high cloud fraction derived by LCH from observations within the domain considered by them is not really evidence that tropical cloud anvil area is inversely proportional to 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.

One final note we would like to make here is about the tropical low clouds. In LCH, the low cloud cover is assumed to be constant, which results in a significant contribution to the negative feedback they find. ERBE estimates suggest a low cloud forcing in the trade cumulus regions adjacent to deep convection over the warm water (Hartmann et al., 2001). The most reflective low clouds in the tropics are over the cooler SST, far removed from deep convection. This calls a question on the assumption of constant low cloud cover in the simple 3.5-box model. On the other hand, 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 also needed to examine the variation of tropical low cloud properties and its implication in climate sensitivity.

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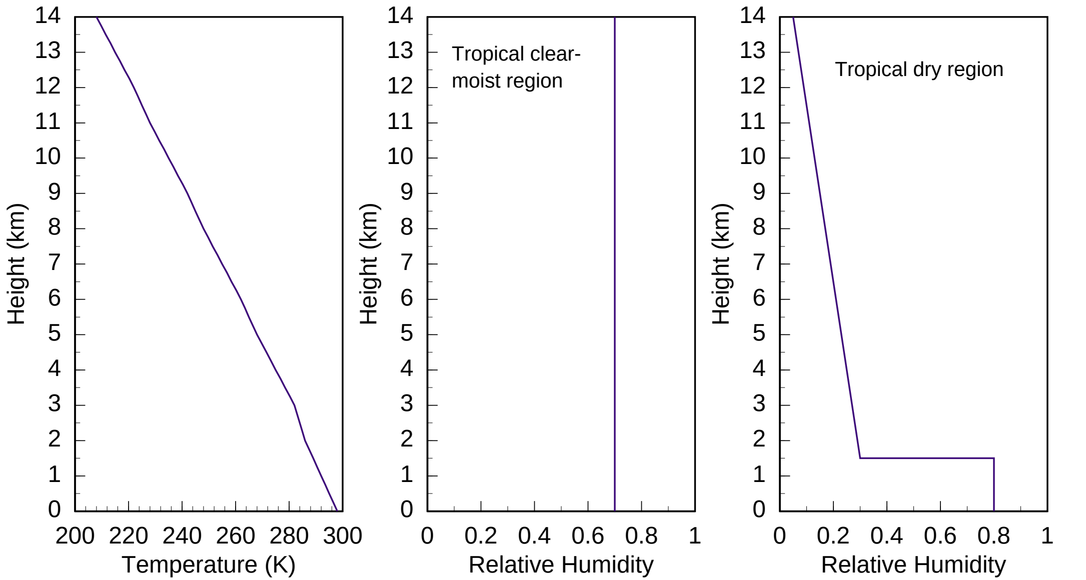

Fig. 1. Tropical atmospheric profiles used in clear-sky OLR simulations: **(a)** temperature; **(b)** relative humidity for tropical clear-moist region; and **(c)** relative humidity for tropical dry region.

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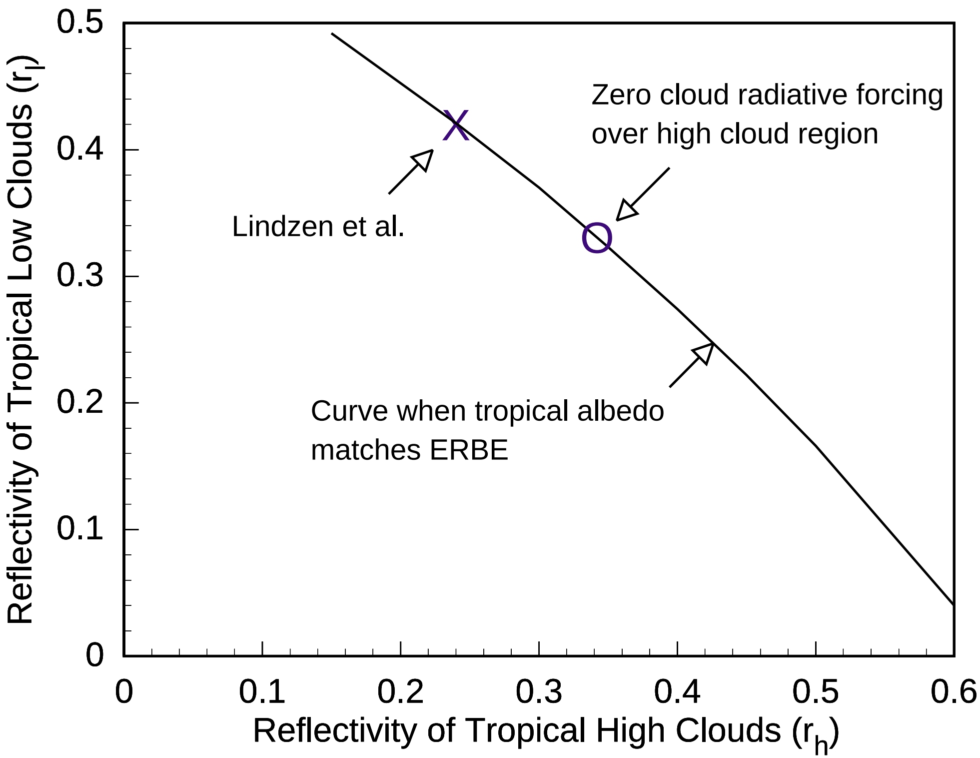

Fig. 2. Reflectivity of tropical low clouds versus reflectivity of tropical high clouds, required to match the tropical ERBE reflectivity in the 3.5-box model by LCH. The symbol "x" represents the values used in LCH while the symbol "O" represents the values when the TOA cloud radiative forcing in the cloudy-moist region is zero.

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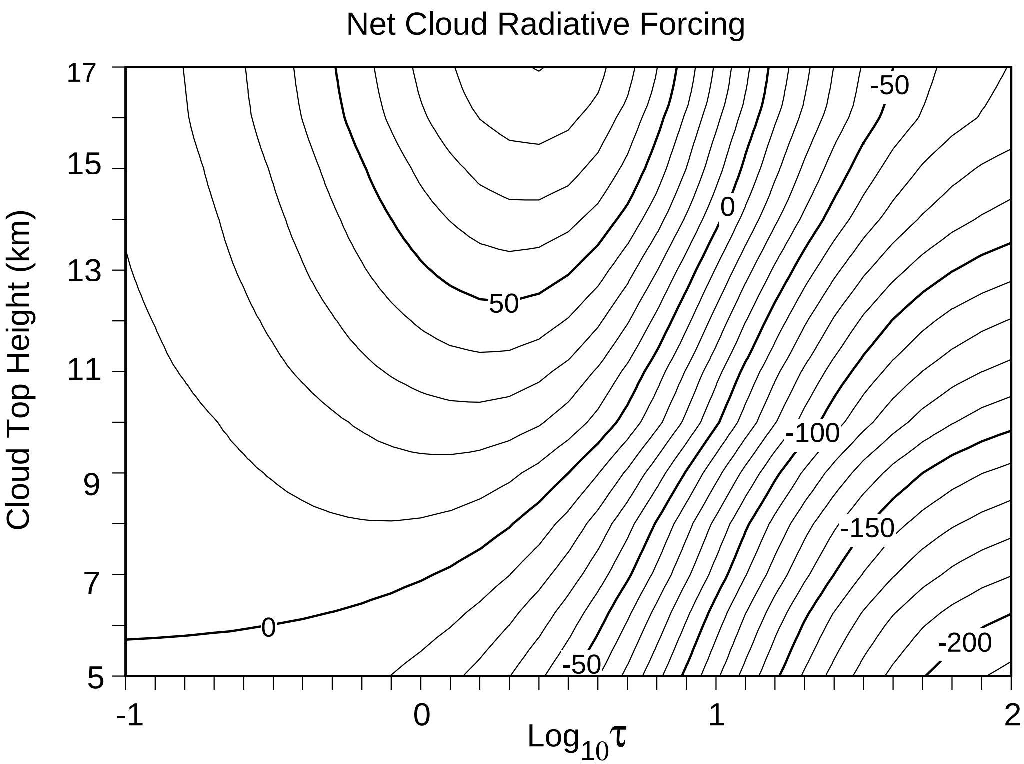

Fig. 3. Calculated daily-averaged cloud radiative forcing (W m^{-2}) for overcast high clouds as functions of cloud top height and cloud visible optical depth. The solar radiation is considered at equinox conditions at the equator.

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a) ISCCP Cloud Fraction in West Pacific (%)

H (km) \ Optical Depth	0–1	1–4	4–9	9–23	23–60	60–380
13–24	14.9	5.1	0.9	0.8	1.8	0.7
9.5–13	0.7	8.5	5.0	4.2	3.4	0.6
7–9.5	1.9	6.4	2.3	1.0	0.4	0.1
5–7	2.2	1.4	1.2	1.0	0.3	0.1
3.5–5	1.0	1.4	1.3	0.5	0.1	0.0
2–3.5	0.5	1.1	1.0	0.3	0.1	0.0
0–2	0.4	1.6	1.1	0.2	0.0	0.0

b) Cloud Radiative Forcing ($W\ m^{-2}$)

H (km) \ Optical Depth	0–1	1–4	4–9	9–23	23–60	60–380
13–24	+8.1	+4.0	+0.4	-0.1	-1.0	-0.5
9.5–13	+0.1	+1.7	-1.1	-3.2	-4.1	-0.9
7–9.5	+0.1	-0.7	-1.4	-1.2	-0.7	-0.1
5–7	-0.1	-0.5	-1.0	-1.4	-0.5	-0.1
3.5–5	-0.1	-0.4	-1.9	-0.9	-0.2	-0.1
2–3.5	-0.1	-0.4	-0.9	-0.5	-0.1	0.0
0–2	0.0	-0.6	-1.0	-0.3	-0.1	0.0

Fig. 4. **(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 Fig. 4a, for the region 0–15 N, 120–150 E during July and August 1985 and 1986.

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