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Comments on "Tropical convection and the energy balance at the top of the atmosphere"

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Comments on
"Tropical convection and the energy balance at the top of the atmosphere"

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Popular Summary

Submitted to Journal of Climate

Analyses of the Earth Radiation Budget Experiment (ERBE) data show that the effects of clouds on the solar and thermal infrared radiation in the tropical deep convective regions have a similar magnitude but opposite signs.

This small difference in the effects of clouds on radiation led Hartmann et al. (2001) to conclude that the contrast in the net radiation at the top of the atmosphere between the convective and non-convective regions must also be small. However, we have found that the ERBE data do not generally show a small contrast in the radiation between the convective and non-convective regions, and the model used by Hartmann et al., therefore, seems unlikely to represent the real physical processes involving convection, radiation, and climate in an appropriate way.

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Analyses of the Earth Radiation Budget Experiment (Barkstrom, 1984) data show that the shortwave (SW) and longwave (LW) cloud radiative forcing (CRF) at the top of the atmosphere (TOA) in the tropical deep convective regions have a similar magnitude but opposite signs. The CRF is defined as the difference between the full-sky and clear-sky radiative fluxes. Hartmann et al. (2001) (referred as HMF hereafter) computed the CRF for individual cloud types using a radiative transfer model with the cloud information taken from the International Satellite Cloud Climate Project (ISCCP) (Rossow and Schiffer, 1991) data archive. They found that individually each cloud type had a large impact on the radiation at the TOA, but collectively the effect was small. From the analyses of the ERBE data, they claimed that the net CRF at the TOA is small not only over deep convective regions but also over the non-convective regions. The perceived small CRF led them to conclude that the contrast in the net TOA radiation between the convective and non-convective regions must also be small. HMF used a simple model to investigate the relation between convection and radiation in the tropics. They hypothesized that the small contrast in the TOA radiation between the convective and non-convective regions was due to the high sensitivity of circulation to the sea surface temperature (SST) gradient and the high sensitivity of cloud albedo to vertical velocity.

However, the ERBE data do not generally show a small contrast in the TOA radiation between the convective and non-convective regions, and the model used by HMF, therefore, seems unlikely to represent the real physical processes involving convection, radiation, and climate in an appropriate way.

Figure 1a shows the net radiation and CRF in the Pacific averaged over five years (1985-1989) for the equatorial band 10°S-10°N derived from the ERBE data. The net heating of the earth-atmosphere system attains a maximum value of ~80 $W\ m^{-2}$ west of 160°E and decreases continuously eastward to a value of ~50 $W\ m^{-2}$ at 260°E. The longitudinal distribution of the CRF (Figure 1b) is very similar to that of the net radiation. The CRF ranges from –5 $W\ m^{-2}$ west of the dateline to -32 $W\ m^{-2}$ at 260°E. The contrast between the deep convective region west of 160°E and the suppressed region to the east is 1: 1.6 for the net radiation and 1:6 for the CRF. These contrasts can hardly be considered as small. Averaged over the five ERBE years (1985-1989), the full-sky net TOA radiation in the tropics (30°S-30°N) is 48.6 $W\ m^{-2}$, the clear-sky net radiation is 61.1 $W\ m^{-2}$, and the CRF is -12.5 $W\ m^{-2}$. Thus, the CRF in the tropics as a whole is significant.

HMF suggested that the small CRF in the deep convective region as found by Kiehl (1994) is not a coincidece but is a combined effect of an ensemble of various types of clouds occurring at various heights and with a large range of optical thickness. They further suggested that the small CRF is a result of the feedback between convection, cloud albedo, and the SST gradient. In fact, the small CRF in the deep convective region is easy to explain. Clouds have the effects of reflecting SW radiation and reducing OLR. Except for thin clouds at high altitudes, the SW CRF in the tropics is greater than the LW CRF. This can be seen in Fig. 1b that the net CRF at 260°E is large, approximately $-30\ W\,m^{-2}$. This region is in the descending branch of the Walker circulation where the deep convection is suppressed and the boundary-layer clouds are extensive. For those boundary-layer clouds, the contrast between the cloud-top temperature and the SST is small, and so is the LW CRF. On the other hand, those clouds are highly reflective to SW radiation, and the SW CRF is large. The result is a large negative CRF. Westward from 260°E, the boundary-layer depth and, hence, the cloud height increase, but the cloud cover decreases. The increase in cloud height enhances the difference between the cloud temperature and the SST, which leads to an enhanced LW CRF. Thus, the net CRF decreases westward toward the dateline. West of the dateline is the deep convective region, where high-level clouds are extensive. The CRF is small, within $-10\ W\,m^{-2}$.

In the simple model of HMF, the tropics is divided into a convective region and a non-convective region. Each of these two regions is characterized by the SST, T, the net radiation at TOA, R, and the areal extent, A. The rate of change of the SST is given in their model by

$$c\frac{dT_1}{dt} = R_1 - \frac{M}{A_1}c_p(\theta_{e1} - \theta_{e2}) + Exp_1 \tag{1}$$

$$c\frac{dT_2}{dt} = R_2 + \frac{M}{A_2}c_p(\theta_{e1} - \theta_{e2}) + Exp_2 \tag{2}$$

where c is the heat capacity of the ocean mixed layer, M is the mass flux between the two regions, c_p is the heat capacity of air, θ_e is the saturation potential temperature of the surface air, Exp is the export of heat out of the domain of interest, and the subscripts 1 and 2 denote the convective and non-convective regions, respectively.

Assuming that exports of heat out of the convective and non-convective regions are the same, equilibrium solutions for T_1, T_2, R_1, and R_2 are obtained by time integration of Eqs. (1) and (2), so that

$$R_1 - R_2 = Mc_p(\theta_{e1} - \theta_{e2})\left(\frac{1}{A_1} + \frac{1}{A_2}\right) \tag{3}$$

Thus, the imbalance between R_1 and R_2 equals the transport of heat from the convective region to the non-convective region.

The mass flux between the two regions, which is related to the vertical pressure velocity ω, is given by

$$M = A_2 \frac{\omega_2}{g} = -A_1 \frac{\omega_1}{g} \tag{4}$$

where g is the gravitational acceleration. The convective area fraction A_1 is set to be 0.3. The vertical pressure velocity in the convective region is assumed to be linearly proportional to the temperature contrast between the convective and non-convective regions,

$$\omega_1 = -\gamma(T_1 - T_2) = -\gamma\Delta T \tag{5}$$

where γ is a disposable constant.

The net radiation at TOA is given by

$$R_i = S(1-\alpha_i) - F_i \tag{6}$$

where S is the insolation at TOA, α is the albedo, F is the OLR, and i=1 or 2. The OLR in each region is assumed to vary only slightly with the SST and is practically fixed at 190 W m^{-2} and 280 W m^{-2} in the convective and non-convective regions, respectively. The albedo in the non-convective region α_2 is fixed at 0.1, and the albedo in the convective region is assumed to be a function of vertical pressure velocity and SST,

$$\alpha_1 = \alpha_2 - \lambda\omega_1 q^*(T_1) \tag{7}$$

where λ is a disposable constant, and q^* is the saturation water vapor mixing ratio.

Equation (5) has the convection and, hence, atmospheric overturning increasing with increasing temperature contrast between the two regions. Equation (7) has the cloud albedo in the convective region increasing with increasing convection for a positive λ. As α_1 increases with increasing T_1, the net radiation R_1 will decrease, which will then lead to a decrease in both T_1 and the temperature contrast between the two regions. The decrease in the temperature contrast will cause weaker convection, which, according to Equation (7), will reduce the albedo α_1 for a positive λ. Thus, it constitutes a negative feedback loop between SST, convection, cloud albedo, and radiation. The strength of the feedback is dependent upon the values of γ and λ.

From Equation (7), we have

$$\lambda = -\frac{1}{q^*(T_1)} \frac{\Delta\alpha_1}{\Delta\omega_1} \tag{8}$$

To estimate the albedo sensitivity parameter λ, HMF derived values of $\Delta\alpha_l$ and $\Delta\omega_l$ by taking the albedo difference and the vertical velocity difference between the convective and non-convective regions,

$$\Delta\alpha_1 = \alpha_1 - \alpha_2 \tag{9}$$

$$\Delta\omega_1 = \omega_1 - \omega_2 \tag{10}$$

so that

$$\lambda = -\frac{1}{q^*(T_1)}\left(\frac{\alpha_1 - \alpha_2}{\omega_1 - \omega_2}\right) \tag{11}$$

By assuming an albedo contrast of 0.25 between the convective and non-convective regions, they found

$$\frac{\alpha_1 - \alpha_2}{\omega_1 - \omega_2} = -1.25 \times 10^2 \, (Pa\, s^{-1})^{-1} \tag{12}$$

and λ=5x10^3 $(Pa\ s^{-1})^{-1}$ for q^*=0.02. Thus, λ has a positive value, and the albedo in the convective region increases as convection strengthens.

Since, as applied by HMF, $\Delta\alpha_l$ and $\Delta\omega_l$ are the changes in albedo and vertical velocity of the convective region between either two different climate states or two different times, Equations (9) and (10) are of course incorrect. Instead, the correct representations should be

$$\Delta\alpha_1 = \alpha_1{}^a - \alpha_1{}^b \tag{13}$$

$$\Delta\omega_1 = \omega_1{}^a - \omega_1{}^b \tag{14}$$

where the superscripts a and b denote two different climate states. Therefore, the sensitivity of albedo to the vertical velocity estimated by HMF based on the differences between the convective and non-convective regions is unrelated to the sensitivity given in Equation (8). Equation (8) considers only the albedo and vertical velocity in the convective region. There is no evidence that albedo in the convective region increases with increasing vertical velocity as HMF claimed.

From Equations (5) and (7), HMF estimated the sensitivity of albedo to SST from

$$\frac{\Delta\alpha}{\Delta T} = \lambda\gamma q^* \tag{15}$$

Based on the results of Ramanathan and Collins (1991) on the relation between the interannual variations of the SST and the absorption of solar radiation, HMF estimated that

$$\frac{\Delta\alpha}{\Delta T} = 0.04 \text{ K}^{-1} \tag{16}$$

and $\lambda\gamma \approx 2$ for q^*=0.02. They found that for $\lambda\gamma \approx 2$ the model predicted a strong feedback that brought the net radiation in the convective region close to that in the non-convective region.

There are problems with this analysis of the albedo sensitivity to SST. Firstly, Equation (15) is not correct. From Equations (5) and (7), the albedo sensitivity should be

$$\frac{\Delta\alpha_1}{\Delta(T_1 - T_2)} = \lambda\gamma q^* \tag{17}$$

so that the albedo in the convective region changes with the temperature contrast between the convective and non-convective regions.

Secondly, the sensitivity of albedo to SST cannot be derived by comparing the data associated with interannual variations. To demonstrate this point, we show in Figure 2 a scatter-plot that relates albedo to SST in the central equatorial Pacific (10°S–10°N, 170°E-150°W) during April 1985 and April 1987. These two months are the same as that used in Ramanathan and Collins (1991). Each point represents monthly mean albedo from ERBE data archive and SST from the National Centers for Environmental Prediction (NCEP) analysis (Reynolds and Smith, 1994). The spatial resolution for each point is 2.5°x2.5° latitude-longitude. The year 1987 is an El Nino, and the SST in the central and eastern equatorial Pacific was abnormally high. Figure 3 shows the longitudinal distribution of equatorial SST of these two months. Compared to April 1985, the SST in the central and eastern equatorial Pacific increased significantly by 1.0–1.4*°C* in April 1987, but the warm pool region (140°E–160°E) cooled slightly by ~0.2*°C*. Correspondingly, the center of deep convection shifted eastward by ~50° longitude from the maritime continents to the central equatorial Pacific (Chou, 1994). Associated with the eastward shift of the deep convection, Figure 2 shows that both albedo and SST in the central equatorial Pacific are larger in April 1987 than in April 1985. These changes are unrelated to the sensitivity given in Equation (17) and are irrelevant to the model of HMF. In fact, the central equatorial Pacific was in the descending domain of the Walker circulation during April 1985 but was in the ascending domain during April 1987. The albedo difference between these two months is equivalent to the incorrect equation (9) but not equivalent to the correct equation (13).

Strong updrafts in the convection cores constitute only a small fraction of the tropics. Most of the tropics are in the regions where the air is descending. Even the detrained thick anvil clouds are in the regions of descending motion. The simple model of HMF assigns a fractional area of 0.3 to the convective region. With this large extent of the convective region, it must include both the deep convection cores and the extended anvil and cirrus clouds, both optically thick and thin. It is not clear what the albedo of the convective region α_l in Equation (7) refers to. If it refers to the albedo of the deep convection cores, then the albedo can change very little because the cloud particle content and optical thickness are already high. If the convective region refers to the region comprising both the convection cores and the anvil clouds, then the albedo of this convective region might decrease with increasing SST and the gradient of SST due to the area response as suggested by Lindzen (1997). In Lindzen et al. (2001), the fractional cover of high-level clouds due to the detrainment of deep convective clouds decreases with increasing SST. Recently, Wielicki et al. (2002) reported that the satellite-inferred OLR in the tropics increases during the past decade. They suggested that this increase of OLR could not be accounted for by the change in clear-sky radiation and was due to a decrease in clouds. Results from those studies imply that the mean albedo in the convective region of the HMF model might not necessarily increase with increasing T_l and ω_l. Furthermore, those studies also imply that the change in high-level cloud cover could have a large impact on the LW radiation and, hence, climate. This LW impact was essentially ignored by HMF.

Finally, Equations (1) and (2) are incorrect. The terms involving the rate of SST change are put there for the purpose of obtaining equilibrium solutions by time integration. For equilibrium solutions, the R-terms balance the M-terms as shown in Equation (3). If the mass flux M is defined only for the atmosphere, then Equations (1) and (2) are correct only if the net surface heat flux is negligible in both the convective and non-convective regions. In this case the net radiative heating of the earth-atmosphere system R is the same as the net heating of the atmosphere. Although the net surface heating has been found to be small in the Pacific warm pool during northern hemispheric winter (Chou, et al., 2000), it is not generally so in other tropical regions and seasons. Otherwise the heat fluxes at the sea surface in the tropics would not be an issue in all long-term climate studies. Figure 4 shows the net downward radiation at the top of the atmosphere, the net surface heating, and the net atmospheric heating in the region (30°S-30°N, 100°E-170°W) during January–August 1998. The TOA net radiation is taken from the Tropical Rain Measuring Mission (TRMM) Clouds and the Earth's Radiant Energy System (CERES) archive (Wielicki, et al., 1996). The radiation component of the net surface heating is inferred from Japan's Geostationary Meteorological Satellite radiance measurements (Chou et al., 2001). Whereas the surface turbulent heat fluxes are derived using the NCEP SST analysis and the Special Sensor Microwave/Imager (SSM/I) precipitable water and wind retrievals (Chou, et al., 1997). The atmospheric heating is the difference between the TOA radiation and the surface heating. It can be seen that the net surface heating has a significant spatial variation, ranging approximately from -60 $W\ m^{-2}$ to +60 $W\ m^{-2}$. As a result, the net heating of the atmosphere has little resemblance to the TOA radiation.

On the other hand, if the mass flux M is defined as the total mass transport in both the atmosphere and the ocean, then ω can no longer be the vertical pressure velocity of air. In this case, Equation (7) and all discussions of ω in HMF become irrelevant.

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Figure Captions

Figure 1. Longitudinal distributions of the net radiation (upper panel) and cloud radiative forcing (lower panel) in the equatorial Pacific (10°S–10°N) averaged over five years (1985-1989). The radiative fluxes are taken from the ERBE data archive.

Figure 2. A scatter-plot relating albedo to the sea surface temperature in the central equatorial Pacific (10°S–10°N, 170°E-150°W)) during April 1985 and April 1987. Each point represents monthly mean albedo from ERBE archive and SST from the NCEP analysis. The spatial resolution for each point is 2.5°x2.5° latitude-longitude.

Figure 3. The longitudinal distribution of the sea surface temperature in the equatorial Pacific (10°S–10°N) for April 1985 and April 1987. The SST is taken from the NCEP analysis.

Figure 4. Net downward fluxes at the top of the atmosphere and the surface, and the heating of the atmosphere averaged over the period January–August 1998. Units are $W\,m^{-2}$.

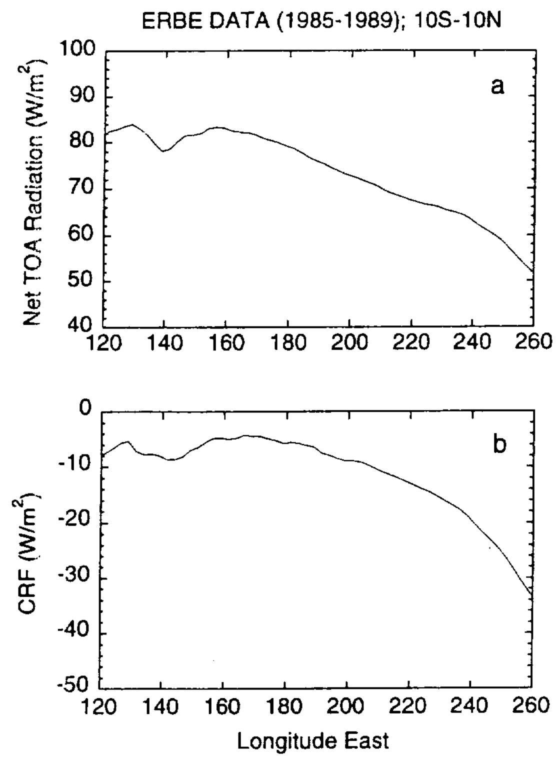
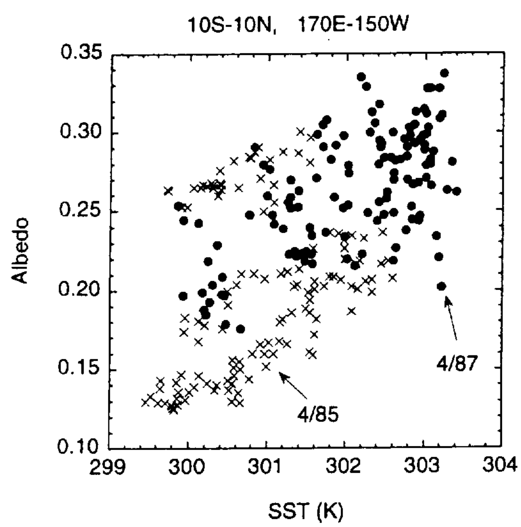

ERBE DATA (1985-1989); 10S-10N
Net TOA Radiation (W/m^2)
a
CRF (W/m^2)
b
Longitude East
100
90
80
70
60
50
40
0
-10
-20
-30
-40
-50
120
140
160
180
200
220
240
260

Figure 1



10S-10N, 170E-150W
0.35
0.30
0.25
0.20
0.15
0.10
Albedo
299
300
301
302
303
304
SST (K)
4/85
4/87

Figure 2

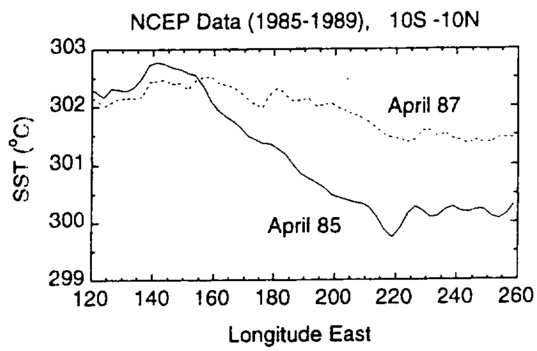

NCEP Data (1985-1989), 10S -10N
303
302
301
300
299
SST (°C)
April 87
April 85
120 140 160 180 200 220 240 260
Longitude East

Figure 3

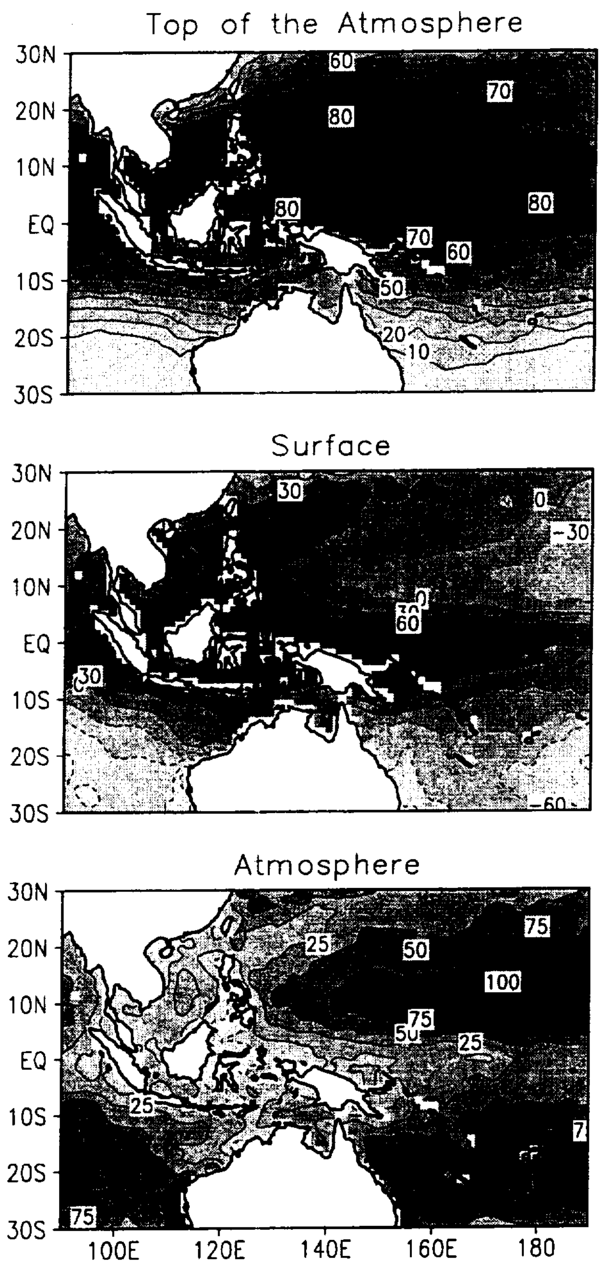

Top of the Atmosphere
30N
20N
10N
EQ
10S
20S
30S
60
70
80
80
80
70
60
50
20
10
Surface
30N
20N
10N
EQ
10S
20S
30S
30
0
-30
60
30
-60
Atmosphere
30N
20N
10N
EQ
10S
20S
30S
25
50
75
100
75
25
25
75
100E
120E
140E
160E
180

Fig. 4