Comments on "The Iris Hypothesis: A Negative or Positive Cloud Feedback?"

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Based on radiance measurements of Japan's Geostationary Meteorological Satellite (GMS), Lindzen et al. (2001, hereinafter LCH) found that the high-level cloud amount averaged over a large oceanic region in the western tropical Pacific decreases with increasing sea surface temperature (SST) underneath clouds. Using a 3.5-box radiative–convective model, they further found that the response of high-level clouds to SST had a negative climate feedback. Lin et al. (2002) reassessed this iris phenomenon by analyzing the radiation and clouds inferred from the Tropical Rainfall Measuring Mission (TRMM) Clouds and the Earth's Radiant Energy System (CERES) measurements. They found a weak positive feedback between high-level clouds and the surface temperature. The feedback factor ranges between 0.05 and 0.10 instead of between -0.55 and -1.10 as found by LCH. The difference in the feedback factor is due to a larger contrast in albedos and a smaller contrast in the outgoing longwave radiation (OLR) between the high-level cloudy region and the surrounding regions as derived by Lin et al. when compared with that specified in LCH. It appears that the approach taken by Lin et al. to estimate the albedo and OLR is not appropriate and that the inferred climate sensitivity is unreliable.

In LCH's model, the Tropics is divided into a dry region, a clear-moist region, and a cloudy-moist region. The fractional coverages of these regions are assumed to be 0.50, 0.28, and 0.22, respectively. The detrainment of cumulus clouds and the moistening of the upper troposphere have an effect of changing the tropical OLR and albedo. The net effect of cumulus detrainment on the radiation budget at the top of the atmosphere is dependent upon the OLR and albedo contrasts among the three regions, which is critical in estimating the feedback between high-level clouds and the surface temperature. LCH specified subjectively the OLR and albedo for the three regions while requiring that the mean OLR and albedo of the Tropics be consistent with the Earth Radiation Budget Experiment (Barkstrom 1984) inferred values.

A large OLR generally corresponds to a dry atmospheric condition. Following LCH's assumption that the dry region covers 50% of the Tropics, Lin et al. derived the OLR and albedo of the dry region by averaging the OLR and albedo of the 50% of the total TRMM CERES pixels in the Tropics that have the largest OLR. Also following LCH, they used a threshold temperature T_{11} of 260 K to infer high-level clouds, where T_{11} is the brightness temperature measured in the TRMM Visible and Infrared Scanner (VIRS) 11-μm window channel. They found that high-level clouds covered only 10% of the Tropics. The OLR and albedo of the cloudy-moist region were then assumed to be the mean values of OLR and albedo of those pixels with T_{11} of less than 260 K. Last, the OLR and albedo of the clear-moist region were assumed to be the mean OLR and albedo of all the pixels neither in the dry region nor in the cloudy-moist region. Values of OLR and albedo of the three tropical regions derived by Lin et al. using the TRMM CERES data are given in their Table 1.

The positive feedback factor, albeit small, as calculated by Lin et al. is due to the use of a large albedo of 0.51 for the cloudy-moist region, which can be compared with the albedo of 0.35 used by LCH. The albedo

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of the cloudy-moist region is a net effect of the reflection and, to a lesser extent, absorption by high-level clouds, low-level clouds, and the surface. Corresponding to these albedo values of the cloudy-moist region together with the albedos of low-level clouds and the surface specified in the model, the albedo of high-level clouds is 0.46 in Lin et al. and 0.24 in LCH. It can be seen in Table 1 of Lin et al. that the albedo contrast between the cloudy-moist region and the clear-moist region is 0.25 in Lin et al. and 0.14 in LCH. The large albedo contrast in Lin et al. causes a high sensitivity of the tropical albedo to high-level clouds. Thus, the response of the radiation at the top of the atmosphere to high-level clouds is more sensitive in the shortwave spectrum than in the longwave spectrum, resulting in a positive feedback between high-level clouds and the surface temperature. That is, a reduced high-level cloud cover due to a higher surface temperature causes a net heating of the Tropics and a further increase in the surface temperature.

LCH inferred areal coverage of high-level clouds using a threshold temperature of 260 K so that a GMS pixel was identified as being filled with high-level clouds if T_{11} is less than 260 K. This areal coverage of high-level clouds is merely an index for the extent of the detrainment of cumulus anvil clouds. It is not meant to be the total areal coverage of high-level clouds, which includes thin cirrus clouds. With a brightness temperature T_{11} of less than 260 K, it can be expected that the clouds are high and thick. Lin et al. found that the areal coverage of those pixels with T_{11} of less than 260 K is very small (10%) and that the mean albedo is very high (0.51), indicating a large optical thickness. Because the tropical high-level cloud cover is significantly greater than 10%, most of the high-level clouds cannot be detected by using a threshold temperature of 260 K.

LCH specified the areal coverage and albedo of the cloudy-moist region independent of the threshold temperature. The threshold temperature was used in LCH merely to investigate the sensitivity of high-level clouds to the SST beneath clouds, which is defined as the cloud-weighted SST. Besides the temperature of 260 K, some other threshold temperatures also can be used. Figure 1 shows the relation between the high-level cloud cover and the cloud-weighted SST in the tropical western Pacific (30°S–30°N, 130°E–170°W) for three threshold temperatures, 250, 260, and 270 K. The slope of the regression line S, the domain-averaged high-level cloud cover A, and the correlation coefficient R are shown in the figure. Each data point in the figure represents daily and domain-averaged values of cloud cover and the cloud-weighted SST. It covers a period of 20 months from January of 1998 to August of 1999. The results of using the three threshold temperatures to identify high-level clouds are similar. The change in cloud cover per 1°C change in the cloud-weighted SST is −13.5% for all three cases. As shown in Bell et al. (2002), the confidence level of the regression is very high. If Lin et al. used a threshold temperature of 250 K (270 K) instead of 260 K, the areal coverage and the albedo of the cloudy-moist region would be less (greater) than 10% and greater (less) than 0.51, respectively. Because the areal coverage of the cloudy-moist region changes with the threshold temperature, the OLR and albedo of the clear-moist region will also change with the thresh-

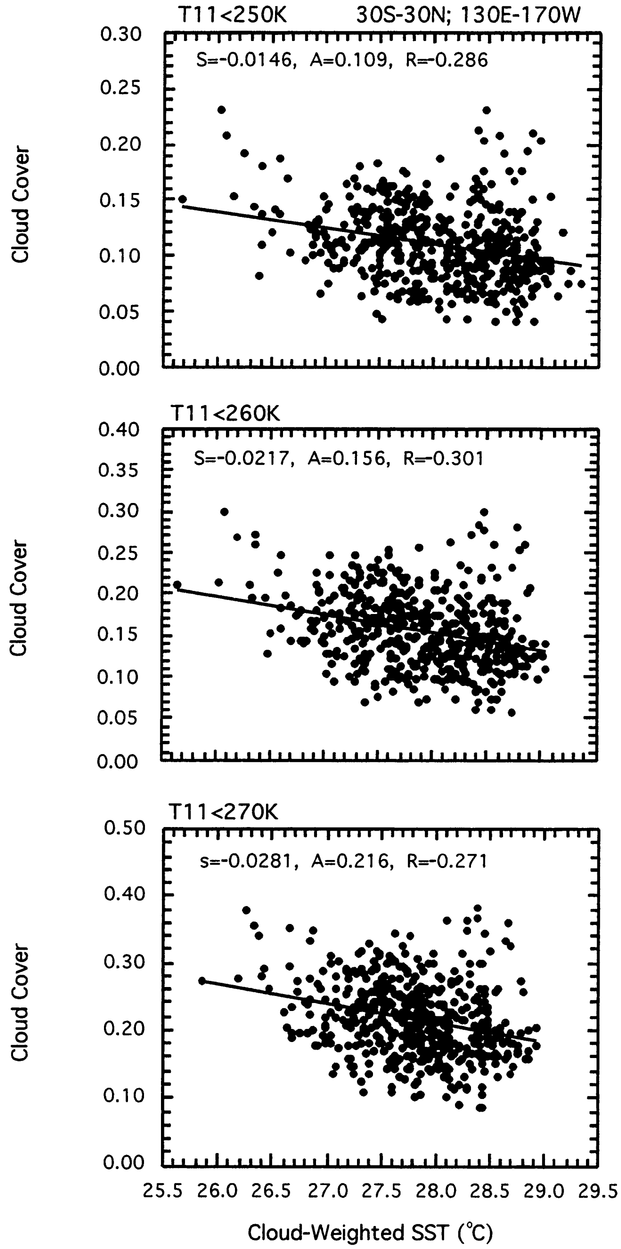

FIG. 1. The relation between cirrus cloud cover and the cloud-weighted SST for the period Jan 1998–Aug 1999. Each data point represents daily-mean and domain-mean values. Cirrus clouds are inferred using the brightness temperature measured in the GMS 11-μm channel, T_{11}. The three panels correspond to the use of three threshold temperatures for identifying cirrus clouds. The parameter S is the slope of the regression line, A is the domain-mean cloud cover, and R is the correlation coefficient.

old temperature if the approach of Lin et al. is followed. Thus, the climate feedback factor will change with the subjectively chosen threshold temperature, which could range broadly from 250 to 270 K.

There are two sources of thin cirrus clouds. One is the detrainment of deep convective anvil clouds, which spread, precipitate, and evaporate to become thin cirrus in the neighborhood of cumulus cloud clusters. The other is the thick cumulus clouds that are left behind propagating large-scale atmospheric disturbances and decay rapidly to become thin cirrus that contribute to the supply of water vapor in the upper troposphere. The upper-tropospheric water vapor may later form thin cirrus clouds because of atmospheric wave motions (Boehm and Verlinde 2000). These thin cirrus clouds are widespread and can persist for a long period of time because of large-scale lifting of air in the Tropics (Boehm et al. 1999). By analyzing the National Oceanic and Atmospheric Administration High Resolution Infrared Radiation Sounder radiance data, Wylie et al. (1994) estimated a global cirrus cloud cover of 40%. By taking into consideration the extended areal coverage and the low albedo of thin cirrus clouds, LCH specified the mean albedo of high-level clouds to be 0.24. With a fractional coverage of 0.25 and an albedo of 0.42 specified for the low boundary layer clouds, the albedo of the cloudy-moist region is then 0.35, which can be compared with the albedo of 0.51 as derived by Lin et al. using TRMM CERES and VIRS data. It appears that Lin et al. significantly underestimated the high-level cloud cover and overestimated the high-level cloud albedo and, hence, overestimated the sensitivity of shortwave radiation to high-level clouds.

We conclude that Lin et al. greatly underestimated the areal coverage and overestimated the albedo of the cloudy-moist region. The areal coverage of high-level clouds in the Tropics should be much greater than the value of 10% as estimated by Lin et al. The albedo of 0.51 of the cloudy-moist region as estimated by Lin et al. is representative of thick anvil clouds but is not the mean albedo of high-level cloudy regions. If we assume that their estimates of the OLR in the three tropical regions are appropriate for studying the climate sensitivity, the feedback factors of high-level clouds should remain negative as suggested by LCH, although the magnitudes are somewhat smaller, typically by 20%, which is still large when compared with the weak positive feedback as estimated by Lin et al.

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