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AQ1

•Constraints to the tropical low-cloud trends in historical climate simulations

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Abstract

Given an importance and difficulty in evaluating long-term trends in the tropical low-cloud amount (C_l), we examined mechanisms that determine the C_l trend in 20th century experiments using two different versions of the climate model called the Model for Interdisciplinary Research on Climate. The C_l trend patterns are coherent with trends in vertical velocity (ω) and lower-tropospheric stability (LTS). While the mean LTS trend varies and gives a stronger constraint to the C_l trends, the ω trend cannot do so due to mass conservation. Two of three reanalysis products support the positive LTS trend, but it is inconclusive because of the diversity in pattern and sign. Copyright © 2011 Royal Meteorological Society

AQ2

Keywords: low cloud; 20th century trend; •CGCM

1. Introduction

Simulations by global climate models (GCMs) have been improved during the past two decades in many aspects of climate. Nevertheless, there is a substantial divergence among GCMs in terms of the mean state, variability, and climate sensitivity under the doubling of CO_2. In particular, different sign or magnitude of the cloud-radiative feedback is still one of the largest sources of uncertainty in the climate change simulations (Soden and Held, 2006; Webb *et al*., 2006; Solomon *et al*., 2007). Bony and Dufresne (2005) demonstrated that the change in shortwave radiative forcing associated with low clouds (combination of stratiform, stratocumulus, and shallow cumulus clouds) has been the most different among GCMs. Namely, low clouds that have a net cooling effect in mean climate increase under the global warming and act as negative feedback in some GCMs, but vice versa in the others. A dominant role of low-cloud feedbacks in the models' climate sensitivity can even be seen in idealized GCMs (Medeiros *et al*., 2008).

It has been widely recognized that the tropical low-level cloud fraction (C_l) is partly controlled by the large-scale environment, especially over the subsidence regime. Klein and Hartmann (1993), and later Wood and Bretherton (2006), identified that the inversion strength above the atmospheric boundary layer (ABL) provides a good measure to the distribution and seasonal cycle of C_l. This thermodynamic constraint is typically measured by lower-tropospheric stability (LTS), defined by the difference in potential temperature (θ) between 700 and 1000 hPa levels. There is also a dynamic constraint that affects C_l: vertical pressure velocity (ω) at 500 hPa or the low-level divergence (Zhang *et al*., 2009).

While the global-mean surface air temperature (SAT) shows a continuous warming trend during the past decades, it is difficult to identify the trend in C_l and their feedback to climate because of the lack of cloud observations: shipboard measurements cannot resolve the vertical structure and is too sparse over oceans (Norris, 2009), whereas the satellite remote sensing provides only 20-year records at the longest (Rossow and Schiffer, 1999) and is dominated by artificial trends (Evan *et al*., 2007). Clement *et al*. (2009) discussed the low-cloud feedback associated with the Pacific decadal oscillation, but not trend, by combining satellite cloud products and GCM simulations. They concluded that the low clouds over the northeastern Pacific served as positive feedback, and further suggested a similar feedback at work over the entire Pacific under the global warming. Such an extrapolation may, however, be controversial as the metric constructed over a particular regime is used for arguing the cloud feedback in other regimes. In the present study, we address the issue of the low-cloud feedback based on two sets of historical climate simulations. A finding from the analysis is then applied to observations to infer a possible low-cloud change and the resultant radiative feedback during the past decades.

2. Twentieth century experiments by MIROC

We use two different versions of a climate model: the Model for Interdisciplinary Research on Climate

(MIROC), which has been developed at the Atmosphere and Ocean Research Institute [Former Center for Climate System Research (CCSR)], the University of Tokyo, National Institute for Environmental Studies (NIES), and Japan Agency for Marine-Earth Science and Technology. One is the MIROC version 3.2 (hereafter referred to as MIROC3), which has contributed to the Intergovermental Panel on Climate Change (IPCC) Fourth Assessment Report (AR4) (Solomon *et al.*, 2007). The other is the MIROC version 5.0 (MIROC5), which will be used for the IPCC Fifth Assessment Report (AR5). MIROC3 is a global atmosphere–ocean–land–sea ice model and includes an interactive aerosol module (K-1 model developers, 2004). The model resolution is T42L20 for the atmospheric component, and is $\sim 1°$ for the ocean component. MIROC5 is the latest version in which substantial changes are made to the parameterization schemes for most of atmospheric physical processes, ocean dynamics, and sea ice (Watanabe *et al.*, 2010). The atmospheric model resolution was doubled from that in MIROC3 for both horizontal and vertical dimensions, but the ocean model resolution is nearly unchanged. No flux correction is applied to both models.

Using MIROC3 and MIROC5, we conducted 20th century simulations following the experimental design proposed by the Climate Model Intercomparison Project. A ten-member (three-member) ensemble was made with MIROC3 (MIROC5) from January 1851 to December 2000. The 20th century simulation using MIROC3 is the same as that analyzed by Nozawa *et al.* (2005). Natural and anthropogenic forcing agents are almost identical between the two sets of experiments, except for some updates, such as the historical solar irradiance data (Lean *et al.*, 2005) and surface aerosols emission data, for the MIROC5 runs. We confirmed that these changes were not crucial for the results obtained in this study. In the next section, we focus on the 20th century (1901–1999) linear trend, denoted as Δ, based on annual- and ensemble-mean fields. Time series of the global-mean SAT are similar to each other, and well reproduce the observed changes during the 20th century. Yet, we should bear in mind that the equilibrium climate sensitivity is different by about 1°: 3.6 K in MIROC3 and 2.6 K in MIROC5 (Watanabe *et al.*, 2010). We have also conducted the 20th century runs using MIROC3 with higher horizontal resolution, known as MIROC3.2 (hires), and doubling CO_2 experiments using MIROC5 with lower resolution. They revealed that the results presented here are not seriously affected by changing the resolution.

3. Results

Linear trends in C_l and sea surface temperature (ΔC_l and ΔSST) are first compared between the two models [The definition of C_l followed the ISCCP (Rossow and Schiffer, 1999)] (Figure 1). In MIROC3, C_l decreases by 1–3% century^{-1} over the subtropical oceans and increases over the equatorial eastern Pacific and Atlantic (Figure 1(a)). The ΔC_l pattern is qualitatively different in MIROC5: increase by 0.5–2% century^{-1} over the subtropics except for limited areas around the eastern boundaries, and decrease over the equatorial Pacific (Figure 1(d)). These patterns show a considerable similarity to the respective ΔSST *relative* to the tropical mean (Figure 1(b) and (e)). Namely, C_l decreases where ΔSST is higher than the surrounding area. Overall, the trends are statistically significant at the 95% level because of ensemble averaging, but ΔC_l is less significant over some regions in MIROC5 due to larger natural variability.

It is noticeable that, despite the positive SST trend almost everywhere in the tropics, both positive and negative values are found in ΔC_l. The tropical-mean C_l is slightly decreasing by -0.28% century^{-1} in MIROC3 and increasing by 0.47% century^{-1} in MIROC5 (Figure 1(c) and (f)). These trends are significant at the 95% level and consistent with the positive and negative cloud shortwave radiation feedbacks identified in the $4 \times CO_2$ experiments in MIROC3 and MIROC5 (Watanabe *et al.*, 2010).

Spatial coherence between ΔC_l and ΔSST in the tropics, as seen in Figure 1, is quantified by their correlation coefficients, showing $r = -0.68$ (-0.54) for MIROC3 (MIROC5) (Table I). As has been argued in literature (Klein and Hartmann, 1993; Clement *et al.*, 2009), ΔC_l may also be explained by the trends in ω at 500 hPa ($\Delta\omega$) and in LTS (ΔLTS). The ΔC_l patterns are positively correlated with both $\Delta\omega$ and ΔLTS, indicating that either subsiding tendency or more stable stratification favors the production of C_l. In MIROC3 (MIROC5), ΔC_l is more coherent with ΔLTS ($\Delta\omega$), probably reflecting changes in C_l in different cloud regimes. Indeed, both observations and the MIROC control runs show that the C_l anomaly is highly correlated with the anomalous ω (LTS) primarily over the ascending (descending) regions (not shown). Both $\Delta\omega$ and ΔLTS are thus likely the major constraints to ΔC_l, but they are not independent of each other ($r = 0.4$ and 0.23; Table I) and are dependent on ΔSST.

The question how the tropical-mean C_l trend, denoted as $\langle \Delta C_l \rangle$, is determined is investigated by means of a probability density function (PDF) of $\Delta\omega$ and ΔLTS. Namely, we reconstruct ΔC_l as

$$\Delta\tilde{C}_l = \int_x P_x \Delta C_l(x) \mathrm{d}x \qquad (1)$$

where x denotes either $\Delta\omega$ or ΔLTS and $\Delta C_l(x)$ is the composite of ΔC_l with respect to x. In principle, we can choose any variable for x if it is highly correlated with ΔC_l.

Figure 2(a) shows the composite of ΔC_l referring $\Delta\omega$ to as x, and the PDF of $\Delta\omega$, i.e. $P_{\Delta\omega}$. As expected from Table I, ΔC_l tends to be positive where $\Delta\omega$ is positive, and vice versa. $P_{\Delta\omega}$ in the two models reveals a similar shape having the center slightly shifted to the

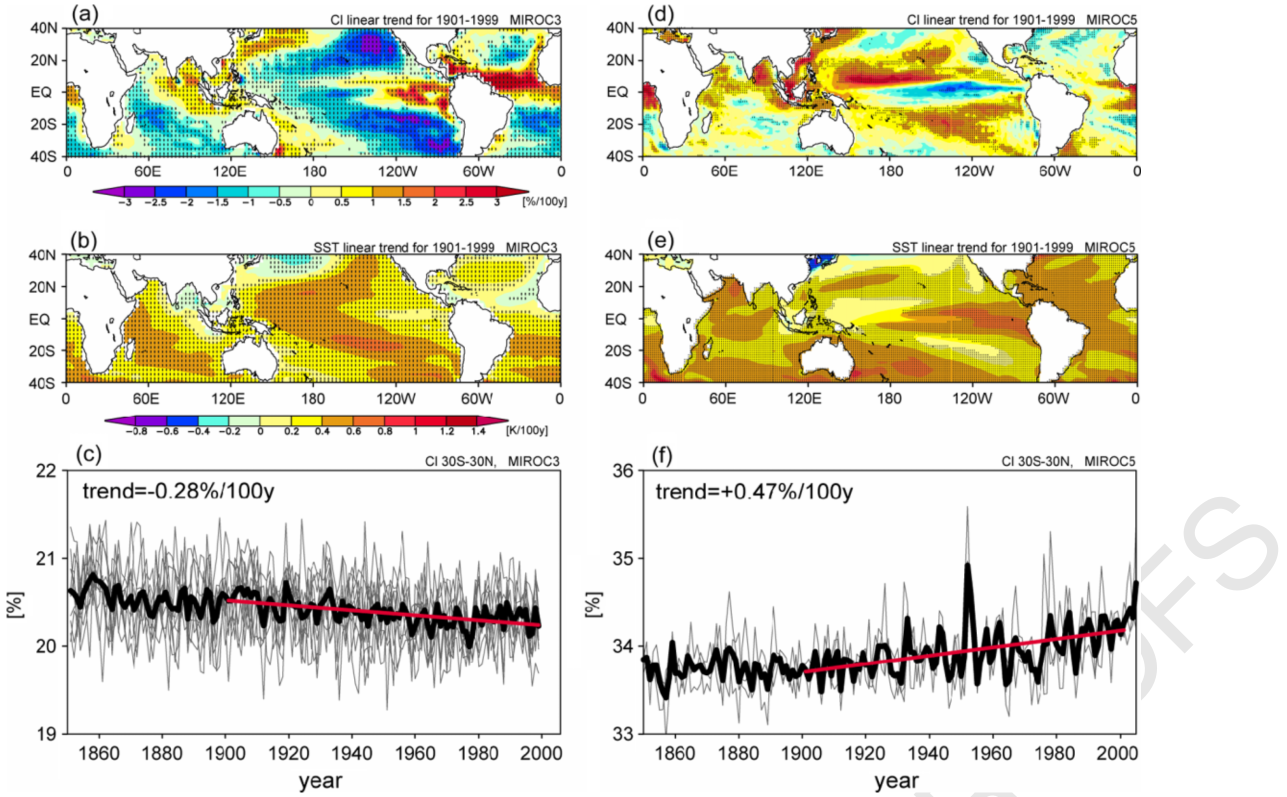

Figure 1. Linear trends in (a) C_l and (b) SST for 1901–1999 in MIROC3. The units are % K per century, respectively. Values significant at the 95% level are stippled. (c) Tropical-mean (30°S–30°N over oceans) C_l time series (thick curve; ensemble-average, thin curves; individual members). (d)–(f) As in (a)–(c) but for MIROC5.

Table I. Pattern correlation between the linear trends (Δ) of various quantities for 1901–1999. The values in bold are for MIROC3 and in italic for MIROC5. The correlation is calculated over the tropical ocean (30°S–30°N).

	ΔC_l	$\Delta\omega_{500}$	ΔLTS	ΔSST
ΔC_l	—	*0.62*	*0.40*	*−0.54*
$\Delta\omega_{500}$	**0.38**	—	*0.23*	*−0.50*
ΔLTS	**0.73**	**0.40**	—	***−0.75***
ΔSST	**−0.68**	**−0.52**	**−0.86**	

negative side but roughly close to the Gaussian. The change in ω within the Hadley cell should be constrained by the conservation of mass, which prohibits a uniform sign of $\Delta\omega$ even if ΔSST were uniform. The reconstructed $\langle\Delta\tilde{C}_l\rangle$ using Equation (1), −0.29 and 0.45% century^{-1}, close to values in Figure 1(c) and (f), is nearly unchanged when $P_{\Delta\omega}$ was replaced between MIROC3 and MIROC5.

The composite of ΔC_l with respect to ΔLTS also shows positive values for positive ΔLTS in both models (Figure 2(b)). However, $P_{\Delta\text{LTS}}$ is very different: PDF centered at around zero in MIROC3, whereas it shifted to the positive side in MIROC5. Indeed, reconstructed $\langle\Delta\tilde{C}_l\rangle$ using Figure 2(b) changes the sign when $P_{\Delta\text{LTS}}$ is exchanged between the models. This demonstrates that a thermodynamic effect of ΔLTS is the key for $\langle\Delta C_l\rangle$. The PDFs of ΔSST do not show a significant difference between MIROC3 and MIROC5; nevertheless, the slope and baseline of ΔLTS are different, suggesting distinct atmospheric responses to ΔSST (Figure 2(c)).

Given the leading role of ΔLTS in $\langle\Delta C_l\rangle$, we examine causes of ⟨ΔLTS⟩, which is decomposed to $\langle\Delta\theta\rangle$ at 700 and 1000 hPa ($\langle\Delta\theta_{700}\rangle$ and $\langle\Delta\theta_{1000}\rangle$). It is clear that the magnitude of $\langle\Delta\theta_{1000}\rangle$ is nearly identical in the two models, but $\langle\Delta\theta_{700}\rangle$ in MIROC5 is much larger than that in MIROC3 (Figure 3(a)). Reasons why the magnitude of $\langle\Delta\theta_{700}\rangle$ is different are further elaborated using the trends in tendency terms, $\langle\Delta\partial\theta_{700}/\partial t\rangle$ (Figure 3(b)). Each tendency has been directly obtained from the model, and is analyzed as in $\langle\Delta\theta_{700}\rangle$ by taking the annual- and ensemble-average. A common feature is found: cooling due to dynamics and cumulus convection, and warming due to radiative processes. The former two arise from the $\Delta\omega$ PDF having the center at a negative value (Figure 2(a)) and a convective heating profile being more top-heavy, whereas the latter warming trend is largely attributed to an increasing greenhouse effect. The $\langle\Delta\partial\theta_{700}/\partial t\rangle$ trends due to turbulence and cloud (excluding cumulus convection) have an opposite sign between the models, the former being minor. The most striking difference is therefore the heating/cooling trend due to non-convective clouds. Given a well-fitted $\langle\Delta\theta_{700}\rangle$, the total tendency has no trend by definition and therefore the terms having positive (negative) trend in MIROC5 (MIROC3) should explain the different magnitude of the warming in the lower troposphere.

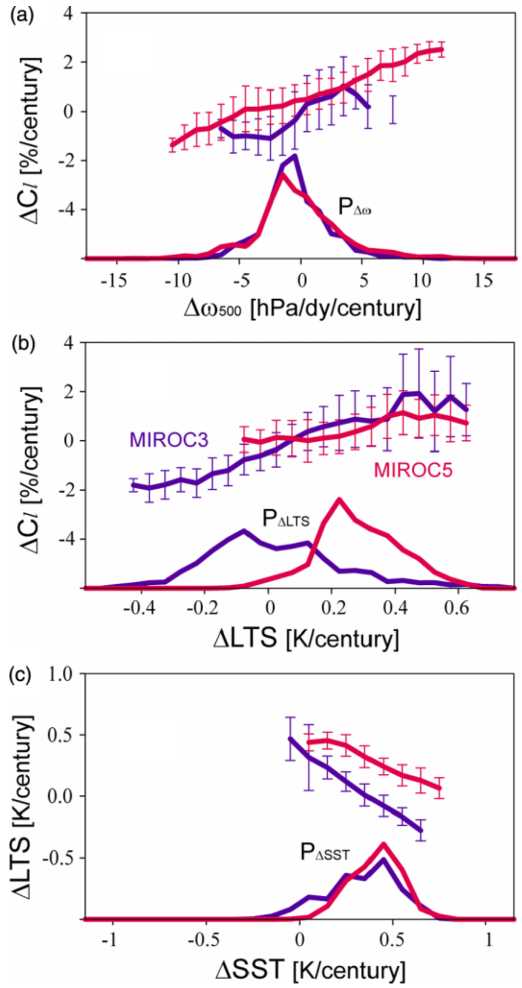

Figure 2. (a) Composites of tropical ΔC_l (% per century) with respect to $\Delta\omega$ (hPa per day per century) and its 1 SD (error bars) in MIROC3 (blue) and MIROC5 (red). The PDFs for $\Delta\omega$ are shown at the bottom of panel. (b) As in (a) but for the ΔC_l composites with respect to ΔLTS (K per century). (c) As in (a) but for the ΔLTS composites with respect to ΔSST (K per century).

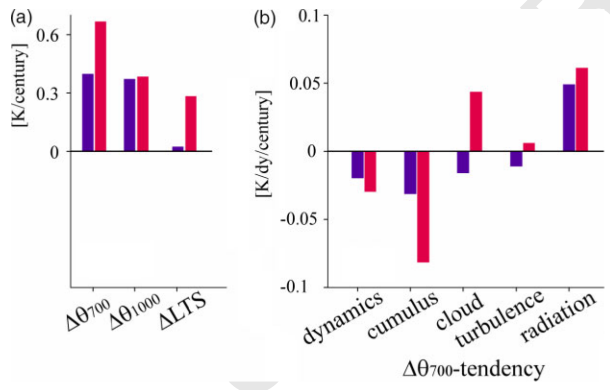

Figure 3. (a) Tropical-mean values of $\Delta\theta_{700}$, $\Delta\theta_{1000}$, and ΔLTS in MIROC3 (blue) and MIROC5 (red). (b) As in (a) but for θ_{700} tendency terms due to individual processes labeled at the bottom.

Tendency terms for individual cloud processes are available only in MIROC5 and the mechanism that leads to the positive trend is explained as follows. In the model's mean climate, evaporative cooling of cloud droplets dominates other terms in the cloud-induced tendency in the tropical middle troposphere between the ABL and melting layer. In a changing climate, less (more) warming accompanies a reduced (enhanced) evaporative cooling of cloud at the lower (upper) part of the layer below (above) 650 hPa, which generates the positive $\langle \Delta\partial\theta_{700}/\partial t \rangle$ trend in MIROC5.

As stated in the introduction, it is hard to verify the long-term C_l trend in the 20th century simulations because of the lack of C_l data. However, assuming that the ΔC_l–ΔLTS relationship holds in nature we can use temperature trend instead. Although reliability of the long-term climate variability derived from reanalysis data is controversial (Bengtsson *et al.*, 2004; Onogi *et al.*, 2007; Allen and Sherwood, 2008), lower-tropospheric temperature data excluding the pre-satellite era may be more reliable. Here, we compare ΔLTS in MIROCs with that calculated from the European Centre for Medium Range Weather Forecasts 40-year reanalysis (ERA40) (Uppala *et al.*, 2005) for 1979–2001, the National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) reanalysis (Kalnay *et al.*, 1996), and the Japanese 25-year reanalysis (JRA25) (Onogi *et al.*, 2007) for 1979–2009 (Figure 4). It is discouraging that the ΔLTS patterns are different from those in MIROCs and even among the reanalysis data. The magnitude of ΔLTS in the reanalysis is considerably large; $\langle \Delta C_l \rangle$ estimated from Figure 4(c)–(e) and an empirical relationship of $\Delta C_l = 5.7\Delta$LTS (Klein and Hartmann, 1993) gives −5.3, 3.2, and 4.1% century^{-1} for the ERA40, NCEP/NCAR, and JRA25 reanalysis, respectively. Difference in magnitude of the trends between models and reanalyses partly arises from natural low-frequency variability in short records of the reanalysis data. Indeed, the ΔLTS trends in MIROCs after 1979 show patterns similar to Figure 4(a) and (b) but with larger magnitudes (not shown).

It is likely that details of quality control and assimilation methods matter for generating discrepancy between the reanalyses. In addition, both SST and tropospheric temperature derived from microwave sounding unit (MSU), which are crucial for accurate estimation of ΔLTS in the reanalysis, suffer from the calibration problem (Christy *et al.*, 2003; Deser *et al.*, 2010). While two data sets show the positive $\langle \Delta$LTS$\rangle$, which is consistent with the temperature trends from corrected MSU data (Mears and Wentz, 2005) and therefore suggests positive $\langle \Delta C_l \rangle$ as in MIROC5, diversity of the pattern and sign shown in Figure 4(c)–(e) indicates that conclusive argument of the long-term LTS trend in nature is still far.

4. Summary and discussion

In the present study, we examined trends in C_l in two sets of the 20th century simulations using

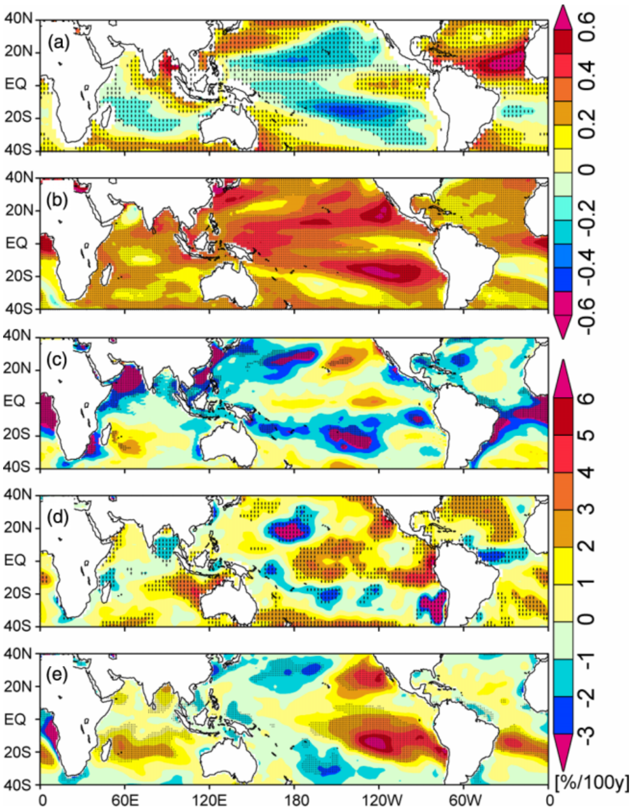

Figure 4. (a)–(b) As in Figure 1(a) and (d) but for LTS in MIROC3 and MIROC5, respectively. (c) The linear trend in LTS in ERA40 for 1979–2001. (d)–(e) As in (c) but for the NCEP/NCAR and JRA25 reanalysis for 1979–2009. Note different color scale between (a)–(b) and (c)–(e).

MIROC3 and its updated version of MIROC5. Consistent with the opposite sign of the cloud shortwave radiative feedback in doubling CO_2 experiments, we observed decreasing (increasing) C_l trend in MIROC3 (MIROC5). Out of two constrains to C_l, ω, and LTS, the thermodynamic effect due to ΔLTS is of primary importance in determining $\langle \Delta C_l \rangle$. The LTS trend is dominated by the trend in θ_{700}, which shows a different magnitude between the two models because of an opposite effect of cloud processes. The positive ΔLTS is also found in two out of three reanalysis data, suggesting a C_l increase during the past decades.

The result that the thermodynamic effect (i.e. change in LTS) is the primary controlling factor for the change in C_l supports the conclusions by Medeiros *et al*. (2008). We also examined the cloud response to uniform SST increase in aquaplanet experiments, and found the C_l changes consistent with ΔC_l in the 20th century runs. The ΔC_l in MIROC5, and that deduced from ΔLTS in reanalysis data, is suggestive of the negative cloud shortwave feedback, which has been obtained in low-order models (Miller, 1997; Larson *et al*., 1999). This may be an encouraging agreement, but some arguments with the simple models are not applicable to our GCM results. For example, they state $\Delta\theta_{700}$ being determined by SST changes in the regions of $\omega < 0$ based on a horizontal homogeneity of θ_{700} set by the convective adjustments. However, coexistence of the negative ΔC_l with higher ΔSST in the convective regions in MIROC3 does not match the argument. It may be useful to construct a simple model in terms of dynamics but involving cloud physics as complicated as the parameterization employed in GCMs.

The reason why ΔLTS in reanalysis data is so different from each other is not obvious. While ΔLTS derived from the NCEP/NCAR and JRA25 reanalysis supports the results in MIROC5, ΔLTS from the ERA40 reanalysis is not. The reanalysis-derived ΔLTS may include analysis errors and should be validated by comparing the same quantity estimated from well-calibrated satellite data such as MSU temperature. A combined analysis of the reanalysis, satellite, and *in situ* measurements may provide observational evidence of the past change in environmental condition for low clouds. Also, parts of the cloud response not simply explained by change in either ω or LTS are yet to be elaborated in detail. Specifically, GCM's ability in simulating boundary layer structure and cloud microphysical property should be severely tested by using satellite products.

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