Climate sensitivity of tropical and subtropical marine low cloud amount to ENSO and global warming due to doubled CO₂

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[1] In this study, we systematically analyzed the sensitivity of tropical and subtropical marine low cloud amount to the short-term climate anomaly associated with the 1997–1998 El Niño and the long-term climate change caused by doubled CO₂ using the International Satellite Cloud Climatology Project (ISCCP) cloud measurements, European Centre for Medium-Range Weather Forecasting (ECMWF) reanalyses, and the sea surface temperature (SST) forced and coupled simulations performed by the latest version of the National Center for Atmospheric Research (NCAR) and Geophysical Fluid Dynamics Laboratory (GFDL) climate models. It is found that the changes in low cloud amount associated with the 1997–1998 El Niño and the doubled CO₂ induced climate change have different characteristics and are controlled by different physical processes. Most reduction in low cloud amount related to the 1997–1998 El Niño occurs in the eastern tropical Pacific associated with an upward large-scale motion and a weak atmospheric stability measured by the 500 hPa vertical velocity and the potential temperature difference between 700 hPa and the surface, and is negatively correlated to the local SST anomaly. In addition to the other mechanisms suggested by the previous studies, our analyses based on the ISCCP observations indicate that the change in atmospheric convective activities in these regions is one of the reasons responsible for the change in low cloud amount. In contrast, most increase in low cloud amount due to doubled CO₂ simulated by the NCAR and GFDL models occurs in the subtropical subsidence regimes associated with a strong atmospheric stability, and is closely related to the spatial change pattern of SST consistent with previous studies. The increase in low cloud amount appears to favor the location where SST is less increased. After removing the background mean SST increase due to doubled CO₂, the results show a clear negative correlation between the change in low cloud and the SST change. An analysis based on the simple atmospheric mixed layer model demonstrates a thermodynamic reason for such a change. The increase in the above-inversion atmospheric stratification due to doubled CO₂ tends to reduce the mixed layer depth in the areas with a small temperature increase, which helps to trap the moisture within the mixed layer, thus, favors low cloud formation.


1. Introduction

[2] Tropical and subtropical low-level marine clouds consist of optically thick stratocumulus clouds, which usually form over the regions associated with relatively cold sea surface temperatures (SSTs) and the atmospheric subsidence, and optically thin shallow cumuli in the trade-wind regime. These low-level clouds play a pivotal role in the global climate system not only by affecting radiative budgets but also by promoting heat and moisture exchange between the sea-surface, the boundary layer, and the overlying troposphere. Thus they have a potential to exert strong feedbacks on and regulate climate change due to increasing greenhouse gases. Klein and Hartmann [1993] showed that a one-percent change in stratocumulus cloud cover could lead to a change in net cloud forcing at the top of the atmosphere (TOA) roughly 1 W m⁻², whereas the change in the direct forcing caused by doubled CO₂ is estimated to be 4 W m⁻², suggesting a moderate change in low-level clouds can markedly offset or amplify the global warming due to increasing greenhouse gases depending on the sign of cloud feedbacks [Slingo, 1990]. However, since in climate models the subtle balance between cloud impact on the solar and terrestrial radiation is affected by a complicated interplay between sub-grid scale processes (via parameterizations) and resolved processes, cloud introduces a feedback whose sign and amplitude are largely unknown. For example, in the low cloud dominated subtropical subsidence regime,
general circulation models (GCMs) do not agree on the sign of net cloud forcing feedback. Wyant et al. [2006] showed that a negative net cloud forcing feedback in the subsidence regime is simulated by the National Center for Atmospheric Research (NCAR) Community Atmosphere Model version 3 (CAM3), while the Geophysical Fluid Dynamics Laboratory (GFDL) Atmospheric Model 2 (AM2) generates a positive net cloud forcing feedback in the same region. Cloud forcing feedback is a complicated process that involves changes in both cloud coverage and cloud condensate. To isolate the problems, in this study we only focus on relatively simple but fundamental questions regarding low cloud amount: How does the amount of low clouds vary in response to climate change? and what physical processes are responsible for such a change? The answers to these questions currently are not well understood but are crucial to cloud-climate feedback issues.

[3] Since there is no observation available to verify the cloud feedbacks in future climate predicted by GCMs, a practical method to assess the climate sensitivity of low clouds is to document the cloud variability associated with the short-term climate anomalies observed in the current climate, such as El Niño-Southern Oscillation (ENSO). There have been a few studies focusing on low cloud variability associated with ENSO. Using ship-observed marine low-level clouds, Bajuk and Leovy [1998] and Park and Leovy [2004] examined the ENSO-related variations of different low cloud types over the tropical oceans. Klein et al. [1999b] showed the relationship between the SST anomaly in the central equatorial Pacific associated with El Niño and the low cloud variation in the remote ocean basins. Sun et al. [2003, 2006] and Zhang and Sun [2006] evaluated the cloud feedbacks through comparing GCM simulations with the International Satellite Cloud Climatology Project (ISCCP) observations. ENSO events are also widely used for evaluating model performance and testing cloud-climate interactions in GCMs [e.g., Cess et al. 2001a, 2001b; Lu et al., 2004]. The underlying philosophy is that if GCMs are to predict reliable long-term cloud-climate feedbacks, they must show their ability to reproduce the cloud variations associated with the climate anomaly occurring on shorter timescales.

[4] However, the potential use of ENSO events for assessing the change in low clouds due to increasing greenhouse gases is meaningful only if the following issues are clarified. First, what causes the change in low clouds during ENSO? Second, Can GCMs faithfully reproduce the observed change in low clouds? Third, what is the main difference of the change in low clouds in response to ENSO and doubled CO₂? and finally, what controls the low cloud variability in the doubling CO₂ experiment? The importance of this four-step investigation can be illustrated by Figure 1, which compares the observed change in marine low cloud amount associated with the 1997–1998 El Niño with those caused by doubled CO₂ simulated by GFDL AM2, NCAR CAM3, and a branch version of NCAR CAM3 that uses a new moist turbulent mixing scheme and a shallow convection scheme (details will be described in section 2). In this statistical analysis, the change in marine low cloud amount is binned based on the SST change relative to the mean SST change between 40°S and 40°N. This figure indicates that the change in low cloud amount in response to doubled CO₂ is not directly related to the local SST change but to the spatial change pattern of SST, a result consistent with Williams et al.’s [2003] findings.

[5] Furthermore, these statistics seem to show a similarity of the change in low cloud amount in response to El Niño and doubled CO₂, suggesting that El Niño might be considered as a surrogate for the climate change induced by doubled CO₂ in terms of the climate sensitivity of low cloud amount. However, a statistical result may gloss over the true physical mechanisms underneath a phenomenon. Thus this paper aims to provide a mechanistic analysis of the change in low cloud amount in response to El Niño and doubled CO₂. First, we will show that the change in low cloud amount in response to El Niño and doubled CO₂ not only occurs in different climate regimes but also is controlled by different physical processes. Second, we will provide a thermodynamic mechanism to explain Williams et al.’s [2003] findings that the change in low cloud amount caused by doubled CO₂ is related to the spatial change pattern of SST.

[6] This paper is organized as follows. Section 2 briefly describes the GCMs and observations that we used in this study. In section 3, we examine whether different GCMs can faithfully replicate the observed change in low cloud amount associated with ENSO if they are forced by the observed SST, and provide physical explanations why low cloud amount responds to ENSO in such a way. We will focus on the 1997–1998 El Niño. In section 4, we focus on analyzing the similarity and the difference of the change in low cloud amount in response to doubled CO₂ predicted by the NCAR and GFDL coupled GCMs. We further explore the mechanisms that control the change in low cloud amount due to doubled CO₂ using a simple atmospheric mixed layer model. In section 5, we present a further discussion on the results obtained in sections 3 and 4. Finally, a brief summary is provided in section 6.

2. Simulations and Observations

[7] Two types of GCM simulations are used in this study. The first was forced by the observed SST. This type of simulation was carried out from 1978 to 2001 by three GCMs: the standard NCAR CAM3, GFDL AM2, and a branch version of CAM3 (named as CAM3-UW hereafter). The detailed description of the relevant model physics and the simulated climatology by CAM3 and AM2 are referred to Collins et al. [2004] and Anderson et al. [2004], respectively. Here we only highlight the differences of model physics and vertical structure between the NCAR CAM3 and GFDL AM2 that are closely related to low clouds. In AM2, cloud amount is determined prognostically following the parameterization of Tiedtke [1993]. A K-profile scheme based on Lock et al. [2000] is used for treating the convective boundary layer. The mixing across the top of the convective layer is prescribed by an entrainment parameterization. Both deep and shallow convections are handled by the Relaxed Arakawa-Schubert scheme [Moorthi and Suarez, 1992]. AM2 has nine full levels in the lowest 1500 m. In CAM3, cloud amount is diagnosed mainly based on relative humidity, but the cloud scheme also considers the effect of atmospheric stability and convective mass fluxes. CAM3 uses a ‘dry’ bulk non-local K-profile
turbulence scheme that neglects the moist thermodynamics in calculating stratification of partially or totally saturated layers. CAM3 uses the Hack scheme [Hack, 1994] to represent shallow cumulus convection. CAM3 has five full levels in the lowest 1500 m. CAM3-UW uses the same dynamic core and physical parameterizations as CAM3 except that it employs a new ‘moist’ turbulence scheme, which includes an explicit cloud top entrainment parameterization [Grenier and Bretherton, 2001], and a new mass-flux based buoyancy-sorting shallow convection scheme [Bretherton et al., 2004] in which cloud base mass flux and convective triggering are controlled by the mixed layer turbulent kinetic energy (TKE) and a convective inhibition. CAM3-UW has eight full levels below 1500 m.

Figure 1. Binned change in low cloud amount based on the SST change after removing the mean SST change over the ocean basin between 40°S and 40°N associated with the 1997–1998 El Niño (observations) and doubled CO$_2$ simulated by the NCAR and GFDL GCMs. Solid lines scaled to the right indicate the PDF of each temperature bin. CAM3-SOM and AM2-SOM represent the standard NCAR and GFDL atmospheric models coupled with a slab ocean model. CAM3-UW-SOM is a branch version of CAM3-SOM that uses a new moist turbulent mixing scheme and shallow convection scheme (see details in section 2).

[8] In the SST forced simulations, the ISCCP simulator, a diagnostic method developed by Klein and Jakob [1999a] and Webb et al. [2001], was activated. Since the ISCCP simulator emulates the ISCCP algorithm to mimic satellite observations, the diagnosed clouds in simulations can be directly compared with the ISCCP observations. Thus it provides a convenient means to evaluate models’ performance.

[9] The run-to-steady state doubling CO$_2$ experiments were performed by CAM3, CAM3-UW, and AM2 coupled with a slab ocean model. These experiments are named as CAM3-SOM, CAM3-UW-SOM, and AM2-SOM in the paper, respectively. The simulations were executed for 50 years and the last 20-year data were used for analyzing
the climate sensitivity of low clouds to the doubled CO$_2$
induced climate changes.

[10] The main observational data used in this paper are the ISCCP D2 products, which consist of monthly mean global cloud amount of different cloud types classified based on their cloud top pressure $P$ and optical thickness $\tau$ [Rossow and Schiffer, 1991; Rossow et al., 1996]. Total nine types of clouds are classified by ISCCP D2. These include, high clouds ($P < 440$ hPa): “cirrus” ($\tau < 3.6$), “cirrostratus” ($3.6 < \tau < 22.6$), and “deep convective clouds” ($\tau > 22.6$); mid clouds ($440$ hPa $< P < 680$ hPa): “altocumulus” ($\tau < 3.6$), “altostatus” ($3.6 < \tau < 22.6$), and “nimbostratus” ($\tau > 22.6$); low clouds ($P > 680$ hPa): “cumulus” ($\tau < 3.6$), “stratocumulus” ($3.6 < \tau < 22.6$), and “stratus” ($\tau > 22.6$). As demonstrated by Hahn et al. [2001] and other studies, the ISCCP defined cloud types may not necessarily correspond to those obtained from surface observations. Thus although ISCCP D2 clouds are named following the convention of cloud morphology, they are not necessarily the same clouds as their names suggested, rather, they should be interpreted based on cloud top pressure and optical thickness. For example, ISCCP “cumulus” and “stratus” are not the real cumulus and stratus. They only represent the low clouds with a thin and thick optical depth. To prevent confusion, all the quoted cloud names hereafter refer to the ISCCP D2 cloud types. Since the ISCCP D2 data set is available from 1983 to 2001, we can examine in detail how different types of low clouds vary in response to the major ENSO events such as the 1997–1998 El Niño.

[11] However, due to the ‘top-down’ satellite view, the ISCCP observations may under-estimate the actual low cloud amount due to the possible obscuring from mid and high clouds. Since the change in low cloud amount associated with ENSO is accompanied by the change in mid and high clouds in the same region, it is important to examine this obscuring effect before using this data set to investigate the change in low cloud amount associated with ENSO. Note that ISCCP does not measure cloud amount directly, rather, the VIS/IR/NIR channels of ISCCP measure pixel radiances. Through the comparison with clear sky radiances, the pixel cloud optical thickness and cloud height are determined. Thus if ISCCP detects, for example, high “cirrus” ($\tau < 3.6$) only, there must be no thick low “stratocumulus” or “stratus” ($\tau > 3.6$) underneath, otherwise, the ISCCP retrieved cloud optical thickness will be larger than 3.6 and the detected clouds will be no longer “cirrus” any more. In other words, if there exists high clouds, the “stratus” may have a smaller optical thickness threshold (Dr. William B. Rossow, personal communication). Therefore the ISCCP low cloud observations are reliable under circumstances in which the obscuring problem is not severe.

[12] In this study, we have carefully examined the ISCCP cloud climatology in the tropics and subtropics and its perturbation associated with the 1997–1998 El Niño event (Figure 2). As shown in Figure 2a1, the maximum reduction in low cloud amount (taken as the sum of “cumulus”, “stratocumulus”, and “stratus”) occurs in the eastern tropical Pacific roughly bounded by (10°S–6°N and 130°W–82°W). In this region, the annual mean high and mid cloud amounts are less than 10% in normal years (Figures 2a1, 2b1, 2c1, and 2d1). During the 1997–1998 El Niño, high clouds increase substantially in this region, but most of them are the optically thin high “cirrus” clouds ($\tau < 3.6$, Figure 2a3), while most decrease in low cloud amount comes from thick low “stratocumulus” and “stratus” ($\tau > 3.6$, Figure 2e2). Thus the substantial increase in high “cirrus” will not affect the reliability of the detected change in thick low clouds since it is impossible for “cirrus” to obscure “stratocumulus” and “stratus” for the reason that we stated previously. The same also holds for thin mid “altocumulus”. Therefore the increase in thin mid and high clouds will not critically affect our analysis on low clouds since it is the thick low clouds that change most during the 1997–1998 El Niño. The only chance for obscuring thick low clouds is from thick mid and high clouds. To account for this possible obscuring, Figure 2e3 plots the total change in thick clouds (the sum of “stratocumulus”, “stratus”, “nimbostratus”, “cirrostratus”, and “deep convective clouds”). It clearly shows that the decrease in thick low clouds in the eastern tropical Pacific cannot be solely explained by the obscuring from the increase in thick mid and high clouds in the same region. Accounting for the possibility that the increased thick mid and high clouds are always accompanied by the underlying low clouds, there is still more than 20% decrease in thick low cloud amount in the eastern tropical Pacific. Keep in mind that this obscuring is just a hypothesis, low clouds may not necessarily lie right underneath mid and high clouds, which means that the real decrease in low cloud amount in the eastern tropical Pacific should be larger than 20%. The change in ISCCP low clouds during the 1997–1998 El Niño is consistent with the surface-based observations of other ENSO events. Park and Leovy [2004] showed that low clouds decrease in the eastern equatorial Pacific during positive ENSO phase. However, we note that the ISCCP observed change in low cloud amount associated with ENSO in the western and central tropical Pacific may be significantly affected by mid and high clouds since low clouds in this region are basically the thin shallow “cumuli” [Jin and Rossow, 1997] or “nimbostratus”.

[13] In addition to the ISCCP measurements, the European Centre for Medium-Range Weather Forecasting (ECMWF) reanalyses are used for the analysis of the atmospheric thermodynamic and dynamic state.

3. Low-Cloud Amount Change in Response to the 1997–1998 El Niño

[14] To examine if the NCAR and GFDL models can faithfully reproduce the observed change in low cloud amount associated with ENSO, Figure 3 directly compares the change in the sum of ISCCP “cumulus”, “stratocumulus”, and “stratus” associated with the 1997–1998 El Niño with those produced by the ISCCP simulators activated during the simulations. For comparison, the changes in the vertically projected low cloud amount (below 680 hPa) directly simulated by the three models are also provided. This vertically projected cloud amount considers the vertical cloud overlapping effect. In the NCAR models, the treatment of cloud vertical overlap follows Collins [2001], while AM2 uses a
random overlap assumption. Compared with the ISCCP measurements (Figure 3a1), CAM3 and CAM3-UW (Figures 3a2 and 3a3) reasonably reproduce the ISCCP observed change in low cloud amount, particularly in the eastern tropical Pacific. Both CAM3 and CAM3-UW fail to reproduce the increase in low cloud amount over the western warm pool. AM2, on the other hand, significantly under-estimates the decrease in low cloud amount in the eastern tropical Pacific and the increase in the western warm pool.

As we discussed previously, even fully taking into account of the obscuring effect from mid and high clouds, the thick low clouds still decrease more than 20% in the eastern tropical Pacific. This reduction in ISCCP type low clouds is also where SST increases most during the El Niño (Figure 3b), it suggests that there is a negative correlation between the SST anomaly and the change in low cloud amount.

To clearly show this relationship, we computed the correlations between the Nino 3.4 ($5^\circ$S–$5^\circ$N, $120^\circ$W–$170^\circ$W) SST anomaly and the change in low cloud amount, and the correlations between the local SST anomaly and the change in low cloud amount using the data from 1983 to 2001 (Figure 4). The Nino 3.4 SST anomaly is chosen since it usually corresponds to the strongest change in SST and is often used as an index for defining ENSO events. A correlation of the change in a variable to the Nino 3.4 SST anomaly can highlight the relationship between this variable and ENSO. Thus the correlation between the change in low cloud amount and the Nino 3.4 SST anomaly...
can tell us how much the change in low cloud amount elsewhere all over the world is related to the ENSO perturbations. For a better comparison with the ISCCP observations, both the ISCCP simulator generated low clouds (the sum of “cumulus”, “stratocumulus”, and “stratus”) and the vertically projected low clouds were used for the calculation. The spatial pattern of correlations between the Nino 3.4 SST anomaly and the change in ISCCP low cloud amount is well reproduced by the NCAR and GFDL GCMs. This is certainly an encouraging result considering all three models using different turbulence and cloud parameterizations and different vertical resolutions. Interestingly, all three models also have similar biases to observations, i.e., an over-estimated negative correlation in the central Pacific and an under-estimated positive correlation in the central Indian Ocean. CAM3 further produces a spurious double negative correlation structure striding over the Equator. This bias is less severe in the CAM3-UW simulations although it still exists. Note that this correlation may not well represent the real correlation between the change in low cloud amount and the Nino 3.4 SST anomaly since the s are the ISCCP type clouds that may be contaminated by cloud obscuring, especially in the central Pacific and the western warm pool. Nevertheless, the comparison provides a means to evaluate the model’s performance.

Figure 3. Left column: Changes in the sum of ISCCP “cumulus”, “stratocumulus”, and “stratus” associated with the 1997–1998 El Niño compared with those generated by the ISCCP simulators activated in the simulations. (b): SST anomaly associated with the 1997–1998 El Niño. (c2), (c3), and (c4): Changes in the corresponding vertically projected low clouds (considering vertical cloud overlapping) simulated by CAM3, CAM3-UW, and AM2. The white solid lines indicate the zero lines.

However, a substantial difference between the NCAR and GFDL models is shown in the change in the vertically projected low clouds. While a similar correlation pattern remains in the NCAR model simulations, the correlation between the SST anomaly and the change in low cloud amount reduces significantly in the AM2 simulations. Since the change in the projected clouds presumably represents the true change in low cloud amount if the vertical cloud overlapping parameterization is appropriate, then, the substantial difference between the ISCCP-view low clouds and the vertically projected low clouds in the AM2 simulation may be attributed to the strong cloud obscuration. However, if the vertical cloud overlapping parameterization is not realistic, then, it is difficult to separate the error associated with the vertical cloud overlapping parameterization and effects from cloud obscuration since both exist simultaneously in the simulations. The same holds for the NCAR model simulations. Currently, we do not have an effective
Figure 4. (a) Correlations between the Nino 3.4 SST anomaly and the change in low cloud amount based on the data from 1983 to 2001. Left column shows the results computed using the ISCCP observed low clouds (sum of “cumulus”, “stratocumulus”, and “stratus”) and those generated by model ISCCP simulators. The correlations computed using the vertically projected low clouds are shown in the right column. (b) The same as (a) but for the correlations between the local SST anomaly and low cloud amount change. The white solid lines indicate the zero lines.
A striking feature is the negative correlation between the local SST anomaly and the change in ISCCP type low cloud amount (Figure 4b) over almost the entire low-latitude ocean basin with the strongest correlation along the Equator in the central and eastern Pacific. There is a good agreement between observations and simulations although the latter over-estimates this negative correlation. Despite the possible cloud obscuring effect, this negative correlation appears to be consistent with the previous observational studies [Norris and Leovy, 1994; Park and Leovy, 2004]. Again, the relationship derived from the ISCCP type clouds remains robust in the NCAR model simulations when low clouds are taken as the vertically projected clouds below 680 hPa, but this negative correlation is substantially reduced in the AM2 simulation, in particular, a weak positive correlation is shown in the southwest Pacific. Despite that, AM2 also agrees with a negative correlation between the local SST anomaly and the change in low cloud amount in the eastern tropical Pacific.

Understanding the physical mechanisms underlying the negative correlation between the change in low cloud amount and SST anomaly may shed light on diagnosing climate sensitivity of low clouds. Observationally, in the stratocumulus regime, the cloud amount of stratocumulus is correlated to the low tropospheric stability (LTS) documented by Klein and Hartmann [1993]. The change from the stratocumulus regime to other low cloud regimes over oceans, however, appears to be related to the large-scale tropospheric vertical velocity. Thus a detailed examination of the relation between the change in low clouds and SST anomaly in different dynamic and thermodynamic climate regimes is useful. To do so, we bin the change in marine low cloud amount and the SST anomaly from 40°S to 40°N based on the 500 hPa vertical velocity and stability measured by the potential temperature difference between 700 hPa and the surface from the ECMWF reanalyses (Figure 5).

Figure 5. Upper panels: the change in marine ISCCP low cloud amount and the SST anomaly associated with the 1997–1998 El Niño between 40°S and 40°N binned based on the annual mean (averaged over the 1989–2001 period) 500 hPa vertical velocity and stability defined by the potential temperature difference between 700 hPa and the surface from the ECMWF reanalyses. Lower panels: the same as the upper panels but binned based on the 500 hPa vertical velocity and stability averaged over the 1997–1998 El Niño period from June 1997 to May 1998. The white dashed lines indicate the zero lines.
This stability and 500 hPa $\omega$ are chosen consistently with Klein and Hartmann [1993] and Bony et al. [2004]. Since the large-scale dynamic and thermodynamic conditions are different between a normal year and an ENSO year, the 500 hPa $\omega$ and the stability are different between normal years (taken as the averaged over the 1983–2001 period) and the 1997–1998 El Niño year (June 97 to May 98). To distinguish this difference, the upper panels of Figure 5 show the change in low cloud amount and the SST anomaly binned based on the normal year 500 hPa $\omega$ and stability. It shows that the reduction in low cloud amount and the SST increase associated with the 1997–1998 El Niño occur in the climate regime with a weak subsidence and a relatively strong atmospheric stability (indicated by the boxes). However, if the change in low cloud amount and the SST anomaly are binned based on the 500 hPa $\omega$ and the stability from the 1997–1998 El Niño year (lower panels of Figure 5), the reduction in low cloud amount and the SST increase occur in the ascending branch of the circulation associated with the weak atmospheric stability (indicated by the boxes), suggesting that the atmospheric circulation change associated with the 1997–1998 El Niño plays an important role in the change in low cloud amount. In addition to this major change in low cloud amount associated with the circulation change, there are weak change in low cloud amount in the other climate regimes. These changes may be associated with local boundary layer processes.

[20] The influence of the general circulation change on low clouds during the 1997–1998 El Niño is clearly shown in Figure 6, which compares the 850 hPa vertical velocity $\omega$ from the ECMWF reanalyses with those simulated by CAM3, CAM3-UW, and AM2. The narrow ascending branch of the circulation in normal years is widened and shifted southward during the 1997–1998 El Niño, which is an evidence of the enhanced atmospheric convection in the eastern tropical Pacific associated with the SST increase in this region. The enhanced convection over the areas with a positive SST anomaly is also supported by the weakening of the atmospheric stability in this region shown in Figure 7. The circulation change and the stability reduction shown in the ECMWF reanalyses are well agreed by all three models. Deser et al. [1993] argued that the reduction in low cloud amount in the warm El Niño year (1987) is associated with the weak cold advection across the equatorial front. Here based on the results shown in Figures 5, 6, and 7, we propose another mechanism for the reduction in low cloud

![Figure 6. Comparison of 850 hPa vertical velocity among ECMWF, CAM3, CAM3-UW, and AM2. Left column: the 850 hPa annual mean vertical velocity $\omega$; Middle column: the 850 hPa $\omega$ averaged from June 1997 to May 1998; Right column: the 1997–1998 averaged $\omega$ minus the annual mean $\omega$. White dotted lines in the figures denote the zero lines.](image-url)
The enhanced convection breaks up the low stratocumulus, reducing low cloud amount relative to a normal year with cooler SSTs. This mechanism appears to be supported by the ISCCP observations that most decrease in low cloud amount comes from thick ISCCP “stratocumulus” and “stratus” shown in Figure 2e2. The mechanism that convection evaporates stratiform clouds in the atmospheric boundary layer has been widely recognized by observations and simulations [Albrecht et al., 1995; McCaa and Bretherton, 2004; de Szoeke et al., 2006].

The enhanced convective activities in the eastern tropical Pacific during the 1997–1998 El Niño are further confirmed by the change in mid and high clouds shown in Figure 8. The increase in mid and high clouds associated with the enhanced convection in the eastern tropical Pacific during the El Niño is clearly shown in the ISCCP observations. CAM3 and CAM3-UW predict a similar change pattern of high clouds (“cirrus”, “cirrostratus”, and “deep convective clouds”) to the ISCCP observations, but both significantly under-estimate the change in mid clouds (“altostratus”, “altostratus”, and “nimbostratus”). AM2, on the other hand, substantially under-estimates the change in both mid and high clouds over the tropical and subtropical ocean basin.

In contrast to the enhanced convective activity in the eastern tropical Pacific, in the western tropical warm pool the decrease in upward motion (Figure 6) and the reduction in higher level clouds (Figure 8) suggest a weakened convection there during the El Niño, which might favor the formation of low clouds. As shown in Figure 3a1, the ISCCP low cloud amount increases over the Indonesia archipelago tropical Indian Ocean during the El Niño. However, as we pointed out, since the ISCCP low clouds in this region may be substantially obscured by mid and high clouds, the suggested mechanism needs to be further examined using in situ observations. For example, the surface in situ observations from the Atmospheric Radiation Measurements at the Tropical Western Pacific site can be used for such an analysis. We will explore this issue in our future research.
To further diagnose the model biases, we examined the separate climatological distribution of “stratus” and “stratocumulus” and their change associated with the 1997–1998 El Niño (Figures 9 and 10). Compared with the ISCCP observations, CAM3 significantly over-estimates the climatological value of “stratus” amount compared with the ISCCP observations. This bias is similar to that of the previous version of CAM. A slight improvement is shown in the CAM3-UW simulation but “stratus” amount is still well over-estimated. The substantially over-estimated “stratus” in CAM3 and CAM3-UW may explain the corresponding large change in “stratus” in response to the 1997–1998 El Niño. AM2, on the other hand, gives a better prediction in climatological “stratus” amount and the change associated with the 1997–1998 El Niño. In contrast, CAM models, especially CAM3-UW, predict a climatological “stratocumulus” amount and the response to the 1997–1998 El Niño very close to the ISCCP observations, while AM2 significantly under-estimates the “stratocumulus” amount and the change related to the El Niño. Another important information shown in the ISCCP observation (Figures 9a1 and 10a1) is that the “stratus” amount is very small compared with the “stratocumulus” amount. This result is consistent with the surface observations that the frequency of true stratus clouds over subtropical oceans is negligible compared with stratocumulus [Norris, 1998]. A further study is needed to explore why CAM models tend to over-predict optically thick “stratus” while AM2 tends to under-predict optically thin “stratocumulus”. The slight improvement shown in the CAM3-UW simulations suggests that the turbulent mixing scheme and its interaction with shallow convection scheme may be a key to obtain a realistic simulation of stratiform clouds.

4. Low-Cloud Amount Change in Response to Doubled CO2

In this section, we focus on the change in low cloud amount in response to doubled CO2 simulated by CAM3-SOM, CAM3-UW-SOM, and AM2-SOM. We attempt to answer the following questions: What are the major differences of the change in low cloud amount in response to...
ENSO and doubled CO$_2$? What are the most likely factors that determine the change in low cloud amount in the doubling CO$_2$ experiment?, and what are the physical mechanisms behind them? To answer these questions, the analyses of the CAM3-SOM, CAM3-UW-SOM, and AM2-SOM simulations are further combined with a simple atmospheric mixed layer model that works well in the current climate.

The simulations of earlier version of the NCAR and GFDL GCMs (CAM2 and AM2p10) show that the change in subtropical marine low cloud amount has a different sign in the doubling CO$_2$ experiment. To find the cause for this inconsistency is one of the motivations for this study. Figures 11a2, 11b2, and 11c2 compare the change in low cloud amount in response to doubled CO$_2$ in the CAM3-SOM, CAM3-UW-SOM, and AM2-SOM simulations. Since the ISCCP simulator was not activated in the doubling CO$_2$ experiments, the vertically projected low clouds are used for all double analysis hereafter. The latest version of the NCAR and GFDL GCMs now at least produce the same sign of the change in low cloud amount in the subtropical subsidence regimes. The change in subtropical low cloud amount predicted by the GFDL GCM has changed its sign from negative (earlier version) to positive (latest version). Such a change may be attributed to the fact that the latest version AM2(p12) uses a high vertical resolution of nine levels below 1500 m instead of five levels in AM2(p10) and the Lock et al. [2000] turbulence scheme for the boundary layer including an explicit cloud top entrainment parameterization instead of Mellor and Yamada [1982] scheme. One of the major differences between CAM3-SOM and CAM3-UW-SOM simulations is the change in subtropical low cloud amount in the northeast Pacific. The increase in low cloud amount in this region shown in the CAM3-SOM simulation has been reduced in the CAM3-UW-SOM simulation, and is closer to what is shown in the AM2-SOM simulation. A possible explanation is that the Lock turbulent mixing scheme used in AM2 is

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**Figure 9.** Left column: the annual mean of “stratus” amount from the ISCCP observations and the CAM3, CAM3-UW, and AM2 simulations. Right column: the changes in “stratus” (r > 22.6) associated with the 1997–1998 El Niño from the ISCCP observations and the CAM3, CAM3-UW, and AM2 simulations. The simulated clouds are from the ISCCP simulators activated in the simulations. White dotted lines (b1, b2, b3, b4) denote the zero lines.
somewhat similar to the turbulence scheme used in CAM3-UW in the sense that both schemes have an explicit cloud top entrainment parameterization. Despite the similarities here and there, the three models show marked differences in the change in low cloud amount in response to doubled $\text{CO}_2$.

[26] To seek what controls the change in low cloud amount in the doubling $\text{CO}_2$ experiment, we plot the corresponding change in SST in Figures 11a1, 11b1, and 11c1. One noticeable difference between the NCAR and GFDL model simulations is that AM2-SOM produces a strong asymmetric SST change between the north and south Pacific. In contrast, a roughly symmetric SST change in the Pacific with respect to the equator is simulated by CAM3-SOM. It is not totally clear why the two models produce such a different SST change. Since the symmetric SST change in the Pacific is broken down in the CAM3-UW-SOM simulation, at least part of the reason may be attributed to the different turbulent mixing scheme and shallow convection scheme used by the two models. To fully understand why NCAR and GFDL model produce such a different SST change pattern, further study is needed.

[27] The most striking feature is that the spatial change pattern of SST is remarkably similar to the spatial change pattern of low clouds but with an opposite phase. The increase in low cloud amount appears to favor the location where SST is less increased and vice versa. This relationship is even more clear if we classify the change in low cloud amount into different climate regimes. Like what we did in the ENSO analysis, we bin the changes in marine low cloud amount, SST, and atmospheric stability due to doubled $\text{CO}_2$ from 40°S to 40°N into different categories based on the 500 hPa vertical velocity $\omega$ and the atmospheric stability ($\theta_{700\text{hPa}} - \theta_{\text{surf}}$) taken from both the current and doubled $\text{CO}_2$ climates, where the SST change is the one after removing the mean SST increase between 40°S and 40°N, which is about 1.81°C, 1.72°C, and 2.14°C in the CAM3-SOM, CAM3-UW-SOM, and AM2-SOM simulations, respectively. Unlike the change in low cloud amount associated with 1997–1998 El Niño (Figure 5), which shifts from a subsidence regime (when it is binned based on the normal year climate) to a weak convective regime (when it is binned based on the El Niño year climate), there is no substantial change in climate regimes associated with the change in low cloud amount due to doubled $\text{CO}_2$. Thus Figure 12 only shows the binned change in low cloud amount, SST, and stability based on the doubled $\text{CO}_2$ climate. The top panels of Figure 12 clearly indicate that most increase in low cloud amount due to doubled $\text{CO}_2$ occurs in the subsidence regimes associated with a strong atmospheric stability, while the low cloud decrease can happen in both ascending and descending branch of the
circulation, but is associated with a weak atmospheric stability. An apparent difference is that the stability corresponding to the maximum increase in low cloud amount is different in the three models, about 25 K in CAM3-SOM and 19 K in AM2-SOM with the CAM3-UW-SOM result lying somewhere in-between.

Comparing the changes in low cloud amount, SST, and stability in the same dynamic and thermodynamic regime, Figure 12 also clearly shows the relationship between these changes. For example, in CAM3 a large increase in atmospheric stability corresponds to an increase in low cloud amount, while a small increase in stability is associated with a decrease in low cloud amount. This is also true for CAM3-UW and AM2, although the change in stability is different between the models. However, the change in stability does not appear to be a good index for distinguishing the sign of the change in low cloud amount in the warmer climate since there is no easy way to determine a critical value of stability increase that separates the increase or decrease in low cloud amount. It appears that the change in low cloud is more cleanly correlated to the SST change after removing the background mean SST increase. As indicated by Figure 12 (upper and middle panels), a positive or negative change in SST approximately corresponds to a decrease or increase in low cloud amount, respectively. This is true for all three models, although the details are different. This result is consistent with Williams et al.’s [2003] finding that the spatial change pattern of SST is correlated to the change in low clouds. This relationship is also clearly shown in Figure 1 in which the change in low cloud amount is binned based on the SST change between 40°S and 40°N after removing the background mean SST increase. An immediate question that one would ask is: why the increase in low cloud amount prefers the location where SST is less increased? Understanding the physics behind this phenomenon is helpful to clarify some issues regarding different cloud-climate feedbacks shown in GCM simulations.

To answer this question, it is useful to simplify the problem and highlight the physical processes that are
involved as indicated by the previous studies [e.g., Miller, 1997]. As shown in Figures 11 and 12, most increase in low cloud amount due to doubled CO$_2$ occurs in the subsidence regime. In this regime, the evolution of boundary layer clouds is controlled by four processes. First, the mixed layer warming caused by both heat input from the surface and entrainment of warmer free atmosphere into the mixed layer; Second, the mixed layer growth controlled by entrainment and large-scale subsidence; Third, the mixed layer moistening or drying depends on the net effect of the surface evaporation and entrainment of dry free atmosphere into the mixed layer; Fourth, the cloud top long-wave cooling effect. These processes can be effectively related in a simple atmospheric mixed layer model framework pioneered by Lilly [1968], which has been extensively used to study the cloud topped boundary layer in the subsidence regime [e.g., Betts, 1989; Garratt, 1992]. Under the mixed layer model framework, the balance within the mixed layer and at the layer top can be expressed as [Lilly, 1968, equations 4 and 5],

$$\frac{\partial \theta_v}{\partial t} = -\frac{1}{h} \left( \frac{w' \theta_{v0}^0}{C} - \frac{w' \theta_{vh}}{C16/C17} \right),$$

$$\left( \frac{\partial h}{\partial t} + W_h \right) \Delta \theta_v + \frac{w' \theta_{vh}}{C} = F_h$$

respectively, where $\theta_v$ is the mixed layer virtual potential temperature; $h$ is the mixed layer depth; $w' \theta_{v0}^0$ and $w' \theta_{vh}$ are the kinetic entrainment buoyancy flux at the mixed layer top and the surface kinetic buoyancy flux, respectively; $W_h$ is the mean vertical velocity at the mixed layer top with downward direction defined as positive; $\Delta \theta_v$ is the difference of the virtual potential temperature across the

**Figure 12.** Left column: the changes in marine low cloud amount from 40$^\circ$S to 40$^\circ$N, the SST change after removing the mean SST increase over the ocean basin, and the changes in atmospheric stability binned based on the 500 hPa vertical velocity and the atmospheric stability of the doubled CO$_2$ climate simulated by CAM3-SOM. Middle and right columns: the same as left column but for CAM3-UW-SOM, AM2-SOM, respectively. White dashed lines in the figures denote the zero lines.
inversion; $F_h$ is the cloud top net longwave radiative flux that tends to cool the cloud layer. If neglecting the time derivative term $\frac{\partial h}{\partial t}$, equation 2 reduces to the equilibrium state described by Betts [1989, equation 2], while neglecting the radiative cooling term $F_h$, equation 2 reduces to cloud free mixed layer situation. Using the mixed layer theory, Zhu and Albrecht [2002] showed that the inversion jump $\Delta h_i$, and the mean potential temperature lapse rate above the inversion $\gamma_t$ under the cloud-free conditions can be related by $\Delta h_i = \gamma_t \frac{\beta h}{1 + \beta^2}$, where $\alpha$ is a coefficient within a range between 1 and 2, and $\beta$ is the ratio of entrainment buoyancy flux to the surface buoyancy flux, i.e., $\beta = \frac{w q_v}{w q_v}$. Assuming this relation can be extended to the cloud topped mixed layer with a correction and represented as

$$\Delta h_i = \xi \frac{\gamma_t \beta h}{1 + \alpha \beta},$$

where $\xi$ is a correction coefficient accounting for the cloud top radiative cooling effect, the possible change in $\beta$ due to clouds, and satisfying the condition $\xi = 1$ for $F_h = 0$. Combining equations (1), (2), and (3), it can be shown that the mixed layer growth is governed by

$$\frac{\partial h}{\partial t} = (1 + \xi \beta) \frac{\partial \theta_i}{\partial t} - W_h + F_h \Delta h_i. \quad (4)$$

[30] The climatologic mean of an arbitrary variable $x$ can be written as

$$\bar{x} = \frac{1}{T} \int_0^T x dt, \quad (5)$$

where $T$ is the time period used for calculating statistical mean of a climate state, for example, $T$ is taken as 20 years for this doubling CO$_2$ analysis. Considering two climate states: the doubled CO$_2$ climate and current climate, the climatologic perturbation of a generic variable $x$ associated with the climate change is defined as

$$\delta x = x_2 - x_1. \quad (6)$$

Applying the climate perturbation defined by equations (5) and (6) to the mixed layer growth $\frac{\partial h}{\partial t}$, it can be shown that the climatologic perturbation of $\frac{\partial h}{\partial t}$ is controlled by

$$\frac{\partial h}{\partial t} = \frac{1}{\xi} \frac{\gamma_t}{\gamma_t (1 + \beta)} \left[ \frac{\partial \theta_i}{\partial t} - \frac{\partial \theta_i}{\partial t} \frac{\beta h}{\gamma_t} \delta \gamma_v \right] - \delta W_h + \delta (\frac{F_h}{\Delta h_i}). \quad (7)$$

where we have assumed coefficient $\xi$ is a slow function of climate state, i.e., not sensitive to climate change to the first-order approximation. Equation (7) indicates that the climate perturbation of the mixed layer growth depends on three factors: the climate perturbation of large-scale thermodynamic state, circulation, and cloud top radiative cooling represented by the first, second, and third term on the right-hand side of the equation, respectively.

[31] The change in low cloud amount is determined by the complicated water vapor and temperature budget within the mixed layer. Since the mixed layer air contains more moisture in a warmer climate due to the enhanced surface evaporation, the change in low cloud amount is largely controlled by the entrainment at the top of the mixed layer although other processes are also involved. Entrainment fundamentally plays two conflicting roles in the evolution of inversion clouds. Weak entrainment, and thus, a shallower mixed layer depth, favors low cloud formation since it helps to trap moisture transported upward by turbulence within the mixed layer. On the other hand, a shallower mixed layer depth allows more moisture to evaporate, the change in low cloud amount is largely controlled by the entrainment at the top of the mixed layer although other processes are also involved. Entrainment fundamentally plays two conflicting roles in the evolution of inversion clouds. Weak entrainment, and thus, a shallower mixed layer depth, favors low cloud formation since it helps to trap moisture transported upward by turbulence within the mixed layer. On the other hand, a shallower mixed layer depth allows more moisture to evaporate, the change in low cloud amount is largely controlled by the entrainment at the top of the mixed layer although other processes are also involved. Entrainment fundamentally plays two conflicting roles in the evolution of inversion clouds. Weak entrainment, and thus, a shallower mixed layer depth, favors low cloud formation since it helps to trap moisture transported upward by turbulence within the mixed layer. On the other hand, a shallower mixed layer depth allows more moisture to evaporate, the change in low cloud amount is largely controlled by the entrainment at the top of the mixed layer although other processes are also involved. Entrainment fundamentally plays two conflicting roles in the evolution of inversion clouds. Weak entrainment, and thus, a shallower mixed layer depth, favors low cloud formation since it helps to trap moisture transported upward by turbulence within the mixed layer. On the other hand, a shallower mixed layer depth allows more moisture to evaporate, the change in low cloud amount is largely controlled by the entrainment at the top of the mixed layer although other processes are also involved. Entrainment fundamentally plays two conflicting roles in the evolution of inversion clouds. Weak entrainment, and thus, a shallower mixed layer depth, favors low cloud formation since it helps to trap moisture transported upward by turbulence within the mixed layer. On the other hand, a shallower mixed layer depth allows more moisture to evaporate, the change in low cloud amount is largely controlled by the entrainment at the top of the mixed layer although other processes are also involved. Entrainment fundamentally plays two conflicting roles in the evolution of inversion clouds. Weak entrainment, and thus, a shallower mixed layer depth, favors low cloud formation since it helps to trap moisture transported upward by turbulence within the mixed layer. On the other hand, a shallower mixed layer depth allows more moisture to evaporate, the change in low cloud amount is largely controlled by the entrainment at the top of the mixed layer although other processes are also involved. Entrainment fundamentally plays two conflicting roles in the evolution of inversion clouds. Weak entrainment, and thus, a shallower mixed layer depth...
the change in averaged vertical velocity over these three regions where the increase in low cloud amount is the largest. CAM3-SOM and CAM3-UW-SOM basically give the similar change in vertical velocity in these areas, but \( \delta W_h \) can be either greater or smaller than zero; while AM2-SOM actually predicts \( \delta W_h < 0 \) above 850 hPa in all three regions, indicating that the change in large-scale circulation is not a reason for generating a tendency of reducing the mixed layer depth.

The second possibility to reduce the mixed layer growth in the subsidence regime in the warmer climate is due to the change in the atmospheric thermodynamic state represented by the first term on the right-hand side of equation (7). To meet \( \delta \frac{\partial \theta_v}{\partial t} < 0 \), it requires that

\[
\frac{\partial \theta_v}{\partial t} - \left[ \frac{\omega}{\partial \gamma_{\theta}} \right] \delta \gamma_{\theta} < 0.
\]  

Considering a finite time interval \( \Delta t \) in the two climate states, relation (8) can be further simplified as

\[
\delta \theta_v < \frac{\omega}{\gamma_{\theta}} \delta \gamma_{\theta}.
\]  

Assuming that the mixed layer temperature increases 1 K due to doubled CO\(_2\) and \( \theta_v \) takes 300 K, then, to satisfy relation (9), the above-inversion stability \( \gamma_{\theta} \) only needs to increase 0.33% of its base value. In other words, the climate perturbation of the mixed layer growth is very sensitive to the climate perturbation of the atmospheric stability above the inversion. Therefore we reach an important result: For a certain increase in \( \gamma_{\theta} \) due to global warming, the smaller increase in the mixed layer temperature, the easier to meet relation (9). As a result, low cloud formation prefers the areas where the mixed layer temperature is less increased since in these areas there is a tendency to reduce the mixed layer depth owing to relation (9). To verify whether the above-inversion atmospheric stability \( \gamma_{\theta} \) does increase due to doubled CO\(_2\), we examined the vertical profiles of \( \theta_v \) and calculated the lapse rate of \( \theta_v \) in the current and doubled CO\(_2\) climates in the three subsidence regimes. The results are shown in Figure 15. The consistent increase in atmospheric stability due to doubled CO\(_2\) is confirmed by all three models, although the detailed structure is different. The simulated increase in atmospheric stability in the warmer climate is consistent with Santer et al.’s [2005] finding. Thus the change in large-scale thermodynamic state is one of the reasons that cause the increase in low cloud amount to prefer the locations where SST is less increased.

According to equation 7, the third factor that may cause a change in the mixed layer depth is the change in the cloud top radiative cooling represented by the third term on the right-hand side of the equation. The effect of this term complicates the climate perturbation of the mixed layer growth, and thus, the climate sensitivity of low clouds. It should be pointed out that the effect of this term is fundamentally different from the other two terms. The change in subsidence and the atmospheric stability above the inversion can be considered as the external forcings that control the mixed layer growth, while the third term
represents an internal mechanism that involves not only the change in low cloud amount itself but the change in cloud condensate as well. Thus the effect of this term cannot be interpreted solely based on the mixed layer model or simple diagnosis using GCM output. Currently, we do not have an appropriate method to determine the sign of the effect of cloud top radiative cooling on the change in low cloud amount. To quantify this effect, the off-line radiative transfer calculation may be needed, but this is beyond the scope of this study.

5. Discussion

The previous analyses on the change in low cloud amount in response to the short-term climate perturbation associated with the 1997–1998 El Niño and the long-term climate change caused by doubled CO$_2$ based on observations and simulations raise several interesting questions. First, according to this study, the increase in low cloud amount in the subsidence regimes due to doubled CO$_2$ favors the location where SST is less increased. The mixed layer model analysis illustrates a thermodynamic mechanism that the increase in the atmospheric stratification above the inversion is responsible for such a change in low cloud amount. This phenomenon and the increase in the atmospheric stratification above the inversion in the warmer climate are well agreed by all three models. What they disagreed on is the specific spatial change pattern of SST, and thus, the details of the change in low cloud amount. Optimistically, if models are to predict a consistent spatial SST change pattern, as well as the change in the above-inversion atmospheric stratification, then, GCMs should be able to generate a robust change in low cloud amount as suggested by this study. The question is: what determines the SST change in a coupled system? This is not an easy question since clouds themselves have an effect on the underneath temperature, the interaction between different processes makes things very complicated. As an important step to attack this problem, a careful examination of surface parameterization and boundary layer turbulent mixing scheme in coupled models are needed.

Second, the cloud effect on global warming due to increasing greenhouse gases is clearly shown in the doubling CO$_2$ experiments. The increase in the global mean surface temperature due to doubled CO$_2$ simulated by AM2-SOM is 2.87°C, which is higher than 2.47°C and 2.37°C predicted by CAM3-SOM and CAM3-UW-SOM, respectively. Although the global mean surface temperature change in response to doubled CO$_2$ is determined by many feedback processes, the larger surface temperature increase in the AM2-SOM simulation appears to be consistent with its predicted larger reduction in low cloud amount. Using multiple linear regression approach, Zhu et al. [2007] decomposed the low cloud radiative feedback into contributions associated with cloud amount and cloud condensate. They found that the positive low cloud amount feedback simulated by AM2-SOM dominates the negative low cloud condensate feedback to result in a net positive low cloud feedback, while CAM3-SOM simulated a negative low cloud feedback in response to doubled CO$_2$, which is mainly determined by the negative low cloud condensate feedback. To understand the different behavior of cloud feedback in different GCMs will be the focuses of our future research.

Third, in this study we compared the simulations from CAM3 and CAM3-UW, which use different turbulence and shallow convection schemes and different vertical resolutions. Yet, both CAM3 and CAM3-UW predict a similar change in low cloud amount (generated by the ISCCP simulator) associated with the 1997–1998 El Niño to that of the ISCCP observations. The similarity between CAM3 and CAM3-UW in the SST forced simulations suggests that the interactions between dynamic processes and model physical parameterizations may be more important than a parameterization alone in determining low cloud feedbacks. This is clearly shown in the coupled simulations where interactions among different processes are fully realized. CAM3-SOM and CAM3-UW-SOM show noticeable difference in the doubling CO$_2$ experiments although they share similarities. For example, the new turbulence and shallow convection schemes used in CAM3-UW-SOM exert a different impact on the change in SST and low clouds in response to doubled CO$_2$ from that of CAM3-SOM.

Figure 15. Change in the lapse rate ($\gamma_0$) above the inversion due to doubled CO$_2$ in the three regions: southeast Pacific, north and south Atlantic (see Figure 11) from the CAM3-SOM, CAM3-UW-SOM, and AM2-SOM simulations.
SOM. This suggests that future research should pay more attention to the interactions between dynamic processes and physical parameterizations rather than physical parameterization alone. To properly address the issues associated with interaction among different parameterizations, novel evaluation approaches need to be developed.

6. Summary

[39] In this study, we systematically investigated the change in low cloud amount in response to the short-term climate perturbation resulting from the 1997–1998 El Niño and the long-term climate change caused by doubled CO\textsubscript{2} using the ISCCP satellite cloud measurements, the ECMWF reanalyses, the SST forced simulations by the NCAR and GFDL GCMs, as well as the coupled numerical experiments executed by these models. Our major objectives are to understand the mechanisms that operate in the change in tropical and subtropical low cloud amount associated with ENSO and climate warming due to doubled CO\textsubscript{2}, and identify model biases shown in different GCM simulations.

[40] The ISCCP cloud observations show that the decrease in low cloud amount in the eastern tropical Pacific associated with the 1997–1998 El Niño mainly comes from optically thick “stratocumulus” and “stratus”. This decrease in low cloud amount is robust even after fully accounting for the possible mid and high cloud obscuring effect due to the ‘top-down’ satellite view. Our analyses indicate that the change in low cloud amount associated with the 1997–1998 El Niño and doubled CO\textsubscript{2} not only has different characteristics, but also is controlled by different physical processes. During the 1997–1998 El Niño, the change in low cloud amount is negatively correlated to the local SST anomaly, especially in the eastern tropical Pacific. Most reduction in low cloud amount during the 1997–1998 El Niño occurs in the eastern tropical Pacific associated with an upward large-scale motion and a weak atmospheric stability measured by the 500 hPa vertical velocity and the potential temperature difference between 700 hPa and the surface. Such a sensitivity of low clouds to the El Niño event can be attributed to the change in the convective activity associated with the SST anomaly. During the 1997–1998 El Niño, observations indicate that the SST increase in the eastern tropical Pacific enhances the atmospheric convection, which shifts the upward motion to further south and breaks down low stratiform clouds, leading to a decrease in low cloud amount in this region. The enhanced convection in the eastern tropical Pacific is supported by the increase in higher level clouds shown in the ISCCP cloud measurements. This mechanism for the reduction in low cloud amount in the eastern tropical Pacific during the 1997–1998 El Niño can be considered as a complement to Deser et al.’s [1993] findings that the reduction in low cloud amount is associated with the weak cold advection across the equatorial front.

[41] In the doubling CO\textsubscript{2} experiments, the coupled NCAR and GFDL models predict an increase in low cloud amount in the subtropical ocean basin in response to global warming. Unlike the change in low cloud amount during the El Niño, most increase in low cloud amount due to doubled CO\textsubscript{2} occurs in the subsidence regimes associated with a strong atmospheric stability. All simulations show that the change in low cloud amount is closely correlated to the spatial change pattern of SST consistent with Williams et al.’s [2003] results. There is a negative correlation between the change in low cloud amount and SST change after removing the mean temperature increase. Using the simple atmospheric mixed layer model analysis, we showed why the increase in low cloud amount appears to favor the areas where SST is less increased. The theoretical analyses indicate that the increase in the atmospheric stratification above the inversion in the doubling CO\textsubscript{2} experiment tends to reduce the mixed layer depth at the locations with a small increase in mixed layer temperature. Such a thermodynamic process helps to trap moisture transported upward by turbulence within the mixed layer, and thus, favors low cloud formation. Based on this analysis, we argued that the different low cloud responses to doubled CO\textsubscript{2} shown in current GCMs are caused by the combination of different SST change pattern and different above-inversion stability change predicted by the models. Thus generating a robust feedback between low-level clouds and climate in the coupled simulations relies on a correct determination of the change in SST and above-inversion atmospheric stability.

[42] Using the simulations forced by the observed SST, we evaluated the performance of the NCAR GCMs (CAM3, CAM3-UW) and GFDL GCM (AM2), and their capability of reproducing the observed change in low cloud amount in response to the short-term perturbation associated with the 1997–1998 El Niño. The NCAR GCMs appear to simulate a total low cloud amount sensitivity to the El Niño perturbation closer to the ISCCP observations than the GFDL GCM. However, both NCAR and GFDL models showed marked difference to the ISCCP observations when clouds are broken down into different categories based on cloud top pressure and optical thickness.

[43] This study only addresses some issues associated with the change in low cloud amount in response to climate perturbations. Issues related to the change in cloud condensate have not been touched here. To thoroughly understand cloud-climate feedbacks, studies on climate sensitivity of cloud condensate also need to be carried out.

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