An Estimate of Low-Cloud Feedbacks from Variations of Cloud Radiative and Physical Properties with Sea Surface Temperature on Interannual Time Scales

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ABSTRACT

Simulations of climate change have yet to reach a consensus on the sign and magnitude of the changes in physical properties of marine boundary layer clouds. In this study, the authors analyze how cloud and radiative properties vary with SST anomaly in low-cloud regions, based on five years (March 2000–February 2005) of Clouds and the Earth’s Radiant Energy System (CERES)–Terra monthly gridded data and matched European Centre for Medium-Range Weather Forecasts (ECMWF) meteorological reanalysis data. In particular, this study focuses on the changes in cloud radiative effect, cloud fraction, and cloud optical depth with SST anomaly. The major findings are as follows. First, the low-cloud amount (−1.9% to −3.4% K−1) and the logarithm of low-cloud optical depth (−0.085 to −0.100 K−1) tend to decrease while the net cloud radiative effect (3.86 W m−2 K−1) becomes less negative as SST anomalies increase. These results are broadly consistent with previous observational studies. Second, after the changes in cloud and radiative properties with SST anomaly are separated into dynamic, thermodynamic, and residual components, changes in the dynamic component (taken as the vertical velocity at 700 hPa) have relatively little effect on cloud and radiative properties. However, the estimated inversion strength decreases with increasing SST, accounting for a large portion of the measured decreases in cloud fraction and cloud optical depth. The residual positive change in net cloud radiative effect (1.48 W m−2 K−1) and small changes in low-cloud amount (−0.81% to 0.22% K−1) and decrease in the logarithm of optical depth (−0.035 to −0.046 K−1) with SST are interpreted as a positive cloud feedback, with cloud optical depth feedback being the dominant contributor. Last, the magnitudes of the residual changes differ greatly among the six low-cloud regions examined in this study, with the largest positive feedbacks (−4 W m−2 K−1) in the southeast and northeast Atlantic regions and a slightly negative feedback (−0.2 W m−2 K−1) in the south-central Pacific region. Because the retrievals of cloud optical depth and/or cloud fraction are difficult in the presence of aerosols, the transport of heavy African continental aerosols may contribute to the large magnitudes of estimated cloud feedback in the two Atlantic regions.

1. Introduction

Low-level marine clouds play an important role in the earth’s radiative energy budget because they reflect much of the incoming solar radiation that would otherwise be absorbed by the underlying ocean (Randall et al. 1984). The response of these clouds to climate change and the enhancement/diminishment of climate change due to radiative imbalance caused by these clouds—cloud feedbacks—varies greatly among general circulation models (GCMs). Cloud feedbacks remain a major source of uncertainty for the estimated climate sensitivity (Bony and Dufresne 2005) and the projected increase in global mean temperature in response to anthropogenic forcings (Randall et al. 2007).

Since low-level clouds are close to the earth’s surface, their main radiative effects are associated with the reflected solar radiation at the top of the atmosphere (TOA). The TOA albedo—the ratio of the reflected solar radiation to solar insolation—of marine low-cloud areas is primarily determined by the cloud fraction and cloud optical depth (τ). An increase in the value of either quantity increases the TOA albedo and increases the cooling at the ocean’s surface. When these quantities both have changes
that are of the same sign, the net change in albedo can be large. Opposing trends in these two quantities can produce a net change in albedo that is small, as was the case for a GCM examined in Zhu et al. (2007a).

The characteristics of low clouds—in particular, those related directly to radiative energy balance—have been studied with surface and satellite observations on a variety of time scales to better quantify cloud feedbacks. Recently, Clement et al. (2009) used surface- and satellite-based data to construct a multidecade record of cloud cover over the northeast (NE) Pacific Ocean (115°–145°W, 15°–25°N). They found that the amount of low-level cloud in the region negatively correlated with SST at −0.82 with a 99% significance level and that this acts as a positive feedback. Using the Earth Radiation Budget Experiment (ERBE) and the International Satellite Cloud Climatology Project (ISSCP) data from 1984–89, Oreopoulos and Davies (1993) found that albedo and cloud cover tended to decrease with SST over the southeastern (SE) Pacific (75°–95°W, 5°–30°S) and SE Atlantic (15°W–15°E, 5°–20°S) Oceans on seasonal and interannual time scales. They estimated the solar absorption variation with SST at 6.1 W m⁻² K⁻¹ for interannual variations. Tselioudis et al. (1992) used a year of ISCCP data to show that optical depths tend to decrease with cloud temperature for low clouds over the ocean in the subtropics. Decreases in both cloud fraction and cloud optical depth with SST were found for subsiding regimes by Bony et al. (1997) using ERBE and ISCCP data for the 1987/88 El Niño/La Niña. Eitzen et al. (2008) used the Clouds and the Earth’s Radiant Energy System (CERES)—Tropical Rainfall Measuring Mission (TRMM) data to examine instantaneous relationships between different physical properties of marine boundary layer clouds and found that optical depth generally decreased with SST.

The changes in cloud and radiative properties with SST are entwined with changes in both dynamic and thermodynamic states of the atmosphere. This has been recognized by many studies and many attempts have been made to link the changes with SST in cloud and radiative properties with those in selected dynamic and thermodynamic measures (e.g., Bony et al. 1997, 2004; Norris and Iacobellis 2005; Clement et al. 2009; Su et al. 2010a). Clement et al. (2009) found that the low-cloud amount is positively correlated with vertical velocity at 500 hPa (ω₅₀₀) and sea level pressure on interannual time scales with a 99% significance level. Monthly mean ω₅₀₀ is often used to divide the tropical atmosphere into dynamical regimes (e.g., Bony et al. 1997). For ascending (ω₅₀₀ < 0) regimes, Bony et al. (2004) showed that the strengths of both the longwave (LW) and shortwave (SW) cloud radiative effect (CRE)—the difference in radiative fluxes between clear and all skies—increase in magnitude as the magnitude of the ascent increases. Stowasser and Hamilton (2006) used ERBE observations of SW CRE and the 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40) analyses of ω₅₀₀ to show that SW CRE does not change significantly with ω₅₀₀ for subsiding regimes. Su et al. (2010a) also confirmed results of Bony et al. (1997) and Stowasser and Hamilton (2006) with CERES–TRMM CRE and ERA-Interim and NCEP ω₅₀₀, as well as CRE and ω₅₀₀ simulated by the U.K. Met Office GCM.

The lower-tropospheric stability (LTS), defined as the difference in potential temperature between the 700-hPa level and the surface, is a reasonable choice for characterizing boundary layer cloud regimes because of the sensitive dependence of low-cloud regimes on low-level stability. LTS has been found to be closely related to the amount of low-cloud cover on seasonal time scales by Klein and Hartmann (1993) among others. Note that this relationship is considerably weaker on daily time scales (Klein 1997). A considerably improved relationship between low-cloud amount and stability was obtained by using the estimated inversion strength (EIS; Wood and Bretherton 2006). The concept of EIS uses the fact that the free atmosphere is typically close to that of a moist adiabat to estimate the magnitude of the strength of the inversion at the top of the boundary layer. EIS is a more regime-independent predictor of low-cloud amount than is LTS under a wide range of climatological conditions. Wood and Bretherton (2006) note that the EIS tends to decrease with temperature for a given LTS. This means that in a warmer climate, if the profile of tropospheric warming follows that of a moist adiabat, the EIS (and hence, low-cloud cover) will remain nearly unchanged, even though the LTS may increase. This is in contrast to the results of Miller (1997) and Larson et al. (1999), who used simple models with low-cloud parameterizations based on the observed relationship between LTS and cloud cover to suggest that there will be a strong negative cloud feedback in a warmer climate.

The interdependence between dynamic and thermodynamic states requires that composing cloud and radiative properties be done within small intervals of dynamic and thermodynamic measures (Williams et al. 2003; Ringer and Allan 2004; Norris and Iacobellis 2005; Su et al. 2010b). In Williams et al. (2003) and Ringer and Allan (2004), changes in observed and simulated cloud cover and CRE are binned in 2D histograms with anomalies in ω₅₀₀ and SST as the axes. Norris and Iacobellis (2005) used ω₅₀₀ and SST advection to composite midlatitude North Pacific cloud properties in bi-intervals while Su et al. (2010b) used ω₇₀₀ and EIS to composite cloud and aerosol properties in the SE Atlantic region. Evaluating models in addition to quantifying cloud feedback was
also a motivation for Clement et al. (2009), who found that very few GCMs reproduced the observed correlations between cloud cover and other meteorological variables—LTS, \( \omega_{900} \), and sea level pressure—in the NE Pacific Ocean over multidecade periods.

The main objective of this study is to obtain an estimate of low-cloud feedbacks from a combination of satellite and model reanalysis data. In this paper, we will use CERES–Terra data to explore the variations in observed radiative fluxes and cloud physical properties with SST for several low-cloud regions on interannual time scales. Although similar studies examined the dependence of fewer cloud/radiative properties on SST or their correlations with SST (e.g., Tselioudis et al. 1992; Oreopoulos and Davies 1993; Norris and Leovy 1994; Bony et al. 1997; Norris and Iacobellis 2005; Stowasser and Hamilton 2006; Eitzen et al. 2008; Wagner et al. 2008; Clement et al. 2009; Ghate et al. 2009; Lin et al. 2009), all of them except for Norris and Iacobellis (2005) did not consider the contributions of dynamic and thermodynamic variations to the changes in radiative fluxes and cloud physical properties with SST. After accounting for these contributions from the systematic changes in thermodynamic and dynamic states with SST, this study will provide a residual estimate of the cloud feedback.

In section 2, the data sources used in this study will be described. Section 3a will present 2D histograms of anomalies in cloud and radiative properties according to thermodynamic and dynamic states. The overall changes of cloud and radiative properties with SST anomaly for all cloud regions combined are presented in section 3b, and the separation of these overall changes into thermodynamic, dynamic, and residual feedback components is shown in section 3c. Section 3d presents changes in low clouds for each individual region. Finally, a summary and conclusions are given in section 4.

2. Data

This study will focus on a 5-yr period (March 2000–February 2005) that corresponds to the first 5 yr of CERES–Terra data (Wielicki et al. 1996; Loeb et al. 2009). CERES provides a number of different monthly gridded (level 3) data products with varying degrees of accuracy. We use the following two CERES–Terra datasets. The first CERES–Terra dataset used in this study is the CERES SRBAVG Edition2D_Rev1 monthly gridded cloud property data. The SRBAVG dataset consists of monthly mean CERES fluxes that are provided on a \( 1^\circ \times 1^\circ \) grid with consistent cloud and aerosol properties from the Moderate Resolution Imaging Spectroradiometer (MODIS). The SRBAVG data are available both with and without geostationary satellite data added (referred to as SRBAVG-GEO and SRBAVG-NonGEO, respectively; hereafter, called GEO and CERES–MODIS). The CERES–MODIS data are based solely upon the twice-daily (1030 and 2230 LT) measurements from the Terra satellite, which has a sun-synchronous orbit (Young et al. 1998). The GEO data combine additional geostationary satellite data measurements with the CERES–MODIS data to minimize temporal sampling errors and incorporate the nonconstant meteorological conditions between the two Terra measurements. The availability of two estimates of the same quantity provides added confidence of the results to be presented later.

The second CERES–Terra dataset used in this study is the CERES–Energy Balanced and Filled (EBAF) radiative flux climatology. This is a radiative flux–only dataset and is based on the GEO fluxes. CERES–EBAF does not provide a version based on CERES–MODIS fluxes. The instantaneous CERES fluxes are estimated from unfiltered radiances using empirical angular distribution models (ADMs; Loeb et al. 2003, 2005) that account for variations in the surface type and cloud properties inferred from MODIS measurements (Minnis et al. 2003, 2008, 2010, manuscript submitted to IEEE Trans. Geosci. Remote Sens.). Loeb et al. (2003) estimated that \( 1^\circ \times 1^\circ \) instantaneous TOA fluxes are accurate to within 10 W m\(^{-2}\) in the SW and 3.5 W m\(^{-2}\) in the LW with little dependence on cloud phase, cloud optical depth, or infrared cloud emissivity. The monthly \( 1^\circ \times 1^\circ \) TOA fluxes are accurate to within approximately 2 W m\(^{-2}\) in both the SW and the LW. CERES–EBAF is obtained by adjusting the GEO TOA fluxes in a manner consistent with the instrumental uncertainty and observations of ocean heat storage (Loeb et al. 2009). In addition, clear-sky fluxes are computed with a weighted average of CERES broadband fluxes from completely cloud-free footprints and MODIS-derived fluxes estimated from the cloud-free portions of partly cloudy footprints (Loeb et al. 2009). The MODIS narrowband radiances are converted to broadband radiances using regression coefficients, and then the “broadband” MODIS radiances are converted to TOA radiative fluxes using CERES clear-sky ADMs (Loeb et al. 2003, 2005). The values of CRE—the difference between clear-sky and all-sky radiative fluxes—are expected to be more realistic with the clear-sky grid cells from the MODIS imager, especially over the nearly overcast regions examined in the present study. For areas with low clouds, the SW CRE is strongly negative, while the LW CRE is weakly positive, causing the net CRE to be negative. The term cloud radiative effect is used rather than cloud radiative forcing here to be consistent with the term used in the data product; Stephens (2005) also argues that cloud radiative forcing is a misnomer since climate forcings are normally understood to be changes imposed on the climate system.
(such as an increase in CO₂) rather than properties of the system.

For cloud properties such as cloud fraction and optical depth \( \tau \), the CERES–MODIS data are believed to be more accurate than the GEO data owing to the higher quality of the MODIS imager data than the geostationary satellite data. However, there is a significant diurnal cycle in marine low-cloud fraction (e.g., Ghate et al. 2009), so the temporal sampling errors in cloud properties may be significant in the CERES–MODIS data. Therefore, the GEO cloud properties will also be analyzed to provide two plausible estimates of the same quantity. The CERES–MODIS low-level cloud optical depth has a random error of approximately 26% with a bias of −3.6% when compared with surface instrument results over the Atmospheric Radiation Measurement (ARM) Southern Great Plains (SGP) site (Dong et al. 2008), which is the only validation of CERES–MODIS measurements of low clouds to date. While the differences between CERES–MODIS and GEO cloud fraction are close to zero during the daytime, cloud fraction differences at night between these two estimates can be as large as 0.09 according to the CERES–Terra SRBAVG data quality summary (available online at http://eosweb.larc.nasa.gov/PRODOCS/terrsrbavg/Quality_Summaries/CER_SRBAVG_Edition2D_Terra_Aqua.html). GEO tends to underestimate nighttime cloud fraction owing to the limited geostationary spectral coverage after dusk. For daytime optical depth, GEO tends to underestimate the CERES–MODIS value by as much as 15% because of larger geostationary imager pixels (CERES–Terra SRBAVG data quality summary). Geostationary satellites are unable to retrieve cloud optical depth at night because their optical depths are derived from the visible channel, so \( \ln \tau \) is linearly interpolated in time between the last daytime geostationary retrieval and the nighttime Terra observation. After \( \tau \) is updated by the nighttime Terra observation, \( \ln \tau \) is then linearly interpolated in time until the first daytime geostationary observation. Note that the nighttime MODIS retrievals of optical depth are not of the same quality as the daytime retrievals because only the infrared channels can be used. For the SST data, the National Oceanic and Atmospheric Administration (NOAA) optimum interpolation version 2 SST analyses on a 1° × 1° grid are used (Reynolds et al. 2002).

Because we are interested in the behavior of marine low clouds, we wish to eliminate grid cells with large amounts of middle and high clouds. Therefore, only those grid cells and months that have a climatological (based on the 5 yr of observations) mean cloud cover of less than 10% for clouds with cloud-top pressures (\( p_c \)) less than 680 hPa are considered. The results presented later are not very sensitive to the precise value of this cloud cover threshold; low-cloud cover and optical depth vary with SST at a rate that is nearly constant (within a few percent of the respective trends calculated with a 10% threshold) for thresholds between 5% and 15%.

There are six regions that are studied here (Table 1 and Fig. 1), five of which (SE Atlantic, SE Indian, NE Pacific, NE Atlantic, and SE Pacific) are the regions defined in Jensen et al. (2008). An additional region, labeled the south-central (SC) Pacific is included in this study owing to its lack of high clouds. The SC Pacific region has much less average low-cloud cover than the SE Pacific region (36.4% versus 67.9% according to CERES–MODIS data), so it represents a climate regime that has more shallow cumuli rather than overcast stratuscumulus. The number of samples per year within each 1° × 1° grid is shown in Fig. 1. The maximum possible number of samples per grid cell over the period of this study is 60 (5 yr × 12). The total number of samples for each region is listed in Table 1 for the selection criteria listed above applied to

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**Table 1. Low-cloud regions. Latitude and longitude boundaries are given for each region. The number of grid points refers to the number of 1° × 1° cells that fit the criteria listed in the text using the CERES–MODIS and GEO cloud property data. The locations are plotted in Fig. 1.**

<table>
<thead>
<tr>
<th>Region</th>
<th>Lat/lon bounds</th>
<th>No. of grid points (non-GEO; GEO)</th>
</tr>
</thead>
<tbody>
<tr>
<td>NE Pacific</td>
<td>20°–40°N; 140°–115°W</td>
<td>9045; 9395</td>
</tr>
<tr>
<td>NE Atlantic</td>
<td>5°–30°N; 45°–10°W</td>
<td>6320; 7495</td>
</tr>
<tr>
<td>SC Pacific</td>
<td>25°S–5°N; 140°–100°W</td>
<td>37 610; 40 875</td>
</tr>
<tr>
<td>SE Pacific</td>
<td>35°S–5°N; 100°–70°W</td>
<td>34 445; 36 220</td>
</tr>
<tr>
<td>SE Atlantic</td>
<td>25°S–0°; 20°W–15°E</td>
<td>36 500; 38 535</td>
</tr>
<tr>
<td>SE Indian</td>
<td>40°–20°S; 85°–115°E</td>
<td>8535; 11 715</td>
</tr>
</tbody>
</table>

**Fig. 1. Number of months of the year with low amounts of clouds with tops above the 680-hPa level. Boxes indicate the six regions in this study. See Table 1 for latitude and longitude bounds of each region. Only grids over ocean are included in the analysis.**
both GEO and CERES–MODIS cloud climatologies. In the subsequent analyses, the CERES–MODIS samples are used when dealing with CERES–MODIS cloud properties and EBAF CREs, while the GEO samples are used for GEO cloud properties.

The meteorological data used to classify the dynamic and thermodynamic states of the atmosphere is from the ERA-Interim reanalysis, which represents the current state-of-the-art in ECMWF reanalysis, with improvements in moist processes, data assimilation, and bias corrections (Uppala et al. 2008). Here, we focus on the vertical velocity at 700 hPa ($\omega_{700}$) and EIS to characterize the dynamic and thermodynamic states of the lower troposphere, respectively. 2D histograms of these states for the low-cloud areas are shown in Fig. 2a. The EIS is defined by Wood and Bretherton (2006) as

$$\text{EIS} = \theta_{700} - \theta_0 - \Gamma^{850}_m (z_{700} - LCL).$$

Here, $\theta_{700}$ is the potential temperature at 700 hPa, and $\theta_0$ is the potential temperature at the surface. The difference between these two quantities is LTS defined by Klein and Hartmann (1993). Here, $\Gamma^{850}_m$ is the moist-adiabatic potential temperature gradient at 850 hPa, and $z_{700}$ and LCL are the heights of the 700-hPa surface and surface-based lifting condensation level, respectively.

The low-cloud basins shown in Fig. 1 are generally associated with subsiding motion and positive values of EIS with the majority of occurrences between 10 and 50 hPa day$^{-1}$ and between 0 and 10 K. Note that the variability of $\omega_{700}$ in these regions is considerably larger (including some negative values) on shorter time scales (Su et al. 2010b). The positive EIS means that an inversion is likely to be present between the surface and 700 hPa, although the exact inversion height and strength are unknown. In general, the higher EIS is, the stronger the inversion is (Wood and Bretherton 2006).

The monthly average value is calculated for the SST (NOAA optimum interpolation version 2 analysis), EIS, $\omega_{700}$ (from the ERA-Interim data at $1.5^\circ \times 1.5^\circ$ spacing, which were interpolated to $1^\circ \times 1^\circ$ grids used by CERES), and cloud physical and radiative properties. The monthly anomalies (denoted by primes) are then calculated by subtracting the monthly average from each month’s value. The anomalies measure the variabilities of the low clouds and their environments with temporal scales longer than the seasonal cycles. For simplicity, $E'$ and $\omega'$ are used to denote the anomalies of EIS and $\omega_{700}$, respectively, while $T'_j$ is used to denote the SST anomaly. The frequencies of occurrence within $E'$–$\omega'$ bins are shown for the low-cloud regions combined in Fig. 2b. The majority of occurrence appears in very small ranges centered at 0 hPa day$^{-1}$ and 0 K. That is, most monthly anomalies are relatively modest in magnitude for both $E'$ ($\pm 1.0$ K) and $\omega'$ ($\pm 10$ hPa day$^{-1}$). The small magnitudes of the variabilities in $E'$ and $\omega'$ are as expected because low clouds are persistent phenomena in the selected regions from year to year, although the seasonal variations (which have been removed in this study) can be very large (e.g., Jensen et al. 2008).

3. Results

a. Mean anomalies

The behavior of radiative flux and cloud property anomalies associated with changes in $\omega'$ and $E'$ is of interest because it allows us to simultaneously see the impact of changes in the dynamic and thermodynamic states on the radiative and cloud property anomalies. This is somewhat similar to Fig. 9 in Williams et al. (2003), but the axes in Williams et al. are based on anomalies of SST and $\omega_{500}$. Norris and Iacobellis (2005) and Su et al. (2010b)
used the full quantities as the axes to examine daily variations of cloud physical properties. When a composite is made with a single variable, the behaviors of cloud property and radiative flux can be different from that obtained when another variable is held constant—a bivariable compositing—owing to averaging over the entire range of the second variable. This statement is readily verified from the results shown in Figs. 3 and 4.

The average CRE anomalies within $\omega'-E'$ bins are shown in Fig. 3 for SW CRE, LW CRE, and net CRE. Only those bins for which the absolute value of the mean anomaly is greater than the uncertainty of the mean anomaly are shown. Here, the uncertainty is calculated according to the one-sample $t$ test (Wilks 2006) with a 95% confidence interval. The magnitude of this confidence interval is proportional to $s_{\text{eff}}^{-1/2}$, where $s$ is the sample standard deviation, and $n_{\text{eff}}$ is the effective number of samples in a $\omega'-E'$ bin. The total effective sample size ($N_{\text{eff}}$) is calculated using lag-1 autocorrelations (Wilks 2006) in the zonal and meridional directions within each of the six regions for months with climatologically small amounts of high clouds in that region and then adding the regional totals up. The value of $N_{\text{eff}}$ varies from ~6000 for SW CRE to ~38000 for ln$. Then, $n_{\text{eff}}$ is calculated by multiplying the number of samples in the bin by $N_{\text{eff}}/N$, where $N$ is the total number of samples. Temporal autocorrelations were also examined, but they are generally small and ill defined because the number of consecutive months used is small in many of the regions examined here. Since $n_{\text{eff}}$ varies widely, the confidence interval also varies (0.6–14.3 W m$^{-2}$ for SW CRE, 0.2–2.5 W m$^{-2}$ for LW CRE, and 0.4–10.0 W m$^{-2}$ for net CRE) among the bins. This causes many of the bins with low frequencies of occurrence (Fig. 2) to have an uncertain sign in the mean anomaly. SW CRE is seen to change rapidly (from $-10$ to 8 W m$^{-2}$) with $E'$, with positive values (less reflection by clouds) being associated with decreased inversion strength. The change in SW CRE with $\omega'$ appears to be smaller, although there is a tendency for strongly negative (positive) values of $\omega'$ to be associated with less (more) reflection by clouds. The magnitudes of the LW CRE anomalies are within $\pm 1$ W m$^{-2}$ for 58 of the total 64 $\omega'-E'$ bins. These are much smaller than those of SW CRE owing to the small magnitude of LW CRE itself for low clouds. The changes in $\omega_{700}$ appear to have slightly more influence on LW CRE than changes in EIS particularly for negative $\omega'$ values, owing perhaps to changes in cloud-top height for more convective cloud regimes. Because of the small magnitudes of the LW CRE anomalies, the histogram of average net CRE anomalies is fairly similar to that of SW CRE anomalies. This indicates that negative values of $E'$ (less stable) are associated with a net radiative warming
While positive values of $E'$ (more stable) are associated with a net radiative cooling. The $\omega' - E'$ bins with frequencies of occurrence greater than 2\% (Fig. 2b) have net CRE anomalies between $-4$ and $4$ W m\(^{-2}\). The lack of significant changes in the CRE with $\omega'$ is broadly consistent with previous studies (Bony et al. 1997, 2004; Stowasser and Hamilton 2006; Su et al. 2010a), which showed that CREs do not change significantly with $\omega'_{500}$ for subsiding dynamic regimes.

As noted earlier in the introduction, decreases in cloud fraction can be offset by increases in optical depth, as was the case for a model evaluated by Zhu et al. (2007a). The average anomalies of cloud fraction and lnT for low clouds based on CERES-MODIS data are shown in Fig. 4. Similar diagrams based upon GEO data are not shown for the sake of brevity. The reason for showing lnT instead of $\tau$ is that solar absorption is directly related to lnT. The low-cloud fraction is defined as the fraction of clouds with tops that have $p_c$ greater than 680 hPa. The value of lnT represents the monthly mean logarithm of $\tau$ over cloudy areas within each grid cell, which means that its value is not influenced by the magnitude of cloud fraction in the same grid cell. As was the case for Fig. 3, there is a wide range of confidence intervals in the bins in Fig. 4 (0.46–9.24\% for cloud fraction, 0.005–0.100 for lnT). In Fig. 4, we see that both low-cloud fraction and lnT tend to increase with $E'$. That is, cloud fraction (lnT) anomalies range from $-10\%$ to $10\%$ ($-0.15$ to $0.2$) as $E'$ increases from $-2$ to $2$ K, while the most frequently occurring anomalies vary between $-4\%$ and $+4\%$ ($-0.1$ to $0.1$). (It is noted that the range of cloud fraction anomalies from GEO data is slightly smaller, owing to its higher temporal sampling frequency, but that of low-cloud lnT from GEO data remains roughly the same.) This indicates that as the low-cloud fraction decreases, lnT of the remaining clouds also decreases, which is consistent with the cloud object analyses of Xu et al. (2005, 2008). Xu et al. examined different low-cloud types based upon footprint cloud fraction criteria and found that $\tau$ is larger for stratiform clouds than for cumulus clouds; neither the low-cloud fraction nor the low-cloud lnT change much with $\omega'$, consistent with the results for SW CRE (Fig. 4). The lack of change in low-cloud fraction with $\omega'$ is consistent with the observed results for low clouds with medium and high optical depths shown in Williams et al. (2003; see their Figs. 9e,f).

Combining the results shown in Figs. 3 and 4, one can see that the change in stability ($E'$) is the dominant process in interannual variations of low-cloud environment, which is accompanied by decreases in cloud fraction and lnT as well as a decrease in solar reflection by clouds. This process is weakly modulated by changes in large-scale subsidence that is also influenced by changes in stability. A further increase in SST resulting from decreases in solar reflection of clouds via a reduction in stability likely enhances the transition in low-cloud regimes from strato-cumulus to cumulus, which further decreases solar reflection by clouds and increases SST—thus causing a positive cloud feedback.

b. Changes of properties with SST for all regions combined

In Oreopoulos and Davies (1993), they tested the hypothesis that albedo and SST remain anticorrelated under interannual variation by using the albedo and SST anomaly data. They only selected those values with magnitudes of albedo anomalies greater than 0.05 (0.025 in a second region) and SST anomalies greater than 0.5 K in magnitude. Using this SST criterion, about 75\% of the samples used in this study would have to be discarded. In general, small values of $E'$ correspond to small values of $T'_w$ with the opposite sign, as shown in Fig. 5, and the results shown in Figs. 3 and 4 indicate that there is no reason to

![Fig. 4. As in Fig. 3, but for (top) low-cloud fraction anomaly; and (bottom) low-cloud lnT.](image-url)
et al. 1992). The slopes and
2
matological applications (e.g., Peterson et al. 1992; Gray
2006) and has been used in other meteorological and cli-
least squares technique (Seneta 1983; Mielke 1991; Wilks
used here is less sensitive to outliers than the conventional
obtaining the slope (e.g., Seneta 1983). The LAD analysis
the confidence intervals were obtained by Mauger and
Norris (2010). The results from a least squares analysis have
the climate signals do not exist in the samples
associated with small $E'$ or $T_s$. Therefore, all samples (see
Table 1 for the numbers of samples in all six regions) will
be used in the linear regression analysis shown below.
The change of a cloud property anomaly $C'$ with $T_s$ can
be estimated by performing a least absolute deviations
(LAD) regression analysis between $C'$ and $T_s$, thereby
obtaining the slope (e.g., Seneta 1983). The LAD analysis
used here is less sensitive to outliers than the conventional
least squares technique (Seneta 1983; Mielke 1991; Wilks
2006) and has been used in other meteorological and cli-
matological applications (e.g., Peterson et al. 1992; Gray
et al. 1992). The slopes and 2σ uncertainties obtained from
the LAD analysis are shown in Table 2. The uncertainties
are calculated using a bootstrap approach (Efron and
Tibshirani 1993), resampling among the points 5000 times
using the effective sample size $N_{\text{eff}}$, which is similar to how
the confidence intervals were obtained by Mauger and
Norris (2010). The results from a least squares analysis have
the same sign as those shown in Table 2, but the least
squares slopes tend to be approximately 5%–30% larger in
magnitude than those from the LAD analysis. The slopes
indicate that a substantial increase in absorbed SW and net
CRE (4.03 and 3.86 W m$^{-2}$ K$^{-1}$, respectively) occurs as
SST increases. This increase in the amount of absorbed
radiation is associated with the values of SW and net CRE
becoming less negative and is due to decreases in both
low-cloud fraction (−3.44% K$^{-1}$ for CERES–MODIS,
−1.87% K$^{-1}$ for GEO) and low-cloud ln$\tau$ (−0.100 K$^{-1}$ for
CERES–MODIS, −0.085 K$^{-1}$ for GEO). The change in
LW CRE with SST is slightly negative (−0.14 W m$^{-2}$ K$^{-1}$),
which indicates that the decrease in low-cloud fraction and
optical depth allows more emission of near-surface
air to space. A slight increase in cloud cover for middle
and high clouds with $p_c < 680$ hPa (0.21% K$^{-1}$
CERES–MODIS, 0.19% K$^{-1}$ GEO) tends to mitigate
the decrease in LW CRE. However, the impacts of
middle and high clouds in the low-cloud regions are still
negligible.
There have been a number of studies analyzing the
change rates of cloud and radiative properties with SST or
SST anomalies for low clouds or subsiding dynamical
regimes (e.g., Tselioudis et al. 1992; Oreopoulos and
Oreopoulos and Davies (1993) obtained a decrease of
3%–6% K$^{-1}$ for low-cloud fraction using the ISCCP
data, which agree with our estimates for the two in-
dividual regions examined in their study. Their estimated
SW absorption increase for low-cloud regions
was 6.1 W m$^{-2}$ K$^{-1}$, which is somewhat higher than
the 4.03 W m$^{-2}$ K$^{-1}$ from our study. Bony et al. (1997)
composited SW and LW CRE, cloud fraction, cloud
optical depth with $\omega_{500}$, and SST. In the low-SST range
(291–298 K), the slopes corresponding to the subsiding
dynamic regimes (stratified $\omega_{500} > 0$ intervals) were
approximately −4% K$^{-1}$, −0.08 K$^{-1}$, 5 W m$^{-2}$ K$^{-1}$,
and −0.8 W m$^{-2}$ K$^{-1}$ for cloud fraction, ln$\tau$, and SW
and LW CRE, respectively (estimated from Fig. 7 in
Bony et al.), which agree well with the slopes obtained
from the present study (Table 2). It is cautioned that
Bony et al.’s results were not based upon anomaly fields or
interannual variabilities. Tselioudis et al. (1992) ob-
tained negative slopes of ln$\tau$ of various magnitudes
with respect to cloud-top temperature for oceanic
low clouds using 1 yr of ISCCP data. Norris and Leovy (1994)
also obtained negative correlations between seasonal
anomalies of low-cloud fraction and SST from 30 yr of
surface observations that were statistically significant in
the subtropical low-cloud regions. Finally, the trend of
cloud fraction obtained from spectrally resolved visible
satellite observations in Wagner et al. (2008) also agrees
with the present study, although their local magnitudes
are larger. The remarkable agreements with previous

![Fig. 5. Relationship between EIS and SST anomalies. The line represents a least absolute deviations fit of $E' = -0.66T_s + 0.01$ K based on all data; the circles represent the mean EIS anomaly for each 0.25-K SST anomaly bin.](image)

### Table 2. Slopes and correlations of physical property anomalies vs SST anomaly.

<table>
<thead>
<tr>
<th>Physical property anomaly</th>
<th>Slope</th>
</tr>
</thead>
<tbody>
<tr>
<td>SW CRE</td>
<td>4.03 ± 0.56 W m$^{-2}$ K$^{-1}$</td>
</tr>
<tr>
<td>LW CRE</td>
<td>−0.14 ± 0.14 W m$^{-2}$ K$^{-1}$</td>
</tr>
<tr>
<td>Net CRE</td>
<td>3.86 ± 0.36 W m$^{-2}$ K$^{-1}$</td>
</tr>
<tr>
<td>Low-cloud fraction</td>
<td>−3.44 ± 0.44% K$^{-1}$</td>
</tr>
<tr>
<td>(CERES–MODIS)</td>
<td></td>
</tr>
<tr>
<td>Low-cloud ln$\tau$</td>
<td>−0.100 ± 0.004 K$^{-1}$</td>
</tr>
<tr>
<td>(CERES–MODIS)</td>
<td></td>
</tr>
<tr>
<td>Low-cloud fraction (GEO)</td>
<td>−1.87 ± 0.36% K$^{-1}$</td>
</tr>
<tr>
<td>Low-cloud ln$\tau$ (GEO)</td>
<td>−0.085 ± 0.004 K$^{-1}$</td>
</tr>
</tbody>
</table>
studies give added confidence to the linear regression analysis adopted in this study using anomaly fields.

c. Separation into components

In the following section, we attempt to split the total change in a cloud property anomaly $C'$ with $T'_s$ into thermodynamic, dynamic, and residual terms. There has not been an approach identical to that proposed below. However, this approach has some similarities to the analysis method presented in Norris and Iacobellis (2005). They also used a two-variable ($\omega_{500}$ and SST advection) composite technique. ISCCP and ERBE data in each ($\omega_{500}$, SST advection) bi-interval were further classified into terciles of SST anomalies and terciles of LTS anomalies. Then, the mean of ISCCP and ERBE parameters in each bi-interval is calculated for each tercile. The middle terciles of SST and LTS anomalies are treated as the normal condition. Then, they took the differences of the mean warm (upper-SST tercile) and cool (lower-SST tercile) cloud/radiative properties associated with the middle LTS tercile for each $\omega_{500}$-SST advection bin. The cloud/radiative property differences are then averaged by weighting by the frequency of each bin and dividing by the average SST difference between the upper and lower terciles to obtain the change rates of cloud/radiative properties. Similar change rates are calculated by taking the differences between the upper- and lower-LTS terciles corresponding to the middle-SST tercile. In the approach proposed below, we will use the partial derivative chain rule and replace the three terciles by summation over multiple bins.

The first, second, and third terms on the lhs of Eq. (2) represent thermodynamic, dynamic, and residual change rates, respectively. Here, the partial derivatives of the form $(\partial A/\partial B)_D$ are estimated from the correlation between $A$ and $B$ within a bin interval of $D$. The frequency within this bin interval is denoted by $P_D$, and $R$ is the residual, obtained by subtracting the thermodynamic and dynamic terms from the term on the lhs of (2)—the total change rate:

$$ \frac{\partial C'}{\partial T'_s} = \sum_\omega \left( \frac{\partial C'}{\partial E'_s} \right)_\omega \left( \frac{\partial E'_s}{\partial T'_s} \right)_\omega P'_\omega \frac{E'_s}{\omega'} \sum_E \left( \frac{\partial C'}{\partial \omega'} \right)_E \left( \frac{\partial \omega'}{\partial T'_s} \right)_E P'_E + R \quad (2) $$

The correlations involving $E'$ are calculated within fixed intervals of $\omega'$ to hold the dynamics as fixed as possible while estimating the changes in $C'$ due to thermodynamics. Similarly, the correlations involving $\omega'$ are calculated within fixed intervals of $E'$ to hold the thermodynamics as fixed as possible. The residual term is interpreted here as the local estimate of cloud feedback if both $\omega'$ and $E'$ are held constant. The residual term includes additional physical processes apart from the thermodynamic and dynamic terms in (2) that are correlated to $T'_s$. These could include processes such as advection of air across SST gradients (Norris and Iacobellis 2005) or temperature advection in different layers of the atmosphere (Gordon and Norris 2010), although these factors may not be as important for low clouds near the tropics as in the midlatitudes. As Bony et al. (2004) note, short-term variations in surface temperature are larger than those in the free troposphere. This causes $E'$ to be largely controlled by $T'_s$, as was shown in Fig. 5. However, as the global climate changes over the long term, surface and free-tropospheric temperatures are likely to vary with one another, leading to smaller variations in EIS. As Wood and Bretherton (2006) note, if future temperature changes in the lower troposphere follow a moist adiabat, changes in EIS will be near zero. If the lower-troposphere temperature changes follow this trend, then the feedback due to long-term climate change will be more similar to $R$ than the estimates given in Table 2.

The thermodynamic, dynamic, and residual contributions to changes in the anomalies of SW CRE, LW CRE, and net CRE with $T'_s$ are shown in Fig. 6. The uncertainties shown represent $2\sigma$ uncertainties calculated using bootstrap resampling with $N_{eff}$ similar to section 3b. The dynamic component of the SW CRE and net CRE trends is generally negligibly small (note that this is also the case if the vertical velocity at 500 hPa is used instead). For example, the dynamic component of net CRE is 0.04 ± 0.03 W m$^{-2}$ K$^{-1}$, compared to the overall change of 3.86 ± 0.36 W m$^{-2}$ K$^{-1}$. This is consistent with the results of Bony et al. (1997), Stowasser and Hamilton (2006), and Su et al. (2010a), who found that the change in observed SW CRE with $\omega_{500}$ is fairly weak for mean subsidence regimes. The thermodynamic trends in SW CRE and net CRE are of the same sign and over half (56% and 61%, respectively) the magnitude as the overall trend, meaning that the majority of the observed decrease in the magnitudes of SW CRE and net CRE with SST can be attributed to decreases in EIS because EIS has a strong negative correlation with SST. This is not the case for LW CRE anomaly, which does not change much with $T'_s$ overall. The residual trends in SW CRE and net CRE are large (44% and 38% of the total, respectively) compared to the respective dynamic portions of the overall trends and are also of the same sign as the overall trend. Therefore, a substantial portion of the overall positive trends in SW CRE and net CRE (1.78 ± 0.50 and 1.48 ± 0.32 W m$^{-2}$ K$^{-1}$, respectively) cannot be attributed to changes in either dynamics or thermodynamics and are associated with a positive cloud feedback.

In the previous section, it was shown that the overall changes in SW and net CRE have the opposite sign to
those of cloud fraction and \( \ln r \). Does this hold true for each of the components, in particular, the residual change? In Fig. 7, the thermodynamic, dynamic, and residual contributions to changes in the cloud fraction and \( \ln r \) of low clouds are shown for both GEO and CERES–MODIS data. As discussed in section 2, the former is able to capture the diurnal variation of low clouds better than the latter but may be less accurate. Most of the decrease in the CERES–MODIS cloud fraction (and almost all of the decrease in the GEO cloud fraction) is due to the thermodynamic term. So, according to the residual change in the GEO data, the cloud fraction feedback is very weak \((0.22 \pm 0.33\% \text{ K}^{-1})\). The residual change in the CERES–MODIS data shows that cloud fraction decreases more strongly with SST \((-0.81 \pm 0.40\% \text{ K}^{-1})\) but much weaker than results that have used SST alone to estimate low-cloud feedback (e.g., Oreopoulos and Davies 1993; Bony et al. 1997). The small residual cloud fraction feedbacks resulting from the large thermodynamic contributions to the overall change—nearly all of the overall changes in GEO and 74% in CERES–MODIS—are consistent with the results of Wood and Bretherton (2006), who found that decreases in EIS are strongly associated with decreases in low-cloud fraction. We suspect that the true low-cloud fraction feedback strength lies between \(-0.81\) and \(-0.22\% \text{ K}^{-1}\). As noted earlier, the optical depth of the low clouds also decreases with SST. Over half of this decrease is due to the thermodynamic term for both the CERES–MODIS and GEO data. The residual terms \((-0.046 \pm 0.004\) and \(-0.035 \pm 0.004 \text{ K}^{-1}\) for the CERES–MODIS and GEO data, respectively) represent a larger percentage of the overall trend for low-cloud \( \ln r \) than for low-cloud fraction, particularly in the GEO data. The low-cloud optical depth feedback is positive, agreeing with the observational results of Tselioudis et al. (1992) and Bony et al. (1997), except for smaller magnitudes (reduced by the thermodynamic change). Since the GEO cloud fraction feedback is nearly zero or negative, the positive residual radiative feedback shown in Fig. 6 must be primarily due to this reduction in cloud optical depth. As is the case for SW and net CRE, the contributions of dynamic change to low-cloud fraction and \( \ln r \) are negligible.

d. Regional trends

The changes in cloud physical and radiative properties among marine low-cloud regions are not uniform across different low-cloud regions, as reported by Oreopoulos and Davies (1993). The overall and thermodynamic, dynamic, and residual changes of net CRE with SST are shown for the six regions in Fig. 8. Note that the 2\( \sigma \) uncertainties in the NE Pacific, NE Atlantic, and SE Indian regions tend to be larger because of smaller effective sample sizes for the regions. Four of the regions

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**Fig. 6.** Changes in (a) SW CRE anomaly, (b) LW CRE anomaly, and (c) net CRE anomaly per degree of SST anomaly for the different categories defined in Eq. (2). Error bars indicate 2\( \sigma \) uncertainties.
(SE Atlantic, NE Pacific, NE Atlantic, and SE Pacific) have changes in net CRE that are at least 5 W m\(^{-2}\) K\(^{-1}\), which is greater than the 3.86 W m\(^{-2}\) K\(^{-1}\) slope for all low-cloud regions combined (Table 2). However, the SC Pacific region has a much smaller change in net CRE with SST (1.0 ± 0.3 W m\(^{-2}\) K\(^{-1}\)) than the other five regions, owing perhaps to the dominant cumulus regimes there. The dynamic component of the change in net CRE is less than 0.4 W m\(^{-2}\) K\(^{-1}\) in magnitude for all six regions, indicating that changes in \(\sigma_{700}\) have relatively little impact on net CRE, as was the case for all low-cloud regions combined. The thermodynamic component is associated with an increase in net CRE for all regions, but the magnitude of this increase varies greatly between regions from 1.5 ± 1.3 W m\(^{-2}\) K\(^{-1}\) in the SE Indian to 5.3 ± 2.2 W m\(^{-2}\) K\(^{-1}\) for the NE Pacific. The thermodynamic component represents most of the overall increase in net CRE for the NE Pacific (72%), SE Pacific (69%), and SC Pacific (121%) regions, while the residual represents a large portion of the overall increase for the SE Atlantic (65%), NE Atlantic (47%), and SE Indian (63%) regions. The residual is positive for all of these regions except for the SC Pacific (−0.2 ± 0.3 W m\(^{-2}\) K\(^{-1}\)), indicating that the positive radiative feedback seen for the overall low-cloud data in Fig. 6 is also present in most of the individual regions.

There is a decrease in cloud fraction with SST for all six regions, shown in Fig. 9 for both GEO and CERES–MODIS data. The magnitudes of the decrease for GEO cloud fraction are consistently smaller than those of CERES–MODIS for the overall and thermodynamic changes (Figs. 9a,b), although the thermodynamic changes are more similar between the two data owing perhaps to the strong dependence of cloud fraction on stability. The lack of the diurnal cycle in the CERES–MODIS data and inaccuracies in the GEO data may contribute to the large differences in the overall and residual changes between the two cloud fraction data (Figs. 9a,d).

There is apparent consistency between the increase in net CRE and decreases in cloud fraction (and cloud optical depth; Fig. 10) for the regions, and the regions with the largest and smallest increases in net CRE (NE Atlantic and SC Pacific, respectively) also have the largest and smallest decreases in cloud fraction (−10.6 ± 1.5% K\(^{-1}\) and −1.1 ± 0.5% K\(^{-1}\) for CERES–MODIS). The overall decrease in cloud fraction with SST for the SE Atlantic region is nearly as large as that for the SE Pacific region (Fig. 9a), which agrees with Oreopoulos and Davies (1993). The decrease in low-cloud fraction with SST in the NE Pacific is consistent with the longer-term trends seen in Clement et al. (2009). The thermodynamic component of the change in low-cloud fraction is...
negative for all six regions, which is consistent with the observed dependence of cloud fraction on EIS across many different regions (Wood and Bretherton 2006) since EIS is negatively correlated with SST for the regions and time scales in this study. The residual changes in low-cloud fraction are negative (implying a positive cloud fraction feedback except for the SE and SC Pacific regions) and roughly half the magnitude of the overall change in low-cloud fraction for the SE Atlantic, SE Indian, NE Pacific, and NE Atlantic regions. The residual change in low-cloud fraction is actually positive for the SE Pacific and SC Pacific regions, which is also the case for the GEO cloud fraction in the SE Indian region. A closer examination, however, shows a few exceptions to the consistency between the increase in net CRE and decreases in cloud fraction (and cloud optical depth). The NE Pacific has a decrease in cloud fraction (−2.8 ± 1.0% K⁻¹) that is smaller than the slope for all six regions combined (Table 2), even though the NE Pacific has a strong increase in net CRE with SST. This implies that the negative cloud optical depth feedback is more important owing to the closer correspondence between the net CRE increase and lnτ decrease (Fig. 10) in all regions except for the thermodynamic change in the NE Atlantic region. This region has the largest decreases in low-cloud fraction and low-cloud lnτ in the overall, thermodynamic, and residual changes among all regions and at least twice as large as those in the SE Atlantic region. Yet the increases in the net CRE are close to those for the SE Atlantic region. Uncertainties in the retrieval of cloud optical depth and cloud fraction caused by transport of large aerosol loading from the African continent (Eitzen et al. 2008; Loeb et al. 2003) may contribute to this inconsistency. Similarly, both the GEO cloud fraction and lnτ show positive trends in the residual change for the SE Indian region, but the net CRE trend is also positive. This result also illustrates that uncertainties in the GEO retrieval are more significant than in the CERES–MODIS retrieval in both quantities. Nevertheless, both retrievals suffer from the cloud identification issue in heavily polluted regions (Loeb et al. 2003).

4. Summary and conclusions

This study has used CERES radiative flux and cloud property data and matched ECMWF reanalysis data to examine how cloud and radiative property anomalies vary with SST anomaly in low-cloud regions. The anomalies
were composited into the changes in thermodynamic state (represented by \( E^\prime \)) and dynamic state (represented by \( \omega^\prime \)). To eliminate (as much as is possible with monthly mean data) the effects of high clouds, only those grid cells and months that had a 5-yr climatological average middle and high \((p_c < 680 \text{ hPa})\) cloud fraction of less than 10% were considered. Mean SW, LW, and net CRE, low-cloud fraction and low-cloud ln\( \tau \) anomalies composited over \( E^\prime - \omega^\prime \) bins tended to change much more with \( E^\prime \) than \( \omega^\prime \). This is broadly consistent with the findings of Bony et al. (1997), Stowasser and Hamilton (2006), and Su et al. (2010a), who found that the change in observed SW CRE with \( \omega_{500} \) is fairly weak for mean subsidence regimes. The composite results indicate that as anomalies in stability increase, both the low-cloud fraction and low-cloud ln\( \tau \) anomalies increase but the net CRE (due mostly to SW CRE) become more negative—more reflection by clouds. If the increase in stability were due solely to a decrease in SST, this result would imply a positive cloud feedback.

The trends of these cloud and radiative properties with SST anomaly were examined for low-cloud areas—over six subtropical oceanic regions—as a whole, and also for six individual low-cloud regions. The low-cloud fraction and low-cloud ln\( \tau \) decreased for every low-cloud region, with the NE Atlantic experiencing the largest decreases per degree of SST and the SC Pacific the smallest. Similarly, every low-cloud region experienced positive changes in net CRE (i.e., less net outgoing radiation at the top of the atmosphere) with SST. For all regions combined, the decrease in cloud fraction is estimated to be \(-3.44 \pm 0.44\% \text{ K}^{-1}\) for CERES–MODIS data (without a full diurnal cycle) and \(-1.87 \pm 0.36\% \text{ K}^{-1}\) for GEO (with a full diurnal cycle) data and the decrease in low-cloud ln\( \tau \) is much closer between the two data \((-0.100 \pm 0.004\) and \(-0.085 \pm 0.004\) K\(^{-1}\)). There are substantial increases in absorbed SW and net CRE of \(4.03 \pm 0.56\) and \(3.86 \pm 0.36\) W m\(^{-2}\) K\(^{-1}\), respectively. These change rates agree broadly with previous studies (e.g., Tselioudis et al. 1992; Oreopoulos and Davies 1993; Norris and Leovy 1994; Bony et al. 1997; Wagner et al. 2008; Clement et al. 2009) despite different datasets and scales of temporal variability. When these trends are extrapolated into the entire globe, the magnitudes would be much smaller and comparable to those of GCMs (e.g., Zhu et al. 2007a,b).

The overall trends in cloud and radiative properties with SST anomaly mentioned above are decomposed into dynamic, thermodynamic, and residual components. Changes in the amount of subsidence at 700 hPa have relatively little effect on cloud and radiative properties—as SST increases, the lower-tropospheric stability decreases—while this accounts for a large portion of the measured
decreases in cloud fraction and cloud optical depth. The residual positive change in net cloud radiative effect ($1.48 \pm 0.32 \text{ W m}^{-2} \text{ K}^{-1}$), the small change in low-cloud amount ($-0.81$ to $0.22\% \text{ K}^{-1}$), and the decrease in the logarithm of optical depth ($-0.035$ to $-0.046 \text{ K}^{-1}$) with SST are smaller and can be interpreted as a positive cloud feedback, with cloud optical depth feedback being the dominant contributor. The residual trends vary among regions but are primarily positive feedbacks (lower-cloud fraction and optical depth, positive change in net CRE) for most of the regions examined here, with the largest positive feedbacks ($-4 \text{ W m}^{-2} \text{ K}^{-1}$) for the SE and NE Atlantic regions and a slightly negative feedback ($0.2 \pm 0.3 \text{ W m}^{-2} \text{ K}^{-1}$) in the SC Pacific region. Uncertainties in the retrieval of cloud optical depth and/or cloud fraction caused by the transport of large aerosol loading from the African continent may contribute to the large magnitude in the two Atlantic regions. Exceptions to the positive feedback are the low-cloud fraction in the SE Pacific and SC Pacific regions, which have slight positive residual changes with SST. The residual ln$\tau$ decreases are large enough in these two regions that the net CRE feedback is positive for the SE Pacific region and near zero for the SC Pacific region. The SE Indian region experienced positive trends in the residual GEO cloud fraction and ln$\tau$, but the net CRE trend is also positive. This contradiction illustrates that uncertainties in the GEO retrieval are more significant than those in CERES–MODIS in both cloud fraction and ln$\tau$. The uncertainties in the trends as listed in Table 2 should not be overlooked and may underestimate the true uncertainty.

Because of the relatively short length of the dataset used here, there remains uncertainty in the low-cloud feedback that will be associated with global climate change. As noted by Zhu et al. (2007b), there are many competing processes involved in the maintenance of boundary layer clouds, even when simplifying assumptions are made. A negative low-cloud fraction feedback in the GCMs studied by Zhu et al. (2007b) was related to an increase in the stability above the inversion when the level of CO$_2$ was doubled. Similarly, if EIS increases in low-cloud areas as the climate warms, the positive feedbacks shown here could be reduced or even reversed in sign. The strength of subsidence (provided that the regime remains subsiding) was not a major factor in the characteristics of low clouds in this study or the doubled CO$_2$ experiments studied by Zhu et al. (2007b). This indicates that small changes in the strength of the subsidence may not play an important role in how low clouds change in the future. Reproduction of the trends seen in this study for the same time period should represent a necessary condition for GCMs hoping to simulate global climate change. Contradictory signs in
low-cloud feedbacks among the GCMs should be examined in relation to the physical processes represented in the models’ parameterizations. This is because there is a transition between stratocumulus to cumulus regimes as SST increases, as revealed from the data analysis presented in this study. For example, the decoupling of the boundary layer is one of the important processes in the transition of cloud regimes. When the transition is absent, as discussed in Xu et al. (2010) with a cloud-resolving model and in Zhang and Bretherton (2008) with a single-column model, the low-cloud feedback tends to be negative. Understanding the different processes with the help of these models will hopefully narrow down the uncertainties in the GCMs if a wide range of low-cloud regimes and their transitions are simulated in these process models.

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