Low-Level Cloud Variability over the Equatorial Cold Tongue in Observations and Models

DAVID K. MANSBACH AND JOEL R. NORRIS

Scripps Institution of Oceanography, University of California, San Diego, La Jolla, California

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ABSTRACT

Examination of cloud and meteorological observations from satellite, surface, and reanalysis datasets indicates that monthly anomalies in low-level cloud amount and near-surface temperature advection are strongly negatively correlated on the southern side of the equatorial Pacific cold tongue. This inverse correlation occurs independently of relationships between cloud amount and sea surface temperature (SST) or lower tropospheric static stability (LTS), and the combination of advection plus SST or LTS explains significantly more interannual cloud variability in a multilinear regression than does SST or LTS alone. Warm anomalous advection occurs when the equatorial cold tongue is well defined and the southeastern Pacific trade winds bring relatively warm air over colder water. Ship meteorological reports and soundings show that the atmospheric surface layer becomes stratified under these conditions, thus inhibiting the upward mixing of moisture needed to sustain cloudiness against subsidence and entrainment drying. Cold anomalous advection primarily occurs when the equatorial cold tongue is weak or absent and the air–sea temperature difference is substantially negative. These conditions favor a more convective atmospheric boundary layer, greater cloud amount, and less frequent occurrence of clear sky.

Examination of output from global climate models developed by the Geophysical Fluid Dynamics Laboratory (GFDL) and the National Center for Atmospheric Research (NCAR) indicates that both models generally fail to simulate the cloud–advection relationships observed on the northern and southern sides of the equatorial cold tongue. Although the GFDL atmosphere model does reproduce the expected signs of cloud-advection correlations when forced with prescribed historical SST variations, it does not consistently do so when coupled to an ocean model. The NCAR model has difficulty reproducing the observed correlations in both atmosphere-only and coupled versions. This suggests that boundary layer cloud parameterizations could be improved through better representation of the effects of advection over varying SST.

1. Introduction

Low-level clouds combine a small greenhouse effect with a generally high albedo and thus contribute significantly to the overall net cooling role of clouds in earth’s climate (Ramanathan et al. 1989). Currently, lack of both resolution and appropriate physical parameterizations prohibit reliable large-scale numerical prediction of cloud–climate feedbacks (e.g., Stephens 2005; Bony and Dufresne 2005). A good strategy for improving our understanding of climate mechanisms and their numerical simulation is to carry out focused studies that elucidate specific ocean–atmosphere–cloud relationships and that can inform and constrain model results. By examining the interannual variability of low-level clouds in the eastern equatorial Pacific, an area of high atmospheric and oceanic variability located on the edge of a persistent stratiform cloud deck, we aim to uncover sometimes-subtle details of marine low-level cloud processes.

Climatologically, low-level stratiform clouds are found in subtropical subsidence regions over the relatively cool eastern side of oceans, and they are most prevalent in seasons of high atmospheric and oceanic variability located on the edge of a persistent stratiform cloud deck, we aim to uncover sometimes-subtle details of marine low-level cloud processes.

Climatologically, low-level stratiform clouds are found in subtropical subsidence regions over the relatively cool eastern side of oceans, and they are most prevalent in seasons of high lower-tropospheric static stability (LTS; Klein and Hartmann 1993). The relationship between low-level cloud amount and LTS, or sea surface temperature (SST), found in the seasonal cycle is also apparent at other time scales. Interannual anomalies in cloud amount and SST are negatively correlated in eastern subtropical oceans (Norris and Leovy 1994), with the maximum effect on clouds occurring
24–30 h downwind from SST anomalies (Klein et al. 1995). There is also evidence of low-level cloud variations preceding SST anomalies in daily data, though the extent, nature, and exact mechanisms involved in two-way low-level cloud–SST relationships remain to be fully analyzed (Klein 1997; Xu et al. 2005). Low-level stratiform clouds display appreciable variability on synoptic, seasonal, and interannual scales, with larger regions generally showing more variability over longer time scales (Rozendaal and Rossow 2003), and year-to-year changes in cloud amount can be substantial relative to the climatological mean.

While most low-level stratiform clouds exist in areas of the subtropics where prevailing winds advect the atmospheric boundary layer equatorward over slowly increasing SST, one exception is the region of the eastern Pacific equatorial cold tongue. Here northward trade winds flow over climatologically decreasing SST between 5° and 1°S and then over rapidly increasing SST as they approach the intertropical convergence zone near 8°N. Previous studies have demonstrated that the boundary layer is sensitive to changes in underlying SST (e.g., Yin and Albrecht 2000) especially on the northern side of the equatorial cold tongue, where advection over the sharp SST gradient generates large latent and sensible heat fluxes that favor development of extensive stratocumulus (Deser et al. 1993). Much less research attention has been devoted to the southern side of the cold tongue, where advection of relatively warm air over cold SST can produce shallow and stably stratified layers near the surface (e.g., Paluch et al. 1999). This is one of the very few areas of the ocean interior where cloudless boundary layers occur at non-negligible frequency (Norris 1999b).

The southern side of the equatorial cold tongue provides a unique setting to study how advection over decreasing SST affects low-level cloudiness. The analysis is simplified since the trade winds are steady, unlike the case for extratropical latitudes. Moreover, this area lies on the northern edge of the extensive southeastern Pacific stratiform cloud region and experiences large year-to-year changes in low-level cloud amount. The present study describes and explains the driving forces behind these interannual cloud variations using satellite cloud data, ship-based synoptic reports, and soundings from two different years in the eastern equatorial Pacific region. To see whether the important elements of our findings are reproduced in climate simulations, we also examine output from two atmospheric general circulation models (AGCMs) run over prescribed SST and output from three coupled atmosphere–ocean general circulation models (CGCMs).

2. Observational data sources

The primary source of cloud information in this study is monthly mean daytime low-level cloud amount from the International Satellite Cloud Climatology Project (ISCCP) D2 visible-radiance-adjusted cloud amount dataset (Rossow and Schiffer 1999). These data are available on a 2.5° × 2.5° grid during July 1983–September 2001. In ISCCP, low-level refers to clouds whose tops are at a pressure greater than 680 hPa, and we look at combined low-level cloud coverage of all optical thicknesses. Since a satellite cannot see low-level clouds when upper-level clouds are obscuring its view, we adjust ISCCP data by assuming that low-level clouds are randomly overlapped with upper-level (middle plus high) clouds. Taking satellite-observed upper-level cloud cover, \( U \), and satellite-observed low-level cloud cover, \( L \), into account, we compute \( L' \), the corrected low-level cloud amount, from \( L' = L/(1 – U) \), after Rozendaal et al. (1995). For convenience, the adjustment for cloud overlap is applied to monthly mean data instead of adjusting instantaneous values for cloud overlap prior to averaging them to monthly means. We found that values calculated from the two methods during July and October for the region 20°S–0°, 110°–85°W had a correlation of 0.997, and in the rest of our analysis we therefore use monthly mean low-level cloud amount adjusted for overlap.

The most important large-scale meteorological parameter for our study is advection by 1000-hPa winds over varying SST: \(- \nabla_{1000} \cdot \mathbf{v} \cdot \text{SST}\). We refer to this as SST advection, and obtain monthly mean wind and SST data from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis (Kalnay et al. 1996). The SST gradient used is the upstream gradient; that is, composed of the finite differences centered half a grid box south and east of the corresponding point for which the advection value is indicated, in order to better approximate the temperature advection in the southeastern trade winds of the southern Tropics.

Since ISCCP cannot provide morphological cloud type information we also examine data from ship-based surface observers in the Extended Edited Cloud Report Archive (EECRA; Hahn and Warren 1999) from 1983 to 1997. Observations in the EECRA include sky coverage by all clouds and by low-level clouds, cloud type, SST, air temperature and pressure, dewpoint depression, and wind speed.

As a complement to the EECRA synoptic reports, we also make use of soundings by research vessels affiliated with the Eastern Pacific Investigation of Climate Processes in the Coupled Ocean–Atmosphere
System (EPIC) program. Of particular interest to us are data from transects along 95° and 110°W in November 1999 and November 2001, described in Pyatt et al. (2005), since these two years correspond to different SST advection conditions south of the equator.

After the treatment of observational data, we briefly examine whether modern general circulation models show a cloud–temperature advection response similar to that in observations. We examine both CGCM and AGCM simulations over the time period 1983–2001, which most nearly matches the ISCCP period.

3. Observational analysis and results

a. Correlation and multilinear regression analysis

Our analysis focuses on the cool-season months of June through November since that is the time of year when climatological low-level cloud amount and interannual variability are largest in the southeastern tropical Pacific, thus implying a greater potential cloud feedback on the climate system. Nonetheless, our results would be qualitatively similar if all months were included. Figure 1 shows the June–November climatological low-level cloud amount and standard deviation of interannual monthly anomalies in the ISCCP data. The heart of the dense low-level cloud region aligns with areas of cooler waters off the South American coast, with a secondary area of extensive low-level cloud amount north of the equatorial SST front.

The area of the northern equatorial cold tongue exhibits high interannual variability in cloud amount due to changes in SST and SST advection associated with El Niño–Southern Oscillation (ENSO) events and tropical instability waves (Deser et al. 1993). The region of pronounced variability east of 110°W between 5°S and the equator is a region of predominant warm SST advection. Since the sign of meridional SST gradient and climatological temperature advection reverse at or just south of 0°, the sign of SST advection south of the equator at any time is typically opposite from that north of the front.

To investigate the conditions associated with low-level clouds in the cold tongue region, we begin with the satellite dataset. From the overlap-corrected ISCCP data, we subtract 19-yr (18 for October and November) averages from the monthly observations, and using the anomalies, proceed with correlation and linear regression analysis at each 2.5°×2.5° grid box. Maps of local correlations between reanalysis meteorological parameters and low-level cloud amount serve as a starting point in attributing low-level cloud variability. For such maps, we use the effective sample size and a standard t test to compute significance of linear correlations. Our method follows that of Klein (1997), but on a month- to-month basis rather than daily.

Since SST has been shown to have a close association with low-level stratiform cloudiness (Norris and Leovy 1994), we use it as our first correlation parameter. In Fig. 2a, monthly anomalies in SST and cloud amount exhibit the expected negative correlation over most of the climatological low-cloud region, although the relationship is statistically significant mainly in the southern region of the stratiform deck and north of the equatorial SST front. Anomalies in SST are also negatively correlated with anomalies in low-level cloud optical thickness (not shown). Another parameter that is closely related to SST and cloudiness is LTS, defined as the difference in potential temperature between the 700-hPa level and 2 m above the surface (θ700 – θ2m) following Klein and Hartmann (1993). The statistically significant positive correlations between cloud amount and LTS anomalies on the northern side of the cold tongue displayed in Fig. 2b are due almost entirely to

![Figure 1](image-url)
variations in SST rather than variations in 700-hPa temperature.

The sense of SST advection is another meteorological parameter that previous studies have found to be strongly related to low-level stratiform cloud amount (e.g., Klein et al. 1995; Klein 1997; Norris and Iacobellis 2005), and Fig. 2c shows that statistically significant negative correlations exist between monthly anomalies in SST advection and low-level cloud amount over the equatorial cold tongue and southwest of the climatological maximum in cloud amount. Although not shown, SST advection is also negatively correlated with low-level cloud optical thickness in these regions. The weak positive cloud–SST advection correlations occur where interannual variability is quite weak (Fig. 1) and consequently are less likely to be influential on the climate system. Considering that both SST and SST advection are related to variations in low-level cloud amount, it is instructive to examine how they are related to each other. Figure 2d presents the spatial pattern of correlations between monthly anomalies in SST and SST advection. The most prominent feature in this plot is the band of positive correlation north of the equator and the band of negative correlation south of the equator. This is because SST varies most strongly along the near-equatorial SST front, with lesser variability of the same sign to the north and south. Since monthly mean SST advection anomalies in this area are mainly determined by SST variability along the largely steady wind streamlines, the latitude of greatest SST variability determines the latitude where the anomalous SST advection changes in sign. At the resolution of the grid used here, that latitude is on or within a degree south of the equator, the same as the location of the climatological SST front. On the northern side of the cold tongue, colder SST is associated with more negative SST advection, both of which favor increased low-level cloud amount. Contrastingly, on the southern side of the cold tongue, colder SST is associated with more positive SST advection, thus producing opposing effects on cloud amount.

The hypothesis that low-level cloudiness is significantly influenced by more than one meteorological factor may be tested by gauging how much additional variance ($R^2$ statistic) is explained by a second parameter in a multilinear regression at each grid box. To determine whether the increase in $R^2$ due to adding a second physical parameter is statistically significant, we employ bootstrap techniques that take into account the degree of statistical dependence of consecutive months in each grid point as well as the $R^2$ increase from the second parameter, described in more detail in appendix A. Figure 3 displays the amount of additional low-level cloud variance that SST advection as a second regressor accounts for when SST is the first regressor.

Although accounting for little extra variance in much
of the domain, SST advection explains significantly more cloud variance than SST alone in a region on the southern side of the equatorial cold tongue (5°S–0°, 105°–95°W), hereafter called the southern cold tongue region. In this region, the fraction of low-level cloud amount variance explained using both regressors is around 0.5, whereas with only SST it is closer to 0.1. Very similar results are obtained if LTS is used instead of SST as the first regressor, and if climatological monthly mean winds are used instead of varying monthly mean winds. Furthermore, if we use nonoverlap-adjusted low-level cloud amount (corresponding to an assumption of minimum overlap between clouds of different heights), or total ISCCP cloudiness (corresponding to a maximum overlap assumption), a very similar pattern also results. Contrastingly, other second regressor variables such as /H925−700, vertical velocity at 700 hPa, or surface divergence do not exhibit any organized pattern that explains significantly more cloud variance.

The sensitivity of low-level cloud amount on the southern side of the cold tongue to interannual variations in advection over the SST gradient is due to the fact that absolute (not anomalous) advection changes sign from year to year. Over most subtropical stratiform cloud regions, SST advection ranges only from moderately negative to strongly negative, but in the southern cold tongue region, SST advection ranges from moderately positive to weakly negative. As listed in Table 1, monthly anomalies in SST advection averaged over the latter area are inversely correlated with anomalies in local and Niño-3.4 SST. Table 1 also shows that low-level cloud amount variability in the southern cold tongue region is more closely related to variability in SST advection than it is to variability in local SST, Niño-3.4 SST, or LTS.

### b. Cold and warm SST advection composites

Compositing analysis is a useful method to illustrate changes in cloud and surface meteorological properties associated with variations in SST advection. We do this by classifying anomalous monthly SST advection values averaged over the southern cold tongue region into lower, middle, and upper terciles, and then examining the cloud and meteorological anomalies associated with each tercile.

Figure 4 presents mean SST and 1000-hPa wind distributions for lower and upper SST advection terciles, along with the differences between the two terciles.

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**Table 1.** Correlation coefficients for monthly mean interannual anomalies of several reanalysis meteorological fields and ISCCP cloud fields averaged over the southern cold tongue region, 5°S–0°, 105°–95°W. Correlations significant at the 90% level are in bold and those also significant at the 95% level are in italics.

<table>
<thead>
<tr>
<th></th>
<th>Low-level cloud amount</th>
<th>Upper-level cloud amount</th>
<th>Niño-3.4 SST</th>
<th>SST advection</th>
<th>Local SST</th>
<th>θ₀₀₀ – θ₂₀₀</th>
</tr>
</thead>
<tbody>
<tr>
<td>Low-level cloud amount</td>
<td>—</td>
<td>—</td>
<td>0.16</td>
<td><strong>-0.41</strong></td>
<td>-0.28</td>
<td>0.36</td>
</tr>
<tr>
<td>Upper-level cloud amount</td>
<td><strong>-0.50</strong></td>
<td>—</td>
<td>0.38</td>
<td>-0.26</td>
<td><strong>0.76</strong></td>
<td><strong>-0.73</strong></td>
</tr>
<tr>
<td>Niño-3.4 SST</td>
<td>0.16</td>
<td>0.38</td>
<td>—</td>
<td>-0.79</td>
<td><strong>0.74</strong></td>
<td><strong>-0.62</strong></td>
</tr>
<tr>
<td>SST advection</td>
<td><strong>-0.41</strong></td>
<td><strong>-0.26</strong></td>
<td><strong>-0.79</strong></td>
<td>—</td>
<td>-0.61</td>
<td><strong>0.49</strong></td>
</tr>
<tr>
<td>Local SST</td>
<td>-0.28</td>
<td>0.76</td>
<td>0.74</td>
<td>-0.61</td>
<td>—</td>
<td>-0.90</td>
</tr>
<tr>
<td>θ₀₀₀ – θ₂₀₀</td>
<td>0.36</td>
<td>-0.73</td>
<td>-0.62</td>
<td>0.49</td>
<td>-0.90</td>
<td>—</td>
</tr>
</tbody>
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very small magnitude of wind vector differences near
the equator (Fig. 4c) indicates that variations in SST
advection are almost exclusively driven by changes in
the SST gradient. The lower tercile, corresponding to
cold SST advection in the anomaly sense, occurs when
the equatorial cold tongue is weak and the along-wind
SST gradient is nearly flat, leading to near-zero SST
advection in the absolute sense (Fig. 4a). The upper
tercile, corresponding to warm SST advection in the
anomaly sense, occurs when the equatorial cold tongue
is strong and the along-wind SST gradient is positive on
the southern side of the cold tongue, leading to sub-
stantial warm SST advection in the absolute sense (Fig.
4b). The SST difference between terciles is reminiscent
of the SST anomaly pattern from classic El Niño events
away from the South American coast (e.g., Rasmussen
and Carpenter 1982; Deser and Wallace 1990), but dif-
fers from them in that it does not have strong equatorial
wind anomalies or sizable SST anomalies along the
coast.

Figure 5 displays meridional profiles of ISCCP and
EECRA cloud amount at low and upper levels for cold
and warm anomalous SST advection terciles. These
profiles were obtained by averaging monthly ISCCP
cloud values and individual EECRA cloud observa-
tions in 2.5° latitude increments between 105° and
95°W, and the error bars show the 95% confidence
interval for each zonal mean using bootstrap methods
described in appendix B. South of the equator, satellite
and surface data both exhibit substantially less low-
level cloud amount with warm anomalous advection
and substantially more low-level cloud amount with
cold anomalous advection. The decrease in ISCCP
low-level cloud amount with warm anomalous advection
cannot be attributed to changes in overlapping higher
clouds since ISCCP upper-level cloud amount is clima-
tologically very small in this region. North of 1.25°N,
satellite and surface data both exhibit substantially
more low-level cloud amount when warm anomalous
advection occurs in the southern cold tongue region.
The opposing changes in low-level cloud amount on
southern and northern sides of the cold tongue are due
to the reversal of the anomalous SST gradient along
surface wind streamlines after the steady southeasterly
winds cross the equatorial SST front.

One advantage of the EECRA data is that they in-
clude surface meteorological measurements and mor-
phological cloud type observations that provide a quali-
tative description of boundary layer structure (e.g.,
Norris 1998a). Figure 6 displays meridional profiles of
these variables for cold and warm anomalous SST ad-
vecton terciles. Warm anomalous advection is associ-
ated with the markedly more frequent occurrence of
clear sky and absence of low-level cloudiness on the
southern side of the equatorial cold tongue, features
that are also present in the climatological mean (Norris
1998b; Park and Leovy 2004). The decrease in low-level
cloud cover with warmer SST advection primarily re-
sults from a decrease in cumuliform cloud cover (de-

defined as cumulus alone and cumulus mixed with stra-
tocumulus). We attribute this implied weakening of
convection to increased stratification of the atmo-
spheric surface layer produced by advection of warmer
air over cooler water, as is suggested by the near-zero
air–sea temperature difference observed under these
conditions.

Figure 6 also shows that near-surface relative humid-
ity is larger and wind speed is smaller for warm SST advection in the southern cold tongue region, indicating that there is less upward mixing of moisture and downward mixing of momentum (e.g., Wallace et al. 1989). The enhanced stratiform cloud cover (defined as stratocumulus alone and stratus) that occurs with warm SST advection may be remnant cloudiness that has not yet dissipated even though it is no longer in turbulent communication with the ocean surface. Cold anomalous advection is associated with less frequent clear sky, more cumuliform cloud cover, a more negative air–sea temperature difference, and a drier and windier surface layer (Fig. 6). All of these conditions are characteristic of the typical cloud regime downwind from major stratocumulus decks.

It is worth noting that in Fig. 6, the conditions in the tercile of anomalously cold temperature advection south of the equator show more cumuliform clouds relative to the warm advection tercile all the way up to 5°N, implying that some of the additional cumulus clouds formed south of the equator persist as they are advected northward. Likewise, the deficit in stratocumulus clouds is evident from 7.5°S to 7.5°N. The net result, shown in both the EECRA and ISCCP data (Fig. 5), is that the cold anomalous advection tercile relative to warm has more low-level cloud cover when averaged over the extent of the cold tongue, 5°S–5°N, in the longitude range 105°–95°W.

The changes in surface wind speed and relative humidity, in addition to the cloud changes, are all consistent with increased boundary layer vertical mixing as in Wallace et al. (1989). An alternative hypothesis holds that anomalous SSTs lead to boundary layer pressure anomalies, and that the corresponding anomalous pressure gradients cause anomalous low-level winds (Lindzen and Nigam 1987). In the eastern tropical Pacific, Hashizume et al. (2001) and Hashizume et al. (2002) find little evidence for winds responding to such an effect. The data used in our analysis, focused as they are on boundary layer cloud variability, are not suitable for a definitive investigation of the Lindzen–Nigam mechanism. This is particularly the case for the reanalysis winds in Fig. 4 since they are likely to be influenced by the boundary layer parameterization used in the

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**Fig. 5.** (left) Low-level and (right) upper-level cloud amount from (top) EECRA and (bottom) ISCCP datasets averaged over 105°–95°W and months belonging to terciles of cold anomalous advection (solid and filled circles) and warm anomalous advection (dashed and open diamonds). Error bars indicate 95% confidence intervals for the zonal means. ISCCP low-level cloud amount and EECRA upper-level cloud amount have been adjusted for overlapping higher and lower clouds, respectively.
model. We do note that in Fig. 6 the meridional pressure gradient is steadily negative over the cold tongue region and has a slope that becomes steeper in the tercile months and latitudes with anomalously cold advection and warm SST. Mean vertical profiles of wind speed from EPIC soundings (not shown) also show that in addition to the surface wind tendencies shown in the EECRA data, wind speeds in the upper parts of the boundary layer were greater over anomalously cold temperature advection and anomalously warm SST, which is consistent with Lindzen and Nigam (1987). Beyond these points, the analysis of mean values of ship reports from 15 yr of EECRA data does not allow for precise calculations of momentum budgets and other methods would be required for more in-depth, quantitative evaluation of the relevant mechanisms.

Quantitative measurements of boundary layer structure are available from EPIC soundings made during November 1999, a month in the warm anomalous advection tercile, and during November 2001, a month on the cool side of the middle advection tercile. Average vertical profiles for these two months are obtained from all soundings between the equator and 5°S along transects at 95° and 110°W (also described in Pyatt et al. 2005). To preserve structure near the top of the boundary layer, soundings are scaled by the height of the base of the trade inversion prior to averaging, and the mean profile is then rescaled by the average inversion base height. The base of the trade inversion is defined as the lowest elevation where the subsequent 10 data points (usually about 200 m) are warmer and the temperature increases by at least 1.5 K, and those few soundings without discernable inversions are discarded. We examine water vapor mixing ratio, saturation water vapor mixing ratio (closely related to temperature), virtual potential temperature, and equivalent potential temperature (calculated according to Bolton 1980). While they are relevant to the structure of most of the boundary layer and the trade inversion, the soundings do not accurately depict surface-layer ef-
effects. This is because the mean soundings miss the bottom 20 to 40 m of the atmosphere, and, moreover, the lowest recordings in many soundings are biased because of the ship’s effect on the environment and the instrument sensors (C. Fairall 2006, personal communication; see also Wang et al. 2002).

Figure 7 shows that strong warm SST advection during November 1999 (1.22 K day$^{-1}$) was associated with a boundary layer that was shallower, cooler, and drier than that observed for weak warm SST advection during November 2001 (0.50 K day$^{-1}$). The differences in boundary layer height in Fig. 7 and the cloud type in Fig. 6 are in agreement with previous studies noting that subtropical cumulus and cumulus-with-stratocumulus occur in deeper boundary layers than those for stratocumulus alone (Albrecht et al. 1995; Norris 1998a). Pyatt et al. (2005) also report that south of the equator, surface relative humidity was greater, surface wind speed was weaker, and cloud fraction was very much smaller during November 1999 than during November 2001. These observations demonstrate that stratification of the near-surface layer caused by warm air flowing over colder water inhibits the upward mixing of moisture needed to sustain cloudiness. The lower inversion height in 1999 is consistent with subsidence pushing the trade inversion further down as reduced buoyancy at the surface and cloud-top radiative cooling act to inhibit entrainment.

Further support for this stratification phenomenon is provided by the study of Paluch et al. (1999), which documented the occurrence of shallow stable surface layers beneath warmer layers over the southern side of the cold tongue and areas of coastal upwelling during September 1996, another month in the warm anomalous advection tercile. In contrast, measurements taken north of the equator in the same field campaign show a deeper dry adiabatic layer starting at the surface, lying beneath a deep moist adiabatic layer with cloud all the way up to the inversion (Paluch et al. 1999). The large-eddy simulation in de Szoeke and Bretherton (2004), which is forced by boundary conditions characteristic of the equatorial Pacific along 95°W, also clearly shows a stable surface layer up to the approximate latitude where the SST front is found.

4. Comparison with general circulation model output

Previous investigations have demonstrated that atmosphere and coupled ocean–atmosphere general circulation models produce unrealistic simulations of low-level cloudiness over the eastern tropical Pacific, which contribute to large biases in the climate state (e.g., Ma et al. 1996; Yu and Mechoso 1999). This motivates comparison of the observational results of the present study to output from the two leading U.S. GCMs in order to determine whether they correctly reproduce the observed relationship between cloud amount and SST advection.

The first model is the Geophysical Fluid Dynamics Laboratory (GFDL) atmosphere-only AM2 (described in GFDL Global Atmospheric Model Development
Team 2004) and two versions of the coupled ocean–atmosphere CM2 (described in Delworth et al. 2006). The GFDL AM2 and CM2 use a prognostic cloud fraction scheme based on Tiedtke (1993). The planetary boundary layer is modeled with a nonlocal scheme based on Lock et al. (2000). This scheme is based on results of various large-eddy simulations and classifies the boundary layer as one of six types, each with a different profile of static stability, decoupling, moisture, and condensate. It determines cloud vertical extent and boundary layer depth by calculating the height of neutral buoyancy for rising and sinking parcels using conserved moist variables (Lock et al. 2000; GFDL Global Atmospheric Model Development Team 2004). Because the AM2 was found to be unreliable in predicting cloud liquid water, the term in Lock et al. (2000) accounting for buoyancy change due to entrainment drying is omitted and the entrainment parameterization adjusted (GFDL Global Atmospheric Model Development Team 2004). The two versions of the CM2, designated CM2.0 and CM2.1, primarily differ in their dynamical cores and particular aspects of cloud tunings. We do not expect that these differences will substantially influence low-level cloud variability over the equatorial cold tongue aside from producing dissimilar basic climate states.

The second model set is the NCAR atmosphere-only CAM3 (Boville et al. 2006; Collins et al. 2006b) and coupled ocean–atmosphere CCSM3 (described in Collins et al. 2006a). The NCAR CAM3 and CCSM3 use a diagnostic cloud fraction scheme based on a relative humidity threshold and an empirical relationship to LTS obtained from Klein and Hartmann (1993). The boundary layer is simulated as in Holtslag and Boville (1993). This is another nonlocal scheme that differs from Lock et al. (2000) in the derivations and values of diffusivities and velocity scales used in the parameterizations. It also lacks a built-in cloud radiative cooling term and does not define distinct boundary layer regimes with different mixing characteristics as Lock et al. (2000) do.

To present relevant elements of the climatology and cloud variability to illuminate the model intercomparison, the calculations for Fig. 1 are repeated for the AGCMs and CGCMs and presented in Figs. 8 and 9, respectively. Between the two AGCMs, the CAM3 shows considerably greater low-level cloud variability, even though it is forced with observed SST, while AM2 levels of cloud variability more closely resemble observations. The area of low-level cloud variability in CAM3 is also concentrated in a band that straddles the equator, while AM2 shows more variability north of 0°.

![Fig. 8. Historically forced AGCMs’ basic states. June–November low-level cloud climatology (contours) and standard deviation of interannual anomalies (shading): (top) GFDL AM2 and (bottom) NCAR CAM3.](image-url)

Each coupled model shows a cold tongue that is unrealistically detached from the South American coast. They each also show the low-level June–November cloud amount variability in unrealistic places in the eastern equatorial Pacific, with CM 2.1 and CCSM3 appearing to show greatest variability over the model cold tongues.

In Fig. 10 the essential features of the observations and models are summarized and presented for intercomparison. The point where each label is plotted in Fig. 10 gives the simultaneous mean anomaly values for areas on the south side of the cold tongue (the abscissa) and the north side (the ordinate) during months in the tercile of most anomalously cold (blue label) and anomalously warm (red label) temperature advection south of the equator. The labels identify the dataset, and the two panels show cloud and SST anomalies separately. The observed values show SST advection.
anomalies of opposite sign across the equator, consistent with an SST front near 1°S latitude and southeastern trade winds in the region. Not surprisingly, the advection anomaly values for both of the AGCMs forced with actual SSTs are close to those in the observations.

The observations show a consistent negative relationship between SST advection and low-level cloud anomaly, as seen by the cool- and warm-tercile mean values falling in the upper-left and lower-right quadrants. Thus, there is a cross-equatorial dipole in cloud anomaly as well as SST advection. Of all the model sets, only the GFDL AM2 data fall in the same quadrants as the observations for cloud amount anomalies, and in fact fall relatively close to the observed values. The NCAR CAM3 falls in the lower-left and upper-right quadrants, indicating a cloud response to SST advection different from that observed in the region.

Previous studies have noted that the NCAR atmosphere model generates too much low-level cloud amount too quickly for warm SST advection conditions, presumably due to insufficient turbulent entrainment of drier air from above (Norris and Weaver 2001; Alexander et al. 2006). Another possible reason is that the CAM3 empirical cloud–LTS parameterization forces low-level cloud amount to be large for cold SST anomalies even when they also produce warm SST advection anomalies. Although not shown, multilinear regression analysis applied to CAM3 output indicates that SST advection does not explain any additional variance in

Fig. 9. As in Fig. 1, but for (top) GFDL CM 2.0, (middle) GFDL CM 2.1, and (bottom) NCAR CCSM 3.
low-level cloud amount over the cold tongue. Contrastingly, multilinear regression analysis of GFDL AM2 output results in a pattern similar to that of Fig. 3. Unlike the case for the atmosphere-only models, the GFDL CM2.0 and CM2.1 and the NCAR CCSM3 do not realistically simulate the north–south dipole behavior of SST advection anomalies in the 110°–90°W region. Figure 10 indicates that the CM2.0 and CM2.1 produce SST advection anomalies with identical sign on both sides of the equator and that the CCSM3 produces anomalies that have the correct signs but are much smaller than observed. None of the models exhibit realistic low-level cloud behavior in the 110°–90°W region.

Figure 9 indicates that each model has a cold tongue significantly detached from the coast (discussed in Wittenberg et al. 2006; Deser et al. 2006). Because some features in the coupled models are merely shifted in longitude, we searched for other 20° longitude increments that exhibited both a north–south dipole in SST advection anomalies and sufficient interannual low-level cloud variability. These longitude ranges are 125°–105°W for GFDL CM 2.0 and 130°–110°W for both GFDL CM 2.1 and NCAR CCSM3 and, as shown in Fig. 10, also do not exhibit low-level cloud anomalies with the observed cross-equatorial sign difference and negative relationship to local SST advection anomalies. The detached cold tongues, along with trade winds south of the equator that tend to be slightly more zonal than in observations, lead to SST advection cooler than observations (never warmer than 0.25 K day−1) in each coupled model in all the averaging regions south of the equator and all terciles. As a result, the clouds do not respond to anomalous SST advection as in the observations, where absolute advection changes more notably than the averaging terciles.

It is difficult to determine from this analysis whether the poor low-level cloud simulations in the CM2.0, CM2.1, and CCSM3 result from incorrect cloud or boundary layer parameterizations or fundamentally different SST and wind distributions. Whereas the meridional component of SST advection in the cold tongue is of primary importance in the observations, zonal advection is relatively more important in the models, even while meridional SST gradients are sharp. This difference might contribute to model inaccuracies in simulating the relationship between low-level cloud and temperature advection.

5. Discussion and conclusions

Satellite and surface cloud observations show that the eastern Pacific south of the equator exhibits pronounced low-level cloud variability on interannual time scales. Although SST and LTS explain some of the variance in low-level cloudiness, advection over the SST
gradient plays an important role in controlling cloud type, cloud frequency, and cloud amount over the equatorial cold tongue, especially in the region extending approximately 1500 km west from the Galápagos Islands. When SST advection is warm, the atmospheric surface layer becomes stabilized, thus inhibiting the upward mixing of moisture from the sea surface. This is evident not only in the ensuing decrease in cloud amount and more frequent absence of low-level clouds reported by the ISCCP and EECRA datasets, but also in the greater surface relative humidity, weaker surface wind speed, and less negative air–sea temperature difference.

EECRA surface cloud observations indicate that remnant stratiform and cloudless conditions occur more frequently and cumuliform clouds occur less frequently for warm SST advection, suggesting that less moisture is being transported to the cloud layer to sustain it against subsidence and entrainment drying. EPIC soundings confirm that warm SST advection is associated with a shallower boundary layer that is drier near the top.

A question that arises is why SST advection anomalies and surface layer stability so strongly influence low-level cloud variability in the southern cloud tongue region (5°S–0°, 105°–95°W). Although the 5° × 10° area immediately west of the southern cold tongue region has a similar distribution of monthly SST advection values, it is climatologically warmer and less cloudy, thus restricting the potential magnitude of interannual cloud variability. Contrastingly, the 5° × 10° area east of the southern cold tongue region is climatologically colder and more cloudy, thus reducing the sensitivity of cloud amount to warm SST advection. The southern cold tongue region is distinct in that it both experiences large variations in SST advection and climatologically occupies the phase space between overcast stratocumulus conditions and sparse trade cumulus conditions.

The observed relationship between low-level cloudiness and SST advection can be used to constrain and improve the simulation of clouds in global climate models, especially since it is independent of the relationships between low-level cloudiness and SST or LTS. The GFDL atmospheric GCM (AM2) run over prescribed observed SST produces a realistic low-level cloud response to SST advection anomalies on the southern side of the equatorial cold tongue. Two versions of the GFDL coupled ocean-atmosphere model (CM2.0 and CM2.1), however, exhibit incorrect relationships between low-level cloud and SST advection. These results suggest that the necessary advection, SST, and LTS phase space observed in the southern cold tongue region simply does not occur anywhere in the CM2.0 and CM2.1 because of their large biases in the SST and surface wind distributions. The NCAR atmospheric GCM (CAM3) run over prescribed observed SST does not produce a realistic low-level cloud response to SST advection anomalies, suggesting the presence of incorrect cloud and boundary layer parameterizations. Possible sources of error are insufficient entrainment of dry air from above the cloud layer or the parameterized requirement that low-level cloud amount increase with larger LTS, irrespective of other meteorological effects. Not surprisingly, the cloud simulation in the NCAR coupled ocean–atmosphere model (CCSM3) is also unrealistic.

The observed response of low-level cloud amount to varying SST advection over the southern cold tongue region is likely to be relevant to coastal upwelling areas and extratropical oceans where advection brings warm air over cooler water. Although beyond the scope of the present study to quantify, we note that the observed inverse relationships between SST and SST advection and between cloud and SST advection imply the existence of a negative cloud feedback on and about the near-equatorial SST front. Here, a cold SST anomaly produces anomalously warm SST advection south of the front, and the decrease in low-level cloud formation south of the front leads to a net decrease in cloud cover over the equatorial region. Simultaneously, there is decreased downward mixing of momentum and dry air to the surface in the anomalously warm SST advection sector, implying a lesser wind speed and greater near-surface humidity and a decrease in latent and sensible heat fluxes. This wind–heat flux feedback acts in concert with the cloud–insolation feedback, unlike the case in Ronca and Battisti (1997), thus underscoring its potential significance. The net increase in insolation and decrease in surface fluxes is pronounced near the climatological local SST minimum, where the ocean mixed-layer depth is often less than half of its value several degrees to the north or south, thus increasing the impact of surface fluxes on SST. The incorrect representation of this feedback in coupled ocean–atmosphere models may be a contributing factor to the development of SST biases in the equatorial cold tongue.

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APPENDIX A

Methods for Calculating Multilinear Regression

Confidence Intervals

The shaded areas indicating confidence levels in Fig. 3 come from rejecting our null hypothesis using bootstrap resampling techniques at each grid box. In this case, the null hypothesis is that the increase in cloud variance explained using the time series of SST advection in addition to SST in a multilinear regression is no greater than the increase in variance explained using a random, but physically feasible, time series in addition to SST. For the physically feasible random series, we randomly reshuffle the portion of the SST advection time series that is linearly independent of the SST series. To extract the portion of the advection series not linearly related to SST, we use

\[
\text{ADV}(t) = \beta_0 + \beta_1 \text{SST}(t) + \epsilon_{\text{ADV}}(t),
\]

where each \( \beta_i \) is optimally determined with a least squares fit. We then regress low-level cloud amount (CA) against SST and a randomly reordered sequence of the residuals, \( \epsilon_{\text{ADV}}(t) \), obtained from Eq. (A1):

\[
\text{CA}(t) = \gamma_0 + \gamma_1 \text{SST}(t) + \gamma_2 \epsilon_{\text{ADV}}(t) + \epsilon_{\text{CA}}(t).
\]

Here, the \( \gamma_i \) coefficients are again optimally determined for this equation, from least squares fit. The sequences \( t_i \) and \( t_j \) are different series of months chosen randomly, with replacement, from all the June–November over the 19-yr sequence for which there are ISCCP cloud data. In each box the length of the sequences is equal to the effective sample size, \( N_{\text{eff}} \), given by whichever is lowest among the \( N_{\text{eff}} \) values for low-level cloud amount, SST, and SST advection anomaly. For each field, \( N_{\text{eff}} = -N[2 \ln(p)]^{-1} \) where \( p \) is the one-month autocorrelation after Leith (1973).

At every grid box we assemble a distinct \( \epsilon_{\text{ADV}}(t_j) \) and compute \( \text{CA}(t_j) \) 10 000 times. We also find the variance explained by regressions using the complete CA, SST, and SST advection (ADV) fields; that is,

\[
\text{CA}(t) = \beta_0 + \beta_1 \text{SST}(t) + \epsilon_{\text{CA}}(t)
\]

\[
\text{CA}(t) = \gamma_0 + \gamma_1 \text{SST}(t) + \gamma_2 \text{ADV}(t) + \epsilon_{\text{CA}}(t).
\]

where the primes denote optimally determined coefficients for each regression that can differ from those in Eqs. (A1) and (A2). If the increase in \( R^2 \) between Eqs. (A3) and (A4) is greater than the 9500th and 9900th largest values of the change in \( R^2 \) between Eqs. (A3) and (A2), we reject the null hypothesis at the 95% and 99% confidence levels, respectively.

APPENDIX B

Confidence Intervals for Zonally Averaged Data

We obtain 95% confidence intervals for ISCCP data averaged over a latitude zone and tercile by calculating the mean of \( N_{\text{ind}} \) months randomly selected with replacement from months belonging to the tercile. All grid boxes in a latitude zone for a particular month are averaged together. This is repeated 10 000 times, and the central 9500 values of the sequentially ordered random means determine the 95% confidence interval. An initial value for the effective sample size, \( N_{\text{eff}} \), is determined for each latitude zone using \( N_{\text{eff}} = -N[2 \ln(p)]^{-1} \), where \( p \) is the one-month autocorrelation (Leith 1973). We round off the quantity \( N/N_{\text{eff}} \) and regard ISCCP values separated by this many months as independent. Months from different years are always considered independent with this method. We then count the number of independent months in each tercile to obtain \( N_{\text{ind}} \) for the ISCCP data.

Confidence intervals for EECRA data averaged over a latitude zone and tercile are determined in a similar manner, except that \( N_{\text{ind}} \) individual synoptic cloud reports are chosen with replacement from all months belonging to the tercile. Since it is very difficult to establish the degree of independence between individual synoptic reports scattered over the ocean and in time, we conservatively assume that \( N_{\text{ind}} \) for the EECRA data is half the nominal number of observations in a latitude zone and tercile.

This bootstrap resampling method does not rely on any assumptions about the probability density function of observations and is adaptable so that we can use essentially similar techniques for both the regular gridded ISCCP data and the surface observations that are irregular in space and time. It is worth noting that the zonal-mean ISCCP data show very similar error bars at almost every latitude if we simply assume a normal distribution of observations about their mean and calculate confidence intervals based on the standard deviation.

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