Marine Low-Cloud Anomalies Associated with ENSO

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ABSTRACT

As a contribution to understanding the possible impact of altered climate regimes on marine clouds, and hence on cloud radiative forcing, ship-observed marine low clouds and precipitation frequency for individual seasons are regressed at zero lag on an index of El Niño–Southern Oscillation (ENSO) for the period December 1955–January 1996 for ocean areas between 40°S and 70°N. Seasonal anomalies of atmospheric circulation parameters, static stability, and SST are also examined in order to illuminate physical mechanisms responsible for observed ENSO cloud variations.

The following extratropical regions exhibit significant ENSO cloud anomalies and are discussed in detail: winter–spring North Pacific, summer North Pacific, winter western North Atlantic, autumn northeastern Atlantic, and western Mediterranean Sea. In all of these regions except the summer North Pacific, cloud anomalies are related to jet stream and storm track anomalies associated with atmospheric teleconnection patterns. The summer North Pacific anomalies are also connected with jet stream and storm track anomalies, but these are associated with a persistent SST anomaly rather than an atmospheric teleconnection. ENSO anomalies in the western and eastern equatorial Pacific are analyzed in greater detail than in previous work, as well as those in the Arabian Sea during winter and summer monsoons. With the exception of the Arabian Sea region in winter, cloud anomalies are consistently related to 1) changes in storm tracks and/or 2) changes in low-level static stability and temperature advection.

1. Introduction

Clouds play an important role in the earth’s radiation budget. Satellite observations reveal that cloud radiative forcing (CRF) on the earth–atmosphere system has a net cooling effect that amounted to $-17 \text{ W m}^{-2}$ during April 1985–January 1986 (Harrison et al. 1990), about four times as large as expected from doubling CO$_2$ (Ramanathan et al. 1989). Slingo (1990) estimated that a 4% global increase of low-cloud amount (mostly marine) has the potential to offset doubled CO$_2$. Because marine low clouds are the major contributor to CRF, it is important to document cloud anomaly variations and their relationships to possible driving factors such as atmospheric circulations and SST changes, and to infer the mechanisms responsible for these variations. In this study we analyze cloud anomaly variations associated with ENSO.

There have been previous studies of cloud variations associated with ENSO. Deser and Wallace (1990) and Deser et al. (1993) examined ENSO-related variations of total cloud amount (TCA) and satellite-derived visible reflectance in the eastern equatorial Pacific. Bajuk and Leovy (1998b, BL hereafter) examined ENSO-related variations of different low cloud types over the tropical Pacific and Indian Oceans. Klein et al. (1999) showed that ENSO-related SST variations in the western Pacific and Indian Oceans are strongly influenced by the modulated insolation in association with ENSO-related cloud variations.

In this study, we use a near-global ocean cloud dataset with refined spatial and seasonal resolution compared with previous studies to 1) document ENSO anomalies of clouds in the western and eastern equatorial Pacific in more detail than previously; 2) document ENSO variability of clouds in regions remote from the equatorial Pacific, specifically the extratropical North Pacific, North Atlantic, and Arabian Sea; and 3) relate these ENSO anomalies of clouds to large-scale circulation and SST changes and draw inferences about the mechanisms for cloud changes from these relationships.

2. Data and analysis methods

a. Cloud variables

Cloud data used in this study are from the Extended Edited Cloud Report Archive (EECRA; Hahn and Warren 1999), which compiles individual land and ship (mainly voluntary observing ship) observations of clouds and present weather. Ship-observed cloud data

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TABLE 1. WMO low-cloud code specification.

<table>
<thead>
<tr>
<th>CL (low cloud code)</th>
<th>Nontechnical specification</th>
<th>Coding prioritya</th>
</tr>
</thead>
<tbody>
<tr>
<td>9</td>
<td>Cumulonimbus, the upper part of which is clearly fibrous (cirriform) often in the form of an anvil, either accompanied or not by cumulonimbus without anvil or fibrous upper part, by cumulus, stratocumulus, stratus, or pannus</td>
<td>1</td>
</tr>
<tr>
<td>3</td>
<td>Cumulonimbus, the summits of which at least partially lack sharp outlines but are neither clearly fibrous (cirriform) nor in the form of an anvil; cumulus, stratocumulus, or stratus may also be present</td>
<td>2</td>
</tr>
<tr>
<td>4</td>
<td>Stratocumulus formed by the spreading out of cumulus; cumulus may also be present</td>
<td>3</td>
</tr>
<tr>
<td>8</td>
<td>Cumulus and stratocumulus other than that formed from the spreading out of cumulus; the base of the cumulus is at a different level from that of the stratocumulus</td>
<td>4</td>
</tr>
<tr>
<td>2</td>
<td>Cumulus of moderate or strong vertical extent, generally with protuberances in the form of domes or towers, either accompanied or not by other cumulus or by stratocumulus, all having their bases at the same level</td>
<td>5</td>
</tr>
<tr>
<td>1</td>
<td>Cumulus with little vertical extent and seemingly flattened or ragged cumulus other than of bad weather; b or both.</td>
<td>By coverc (6 ~ 9)</td>
</tr>
<tr>
<td>5</td>
<td>Stratocumulus not resulting from the spreading out of cumulus</td>
<td></td>
</tr>
<tr>
<td>6</td>
<td>Stratus in a more or less continuous sheet or layer, or in ragged shreds, or both, but no Stratus fractus of bad weather</td>
<td></td>
</tr>
<tr>
<td>7</td>
<td>Stratus fractus of bad weather or cumulus fractus of bad weather, or both (pannus), usually below altostratus or nimbostratus</td>
<td></td>
</tr>
<tr>
<td>0</td>
<td>No stratocumulus, stratus, cumulus, or cumulonimbus</td>
<td>10</td>
</tr>
</tbody>
</table>

Although there are multiple low cloud types, the observer should report only one cloud type as representative low cloud type following this priority. For example, if there are both CL = 9 and CL = 7, the reported cloud type should be CL = 9.

b “Bad weather” denotes the conditions that generally exist during precipitation and a short time before and after.

c The cloud type that has the largest sky fraction has the highest priority among these four low-cloud types.

are used from December 1955 to January 1996. The EECRA ship data are available from December 1951, but the first four years are excluded to avoid spurious shifts in the cloud frequency in the early 1950s attributed to changes in cloud observation techniques initiated in 1949 (Wright 1986; Bajuk and Leovy 1998a).

For the cloud fields, the observer usually reports cloud amount and cloud type for each height level (low, middle, high) following a strict hierarchy from the World Meteorological Organization (WMO 1975). We will focus on the variations of low-level clouds whose nontechnical specifications and coding priorities are summarized in Table 1. Although this classification is mainly based on cloud morphology, each low cloud type is associated with its own distinct boundary layer structure and mean environmental conditions favoring its formation (Norris 1998; Norris and Klein 2000), so low cloud type can be diagnostic of boundary layer conditions. For the purpose of compact discussion within this paper, several low cloud types are grouped according to the similarities in properties and are given author-defined short names which are summarized in Table 2. Henceforth, for example, “stratocumulus” means CL = 5, 8, 4, 6, 7 unless otherwise specified.

Ship observations are inhomogeneously distributed in space and time. Also, depending on the variables, each ship observation is subject to measurement problems that can be serious noise sources, sufficient to obscure the real signals (Ramage 1984, 1987). For the cloud variables, Hahn and Warren (1999) summarized possible systematic biases associated with cloud observations and averaging processes. From individual ship observations of EECRA, we produced 5° latitude × 5° longitude regularly gridded, 3-month seasonal mean data centered on each calendar month from January 1956 to December 1995. For each low cloud or combination of low cloud types, cloud frequency (FQ), amount-when-present (AWP), and amount (AMT) are calculated. Cloud FQ for a single low cloud type or group of low cloud types is defined by the fraction of observations reporting those low cloud types among the total set of observations reporting any low cloud type information. Cloud AWP is the average low cloud cover when those low clouds are observed. Cloud AMT is the product of FQ and AWP. Detailed calculation methods used for the

TABLE 2. Author-defined short names of low cloud types.

<table>
<thead>
<tr>
<th>CL code</th>
<th>Short name</th>
<th>Abbreviation</th>
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<tbody>
<tr>
<td>1</td>
<td>Small cumulus</td>
<td>S.Cu</td>
</tr>
<tr>
<td>2</td>
<td>Large cumulus</td>
<td>L.Cu</td>
</tr>
<tr>
<td>1, 2</td>
<td>Cumulus</td>
<td>Cu</td>
</tr>
<tr>
<td>3, 9</td>
<td>Cumulonimbus</td>
<td>Cb</td>
</tr>
<tr>
<td>1, 2, 3, 9</td>
<td>Cumuliform</td>
<td>Cf</td>
</tr>
<tr>
<td>5, 8, 4</td>
<td>Stratocumulus</td>
<td>Sc</td>
</tr>
<tr>
<td>6</td>
<td>Fair weather stratus</td>
<td>FSt</td>
</tr>
<tr>
<td>7</td>
<td>Bad weather stratus</td>
<td>B.St</td>
</tr>
<tr>
<td>6, 7</td>
<td>Stratus</td>
<td>St</td>
</tr>
<tr>
<td>5, 8, 4, 6, 7</td>
<td>Stratiform</td>
<td>Sf</td>
</tr>
</tbody>
</table>
production of 3-month seasonal mean $5^\circ$ latitude $\times$ $5^\circ$ longitude regularly gridded cloud fields and precipitation $FQ$ are summarized in the appendix. The 40-yr (1956-95), 3-month mean gridded data (480 total data points for each grid box if there is no missing value) were used for the regression analysis.

b. Other variables

The EECRA contains not only cloud data but also coincident surface observations: SLP, SST, surface wind speed and direction, ship deck air temperature, and dew-point depression. As summarized by Ramage (1984, 1987), several surface variables have systematic biases that should be calibrated, especially for climate research. In this paper, however, we did not try to do any calibrations of these variables and produced 3-month seasonal mean, $5^\circ$ $\times$ $5^\circ$ gridded data using the averaging scheme used for the cloud variables. Because our ENSO index used for the simultaneous linear regression analysis has most power on the interannual time scale (Fig. 1), a spurious trend in a cloud variable or in surface wind speed, for example, should not produce serious problems. Surface divergence and vorticity are calculated from the $5^\circ$ $\times$ $5^\circ$ surface wind vector using centered differences. The $5^\circ$ $\times$ $5^\circ$ SST gridded data are further interpolated to a $5^\circ$ latitude $\times$ $10^\circ$ longitude grid for the empirical orthogonal function analysis.

In addition to the ship-observed surface variables, we analyzed the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis dataset (Kalnay et al. 1996). As proxies of synoptic storm track activity, bandpass-filtered (with cutoff frequencies at 0.1 day$^{-1}$ and 0.3 day$^{-1}$) monthly mean root-mean-square (rms) geopotential height, pressure vertical velocity, and meridional velocities were calculated using daily average values for several pressure levels (850, 700, 500, 300, 200 hPa). In addition to the rms fields, monthly mean wind vector, geopotential height, and pressure vertical velocities are used. The resulting monthly $2.5^\circ$ latitude $\times$ $2.5^\circ$ longitude fields are interpolated into 3-month seasonal mean, $5^\circ$ $\times$ $5^\circ$ fields with area weighting.

c. ENSO index and regression analysis

For the representative proxy of tropical SST variations, we did a covariance-based empirical orthogonal function (EOF) analysis using equatorial Pacific and Indian Ocean SST between 20$^\circ$S and 20$^\circ$N (see Fig. 9a for the EOF domain). Using annual cycle-filtered, 3-month seasonal mean, $5^\circ$ latitude $\times$ $10^\circ$ longitude SST anomalies from January 1956 to December 1995, the covariance matrix was constructed with area weighting [cos (lat)]. Each 3-month seasonal mean SST anomaly field is projected on the eigenvector corresponding to the maximum eigenvalue of the covariance matrix, and the resulting projection components are standardized. The first mode is the only mode well-separated from the others by the criteria of North et al. (1982). The standardized principal component for the dominant mode is our basis index representing the dominant variations of tropical SST anomalies or ENSO. To examine annual-cycle-dependent low cloud variations associated with tropical SST anomalies, seasonal ENSO indices are picked out from the above principal component without further normalization (Fig. 1). Our seasonal ENSO indices are dominated by the interannual com-
ponent with a slight regime shift toward more positive ENSO index phase during 1975–77, as noted in previous studies (Zhang et al. 1997). The ENSO indices obtained from the year-round EOF analysis are highly correlated with the eastern equatorial Pacific (5°N–5°S, 180°–90°W) SST anomalies (correlation coefficient, $r > 0.95$ for all seasons except July when $r = 0.93$).

Standardized anomalies were calculated by performing simultaneous linear regression analysis on the seasonal ENSO indices such that the anomalies displayed in the following figures represent the simultaneous linear variations associated with one standard deviation of the positive anomaly of the seasonal ENSO indices. Regression analysis was done only for the grid points where the available number of years is equal or greater than 20 out of 40 years for each season. This procedure leaves several gaps in our domain, particularly over the southeastern Pacific off South America.

To test statistical significance, we need information on the number of independent realizations of the ENSO index and regressed variables. Because our seasonal ENSO index is dominated by interannual variation components, we simply assumed that the number of independent realizations is the same as the number of years available for the regression analysis for each season and grid point. As a check, the same analysis based on the number of independent realizations calculated from the 1-lag autocorrelation (Leith 1973) of the seasonal ENSO index and regressed variables does not produce a large difference. The null hypothesis used for the statistical significance test of correlation coefficients between the ENSO index and regressed variables is the standard Student’s $t$ distribution. Statistically significant grid points at the $95\%$ and $99.9\%$ confidence levels from the two-sided Student’s $t$ test are shaded in the regression maps with light and heavy shading, respectively. When the number of independent realizations is 40, $95\%$ and $99.9\%$ confidence levels roughly correspond to correlation coefficients higher than 0.3 and 0.5, respectively. This rather approximate statistical test associated with an unrealistically high number of degrees of freedom will be complemented by showing coherent variation signals in several variables and analysis indicating responsible physical mechanisms.

3. Overview

Figure 2 shows seasonal ENSO anomaly patterns of total cloud amount (TCA). In January, a strong positive anomaly of TCA is confined between 160°E and 140°W centered at 170°W on the equator, while a strong negative anomaly is centered over the Philippines. By April, the positive TCA anomalies have stretched eastward to Central America and the northern coast of South America. The eastward extended positive anomalies weaken during the following months and by July they are replaced by negative anomalies near the northwestern coast of South and Central America, a characteristic of the July–November period (BL). Between January and July, the Indonesian negative TCA anomaly weakens and shrinks to a small area covering eastern Indonesia. By October, a significant negative anomaly is still in place close to the northwest coast of South America but the Indonesian negative anomaly has begun its seasonal expansion northward.

ENSO-related cloud anomalies are not confined in the tropical Pacific. In the next section, we will discuss in detail the following regions in which significant anomalies of TCA can be seen: 1) areas of TCA decrease in the western equatorial Pacific, 2) the eastern equatorial Pacific during northern autumn, 3) the extratropical north Pacific during winter–spring (where there are no significant TCA changes but there are significant shifts in low-cloud-type frequencies), 4) same area during July–August–September (JAS; hereafter 3-month periods are denoted by the first letter of each respective month), 5) the western North Atlantic during DJF, 6) the northeastern Atlantic and Western Mediterranean Sea during late summer and autumn; 7) the Arabian Sea during JAS, and 8) the same area during NDJ.

Figure 3 shows ENSO-related surface wind vorticity anomalies obtained from the individual ship observations. This field is shown to depict teleconnection patterns at the ocean surface. Beginning in July, surface vorticity anomalies are localized within the equatorial Pacific in a north–south dipole with little or no remote teleconnection pattern in either hemisphere. We see that the west end of this vortex dipole extends westward almost to 120°E. By October, the central Pacific dipole pattern is no longer localized within the Tropics and wave trains have developed in the Pacific in both Northern and Southern Hemispheres. The northern wave train has begun to appear as a pattern propagating away from the west-central equatorial Pacific with a negative center near 30°N, 165°E, positive center near 50°N, 170°W, and another negative center near 50°N, 135°W over the southeastern Gulf of Alaska. Compared to the well-known wintertime Pacific–North American (PNA) wave train pattern (Wallace and Gutzler 1981; see also Fig. 3a), the autumn wave train is shifted westward along with the westward shift of the equatorial Pacific vorticity and TCA anomalies during JJA and SON. In the eastern North Atlantic, we note a weak vortex pair that is related to the positive TCA anomalies over the western Mediterranean region (Park 2004). By January, the equatorial vortex dipole has shifted eastward with its western edge near 150°E. The weak autumn wave train has evolved into the more familiar PNA pattern extending into the western North Atlantic with a positive vortex anomaly centered at 35°N, 60°W. The spring wave train pattern is similar to the winter pattern but the signals are weaker, especially over the western North Atlantic. Qualitatively similar, but somewhat stronger, wave train signals are observed in the NCEP–NCAR reanalysis surface wind vorticity fields (not shown).

Figure 4 shows the annual climatology (Fig. 4a) and
Fig. 2. Seasonal ENSO regression maps of ship-observed total cloud amount (TCA). The 5° lat × 5° lon regression anomalies are smoothed using 1:2:1 zonal and 1:4:1 meridional weighting filter and the same smoothing scheme is applied for Figs. 3–4. Variable types and units of contours are shown respectively on the upper-left and upper-right side of each panel of this and following figures. Grid boxes where the number of years used for the regression analysis is smaller than 20 out of 40 are not plotted. Statistically significant anomalies at 95% and 99.9% confidence levels are shaded by light and heavy gray. The same plotting convention and statistical significance tests are applied for all the following regression maps.
ENSO regression anomalies (Fig. 4b) of ship-observed clear sky FQ for all seasons. Clear sky is frequently observed in coastal regions near the continents, while it is rarely observed over the open oceans except in the eastern equatorial Pacific and eastern equatorial Atlantic Oceans where climatological SST cold tongues exist. During positive ENSO phase, striking ENSO anomalies of clear sky FQ exist over the western and eastern equatorial Pacific. Significant decreases of clear sky FQ also appear in the Gulf of Mexico and Caribbean Sea with
similar decreases along the west coast of North and South America. Although not statistically significant at the 95% confidence level, negative anomalies are observed over the Arabian Sea and the northeastern Atlantic Ocean near Portugal and over the western Mediterranean Sea. These features have not been previously discussed but will be discussed in detail in this paper.

4. Regional analysis

a. Western equatorial Pacific

During northern winter, the tropical Pacific west of 150°E is dominated by the northeasterly winter monsoon. From the northern Philippines northward off the Asian coast, stratiform clouds, including stratocumulus, fair weather and bad weather stratus comprise the dominant winter monsoon cloud types. Frequency of these clouds decreases and frequency of cumulonimbus increases to the south. Stratiform and convective cloud frequency in this region and season are influenced by cold air advection and synoptic storm activity accompanied by cold air outbreaks from the Asian continent as well as large-scale divergence.

During positive ENSO phase in northern winter, significant negative anomalies in stratiform cloud FQ accompany a strong southerly wind anomaly corresponding to weakened northeasterly winter monsoon winds between 10° and 20°N as well as divergence anomalies between 0° and 20°N (Figs. 5a,c). In the vicinity of positive divergence anomalies, an anomalous increase in clear sky FQ occurs with a maximum clear-sky FQ anomaly where climatological clear sky FQ is maximum (Fig. 5b). Bajuk and Leovy (1998b) showed that ENSO cloudiness anomalies in this region were more closely associated with large-scale divergence changes than SST changes, which are small. Our results support the conclusion of BL and suggest that anomalous cold air advection is an additional factor responsible for ENSO cloud variations during winter.

During northern summer, significant ENSO anomalies of divergence, positive clear sky FQ, and negative convective cloud FQ (not shown) are found in the same region, but they have shrunk and retreated southward to near 125°E on the equator (Figs. 5e,f). This is consistent with the finding of BL that the ENSO anomaly of clear sky FQ is mainly associated with the decrease in cumulonimbus FQ during summer. In contrast to winter, convergence and positive stratiform cloud FQ anomalies appear north of New Guinea and over the northern Philippines approximately along a line of mean con-
Fig. 5. ENSO regression maps in the western equatorial Pacific during (left) JFM and (right) JAS. For (c) and (f) the statistical test is shown only for surface wind divergence. Nomenclature for the definition of low cloud types are given in Tables 1 and 2.

vergence formed by the seasonal migration of the ITCZ north of New Guinea (Figs 5d,f). As in winter, there is a close association between divergence anomalies and cloud FQ anomalies.

b. Autumn eastern equatorial Pacific

Figure 6 shows the ENSO anomalies of stratocumulus AMT, clear sky FQ, and static stability parameters in
the eastern equatorial Pacific during October when the eastern equatorial SST cold tongue is well developed (Mitchell and Wallace 1992). On each panel, we plotted a thick dotted line to indicate the axis of maximum SST anomaly, which is located near and slightly downstream (north) from the climatological SST trough. The region north of this line up to about 7°N (near the ITCZ) defines the SST frontal region and maximum SST anomalies near the SST trough result in weakening of the SST front.

In the eastern equatorial Pacific, climatological maximum low cloud fraction is observed over the SST frontal region. Consistent with BL, the major low cloud change responsible for the July through November negative TCA [and low cloud amount (LCA)] anomalies during positive ENSO phase is the replacement of stratocumulus by cumulus (mainly large cumulus) and, to a smaller extent, by cumulonimbus clouds. This feature is consistent with Deser et al. (1993) who showed that ENSO-related interannual variation of low cloudiness is positively correlated with the strength of the SST front. In addition to BL, we identified several new features in our work: 1) most of the observed decrease in stratiform cloud [and increase in cumulus] is over the SST frontal region in which climatological strong cold advection takes place but is weakened during positive ENSO phase (Fig. 6a) and 2) clear sky anomaly occurs in the vicinity of the SST cold tongue where clear sky is already relatively frequent (Figs. 6b and 4a). These cloud anomalies are accompanied by significant negative anomalies of lower-tropospheric stability [LS = $\theta (700 \text{ hPa}) - \theta (1000 \text{ hPa})$, Klein and Hartmann (1993)] over a broad area, which is mainly due to increased SST that is only partly compensated by free air warming (Fig. 6c). Negative anomalies of clear sky FQ seem to be closely related to the positive anomalies of surface convective instability [SI; which is diagnosed by the SST minus ship deck air temperature during nighttime (1800–0900 local hour); note that SI is an instability index] that shows significant increase near the SST trough centered at ~105°W where clear sky FQ significantly decreases.
during positive ENSO phase (Fig. 6d). We suggest the following explanation on these ENSO variation features.

Over regions where synoptic activity is minimal, an individual ship observation reporting clear sky condition indicates the absence of net condensation within the atmospheric boundary layer, that is, clear sky condition implies that moisture transport from the sea surface to the lifting condensation level (LCL) is not sufficient to overcome any drying factors near the LCL. From this point of view, clear sky FQ defined for a 5° × 5° grid box in a 3-month mean can be understood as the probability for the turbulent moisture transport from the sea surface to the LCL to be limited below a certain threshold.

Over the eastern equatorial cold tongue, net surface buoyancy flux from the sea surface to the overlying atmosphere is very weak during July–November when the equatorial cold tongue is particularly strong (Deser and Wallace 1990). Together with weak turbulence, the associated stable stratification in the lower part of the marine boundary layer (MBL) strongly limits moisture transport from the sea surface to the LCL (i.e., convective decoupling develops below the LCL), resulting in maximum climatological clear sky FQ over the SST cold tongue slightly downstream from the SST trough. A weak MBL top entrainment rate and strong subsidence rate over the eastern equatorial cold tongue is likely to increase the probability of sporadic intrusion of the MBL top inversion layer below the LCL as hypothesized by Norris (1998), resulting in a further increase of clear sky FQ. During positive ENSO phase, enhanced surface buoyancy flux associated with positive anomalies of SST and surface wind speed will strengthen boundary layer turbulence, resulting in enhanced moisture transport from the sea surface to the LCL and a decrease of clear sky FQ. Deepening of the MBL associated with enhanced MBL top entrainment rate further enhances moisture transport from the sea surface to the LCL by reducing the probability of sporadic intrusion of the MBL top inversion layer below the LCL.

At the MBL top, enhanced entrainment during positive ENSO phase can produce a significant negative buoyancy flux anomaly through sensible heating and evaporative drying at the MBL top. Reduced turbulent activity and enhanced stratification in the upper MBL, in turn, fosters decoupling between the LCL and inversion base. This additional decoupling limits moisture transport from the lower MBL to the upper MBL, resulting in the dissipation of overlying stratuscumulus clouds. Cumulus clouds develop within the conditionally unstable upper MBL and, when well developed, these clouds may act to further dissipate overlying stratuscumulus clouds by enhancing MBL top entrainment drying as suggested by Bretherton (1992) and Bretherton and Wyant (1997). As a result, clear sky FQ and stratuscumulus cloud AMT (and so LCA) both decrease while cumulus cloud AMT increases in the vicinity of the equatorial cold tongue from negative to positive ENSO phase. The downstream shift of maximum stratuscumulus anomalies from the maximum SST anomalies is likely to be due in part to finite adjustment time of MBL air columns embedded in the flow. Note that the distribution pattern of stratuscumulus cloud anomaly (and so the LCA anomaly) do not fit well the distribution pattern of LS, suggesting that something more is going on here than the simple LS relationship of Klein and Hartmann (1993). These results suggest other factors to be accounted for in boundary layer models.

During positive ENSO phase, the FQs of clear sky and the “no low cloud” condition over the eastern equatorial cold tongue decrease in all months but especially significant decreases occur during May–June and November–December. The area-averaged divergence anomaly reaches its maximum negative value (maximum convergence anomaly) during June and SST anomaly reaches its maximum during November in the cold tongue region (5°N–5°S, 130°–90°W). These anomalies and the associated destabilized MBL may contribute to positive precipitation anomalies (not shown) as well as negative clear sky anomalies.

c. Winter–spring extratropical North Pacific

During northern winter and spring, the well-known PNA wave train circulation pattern is the dominant Northern Hemisphere teleconnection pattern, and it has a dominant influence on atmospheric circulation over the extratropical North Pacific. The Aleutian low is anomalously strong and surface westerly winds and the storm track in the central and eastern North Pacific shift southward during positive ENSO phase (Figs. 7d,e). These shifts are not associated with major changes in TCA but the frequencies of low cloud types contributing to TCA do shift significantly.

Figures 7a–c show ENSO-related cloud anomalies over the extratropical North Pacific during winter–spring. South of the Aleutian Islands, cumulonimbus FQ significantly increases during winter (Fig. 7a) and these anomalies continue into spring with significant enhancement extending to the California coast in March. Large negative anomalies of stratuscumulus occur off the California coast together with weaker increases in both types of stratus (Figs. 7b,c). Patchy increases in stratus extend westward across the Pacific from the California coast together with a well-defined band of positive anomaly of precipitation FQ (not shown, but the ENSO anomaly pattern of precipitation FQ is similar to Fig. 7c with more distinct zonal structure). Anomalous zonal structure of stratus and precipitation FQ is stronger during spring than winter. The negative stratuscumulus anomaly is compensated in part by stratus and in part by an increase in cumulus FQ, which shows significant positive anomalies centered off Baja California. (not shown).

Some of the mechanisms responsible for these cloud anomalies are illustrated Figs. 7d–f. During positive
ENSO phase, perturbations of 200-hPa geopotential height associated with the PNA-type wave propagation elongates the wintertime North Pacific subtropical jet eastward with a southward shift. As a result, positive anomalies of synoptic storm activity diagnosed by the 700-hPa rms $w$ occur in a band extending from $35^\circ N$, $160^\circ E$ to a strong center near $30^\circ N$, $140^\circ W$ and in a patch centered near $30^\circ N$, $130^\circ E$ (Fig. 7e), roughly coinciding with the patterns of increased stratus and precipitation FQ. In addition, strong mean upward motion anomalies (700-hPa mean $v$) occur just south of South Korea and off the California coast in association with positive anomalies of stratus and precipitation FQ (Fig. 7f). The increase in cumulonimbus FQ is due to reduced midtropospheric static stability [$=\theta(500 \text{ hPa}) - \theta(1000 \text{ hPa})$], which is related to enhanced cold air advection in the vicinity of the enhanced cold low pressure south of the Aleutians (Fig. 7d). We conclude that the anomalous synoptic storm activity, mean vertical velocity, and midtropospheric static stability patterns associated with the PNA wave propagation are responsible for the anomalies in stratus, precipitation, and cumulonimbus FQ over the winter North Pacific.

Off the California coast, weakening of the prevailing northeasterly surface winds during positive phase ENSO induces positive SST and warm advection anomalies, both of which are unfavorable for stratocumulus (Fig. 7d, Norris and Leovy 1994; Klein 1997). Reduced mean subsidence upstream of the center of the stratocumulus anomalies may also contribute to a reduction of stratocumulus downstream through the influence of upstream subsidence anomalies on downstream inversion height and strength. Klein et al. (1995) and Klein (1997) have demonstrated a dependence of stratocumulus amount on upwind environmental conditions such as SST, upper-air temperature, and surface wind divergence. Finally, enhanced synoptic activity and associated disruption of the prevailing northeasterly winds and cold advection probably contribute significantly to the negative stratocumulus anomaly in this region (Fig. 7e).

**d. Summer extratropical North Pacific**

ENSO-related cloud anomalies during northern summer are more surprising. In July, a positive TCA anomaly extends eastward from the Yellow Sea to $170^\circ W$ (Fig. 2c) and by August this band extends eastward to $130^\circ W$, reaching maximum zonal extent. The dominant contributing cloud pattern to this TCA (or LCA) anomaly is an increase in FQ of stratocumulus coupled with
a decrease of cumulus, but an increase of bad weather stratus and precipitating midlevel clouds also contribute (Figs. 8a–c). Not shown here, significant positive anomalies of cumulus FQ and negative anomalies of clear sky and fair weather stratus FQ are also observed along the coast of California and Baja California. As the late summer/autumn wave train pattern develops (Fig. 3d), the storm track, cloud, and precipitation anomalies shrink toward the central North Pacific from both east and west, and by September and October, they are concentrated in the region of anomalous westerlies in the southern part of the cyclonic vortex induced by Rossby wave propagation over the mid North Pacific (Fig. 2d).

Norris (2000) showed that during summer, dominant interannual and interdecadal variations of marine stratiform cloud and nimbostratus over the North Pacific are strongly coupled to latitudinal shifts of storm track and SST gradient without corresponding coupling to large-scale circulation anomalies. Consistent with the Norris results, we find that ENSO-related summertime cloud variations are accompanied by a southward shift of the zonally elongated isotherms of SST and synoptic storm activity, with maximum anomalies over the latitudinal band where the meridional gradient of climatology is maximum (Figs. 8e,f). However, we also observed a significant southward shift of the summer North Pacific midlatitude jet in association with the north–south gradient of 200-hPa height anomalies (Fig. 8d). This implies that ENSO-related summertime variations of SST–storm track–MBL clouds are also coupled to the variations of upper-level mean atmospheric circulation, although the ENSO signals in surface mean flows are weak.

A simple hydrostatic calculation indicates that the jet change along 35°N corresponds to a 200-hPa height gradient anomaly of about 20 m over the latitude range 30°–40°N in the central part of the North Pacific. This corresponds to a mean temperature gradient change from the 1000 to 200 hPa layer of about 0.5 K, which is very close to the observed SST gradient anomaly through the same latitude range. Thus, the jet change can be attributed to the SST change if the SST anomaly propagates undamped through the entire troposphere and no additional tropospheric thermodynamic mechanisms are required (consistent with no propagated teleconnection pattern in this season). However, the jet change is not trivial in its consequences because the jet change is associated with 700-hPa rms $\omega$ changes; that is, the storm track anomaly pattern is closely tied to the jet anomaly pattern. Thus, the ENSO anomaly pattern
of SST gradient that persists into summer is accompanied by atmospheric structural and dynamical changes that can account for the low cloud anomalies. In turn, enhanced low cloud amount can reinforce anomalous SST by reflecting incoming solar radiation.

The similar behavior of the dominant regional coupled variations examined by Norris (2000) and ENSO-related coupled variations examined here implies a significant role of ENSO on the coupled interannual and interdecadal variations of SST–storm track–MBL clouds over the summer North Pacific. In order to test this hypothesis, we did a singular value decomposition (SVD) analysis (Bretherton et al. 1992) of TCA anomalies in the North Pacific (20°–60°N, 120°E–120°W) coupled with tropical SST anomalies (20°S–20°N, 50°E–70°W) during JAS. Figure 9 shows the resulting expansion coefficients and heterogeneous maps obtained by regression on the expansion coefficients of the other variables. The SVD statistics in Fig. 9a indicate strong interannual coupling variations between midlatitude TCA and tropical SST during JAS. Spatial distribution patterns are very similar to the ENSO regression maps (compare with Fig. 2c during JJA) and the corresponding expansion coefficients are well correlated with the time series of area-averaged (30°–45°N, 170°E–140°W) TCA and SST (Figs. 9b,d). Although not shown here, the simultaneous correlation coefficient between the principal component of TCA empirical orthogonal basis for the region 20°–60°N, 120°E–120°W and the ENSO index is 0.58 during JAS. The same analysis using LCA produced similar but slightly weaker coupling variations than TCA. All of these analyses indicate that interannual cloud variations in the summertime midlatitude North Pacific are strongly coupled with ENSO. We note that both the SVD expansion coefficients and time series of locally averaged TCA and SST indicate significant regime shifts across the mid-1970s. This suggests that the epochal increases of summer TCA and stratocumulus FQ in the midlatitude North Pacific between 1955–70 and 1980–95 (Norris and Leovy 1994) are largely due to ENSO.

In both northern winter and summer, the axis of synoptic storm activity shifts southward during positive ENSO phase, but the effect of the shifted storm track on the stratocumulus FQ is opposite in the northeastern subtropical Pacific off the California coast and in the central and western North Pacific during the two seasons. During winter, positive ENSO favors less strato-
cumulus in the northeastern subtropical Pacific but during summer, positive ENSO favors a more frequent occurrence of stratocumulus over the central and western North Pacific (Figs. 7b and 8b). We attribute this difference to the different character of the weak cold advection that favors stratocumulus in the two regions: over the northeastern subtropical Pacific, cold advection is the norm but is disrupted by synoptic activity; over the central and western North Pacific, cold advection follows synoptic low pressure systems.

e. Winter western North Atlantic

ENSO-related TCA anomalies in the western North Atlantic start to develop east of Florida in October (Fig. 2c). During the following months, significant anomalies extend westward to the Gulf of Mexico and eastward and northeastward to the central North Atlantic. In February, they reach maximum strength over the Gulf of Mexico and Caribbean Sea and extend westward across Central America to the eastern tropical Pacific and eastward to near Portugal. After February, these anomalies slowly drift southward and dissipate, leaving no significant signals in September.

Figure 10 shows ENSO-related anomalies of clouds and precipitation in the western North Atlantic during January. During northern winter–spring, positive anomalies of LCA (or TCA) are associated with an increase of stratocumulus and stratus AMT (especially bad weather stratus) and precipitation FQ but a decrease of cumulus FQ. The increase in precipitation FQ is consistent with the finding of Ropelewski and Halpert (1987) that precipitation increases at stations around the Gulf of Mexico during positive ENSO phase in October–March.

During northern winter and spring, ENSO-related cloud variations in the western North Atlantic are related to the storm track shift associated with the combined effects of zonally symmetric tropical warming and the PNA pattern. In response to the associated trough–ridge pattern of positive ENSO phase, there are pronounced upper-level westerly anomalies and more synoptic activity in the eastern North Pacific Ocean and in the Gulf of Mexico, and these synoptic activity anomalies, analyzed here as 700-hPa rms $\omega$ anomalies, extend east-northeastward across the central North Atlantic (Fig. 10d). During northern winter, more frequent and/or stronger cold air outbreaks from the adjacent North American continent into the Gulf of Mexico may further contribute to the formation of inversion-topped MBL clouds.

Similar to the summer North Pacific, winter low cloud anomalies over the western North Atlantic are associated with a change in upper-level jet and storm track. But unlike the summer North Pacific where these changes seemed to occur because of the residual SST change,
rather than an atmospheric teleconnection, the changes in the western North Atlantic occur due to propagation of the well-known PNA teleconnection. It is interesting to note that enhanced synoptic activity associated with the PNA teleconnection has opposite influences on stratocumulus AMT over the northeastern subtropical Pacific (decreased stratocumulus) and over the western North Atlantic (increased stratocumulus) during northern winter–spring.

The low cloud amount anomalies in this region, like the TCA anomalies, are remarkably large and could significantly affect the energy budget at these low latitudes even during winter. Over the western North Atlantic (20°–30°N, 95°–65°W) during DJF, climatological (1985–89) net CRF obtained from the Earth Radiation Budget Experiment (ERBE; Barkstrom 1984) data is \(-11\, \text{W m}^{-2}\) (SW CRF = \(-34\, \text{W m}^{-2}\) and LW CRF = \(23\, \text{W m}^{-2}\)) and the ratio of anomalous LCA (and TCA) to climatology (1956–95) is 6%. Thus, for a one standard deviation positive ENSO anomaly, anomalous net CRF at the top of the atmosphere would be \(-0.7 \sim -2\, \text{W m}^{-2}\) with a maximum negative value when TCA anomalies are due to cloud variations with low cloud-top height. Assuming that the ENSO variation of LW radiation at the sea surface is negligible, the empirical formula of Reed (1977) estimates anomalous net CRF at the sea surface to be \(-4 \sim -5\, \text{W m}^{-2}\) for a one standard deviation positive ENSO anomaly.

f. Autumn northeastern Atlantic and western Mediterranean Sea

During July–November, TCA shows significant positive correlation with ENSO in the eastern North Atlantic and western Mediterranean Sea centered near Portugal (Figs. 2c,d). These TCA anomalies are associated with increased FQ of stratiform cloud (mainly bad weather stratus), precipitation and related upper-level clouds, and a significant decrease of clear sky FQ (not shown). Simultaneous correlation coefficients between the domain-averaged (30°–45°N, 20°W–15°E) TCA and the ENSO index are statistically significant at the 95% confidence level during July–November with the highest correlation in September \((r = 0.47, 0.60, 0.75, 0.70,\) and 0.64 from July to November). More than 55% of interannual variance of the northeastern Atlantic and western Mediterranean TCA are explained by the simultaneous variations of eastern equatorial Pacific SST anomalies during September.

Park (2004) discussed in detail the association of sky conditions in this region with the ENSO. He showed that the strong ENSO teleconnection of western Mediterranean sky conditions during the August through October season arises partly from a previously unreported quasi-stationary Rossby wave propagating eastward from the central-west Pacific across the midlatitude North Pacific, North America, and the North Atlantic. The surface marine manifestation of this pattern appears in Fig. 3d. By analyzing the long-term record (1875–1997), Park also found that this remote ENSO influence on the western Mediterranean TCA has shown inter-decadal variations with epoch lengths of about 30 years.

g. Summer Arabian Sea

The northern Indian Ocean is subject to a strong monsoon circulation regime. During May–September, surface winds are dominated by the southwesterly summer monsoon (Fig. 11c) while, during November–March, the northeasterly winter monsoon circulation prevails and, in April and October, the monsoon circulation is in the transition stage. During August, a strong southwesterly surface wind and associated SST cold tongue along the southeast coast of the Arabian Peninsula favors the formation of stratocumulus clouds over the central Arabian Sea (Figs. 11b,c). In addition, monsoon-related convection over the Indian subcontinent produces similar maxima in FQ of precipitating stratus over the eastern Arabian Sea, resulting in maximum FQ of stratiform cloud (\(-55\%\)) centered at 19°N, 63°E just downstream from the minimum SST and west of India. This corresponds well to the location of maximum climatological lower-tropospheric stability.

During positive ENSO phase, summer TCA shows negative anomalies in the Arabian Sea (Fig. 11a) associated with the decrease of stratocumulus and stratus (Fig. 11b) and increase of cumulus cloud (see BL). These cloud variations seem to be associated with a modulation of the southwesterly monsoon circulation and SST. During positive ENSO phase, the southwesterly monsoon circulation weakens and associated weakening of coastal upwelling results in the weakening of cold atmospheric advection and significant positive SST anomalies centered near the SST cold core east of the Arabian Peninsula (Fig. 11d). Warm SST and weakened cold advection are likely to promote the transition of stratocumulus cloud into cumulus cloud, similar to the ENSO-related MBL cloud transition observed in the SON eastern equatorial Pacific Ocean and FMA northeastern subtropical Pacific Ocean (see Figs. 6a and 7b). At the same time, reduced convection over India will decrease bad weather stratus cloud FQ, resulting in the decrease of stratiform cloud but increase of cumulus cloud over the summer Arabian Sea during positive ENSO phase.

In order to gain some insight into the characteristic environments controlling the formation of individual low cloud types over the Arabian Sea, we did a compositing analysis using individual ship-observed low cloud information and daily NCEP–NCAR reanalysis data. First, as indicated in Fig. 10e, we chose a 5° × 5° grid box where significant ENSO signals are observed in cloud fields for the JAS season. Then, we picked out the days when a specific low cloud type or a group of low cloud types (e.g., stratocumulus, bad weather stratus, and cumulus) was reported in the given grid box.
for each season during 1979–95; that is, we produced daily cloud indices. Among the total of 1564 days available, ship observers in the chosen grid box reported stratocumulus on 776 days, bad weather stratus on 445 days, and cumulus on 802 days during JAS. Using these daily cloud indices as compositing bases, we produced composite fields of the surface wind vector and lower tropospheric stability obtained from the daily NCEP–NCAR reanalysis dataset and then calculated composite anomalies by subtracting climatologies. Because the chosen compositing indices are not mutually independent for a given season (i.e., a specific day can be used for the compositing of both stratocumulus and bad weather stratus, e.g.), we will call this procedure heterogeneous compositing.

Compositing results in Figs. 11e,f clearly show that stratocumulus and bad weather stratus during summer favor strong summer monsoon flows and strong lower-tropospheric stability while cumulus favors weak summer monsoon flows and weak lower-tropospheric stability (not shown for cumulus). This implies that the variations of summer low clouds over the Arabian Sea are governed by MBL dynamics, consistent with the ENSO-related summer cloud variations discussed above. Note that the decrease (increase) of stratocumulus clouds with reduced (enhanced) cold advection
is a consistent message, as also documented in the other regions (e.g., winter western equatorial Pacific, autumn eastern equatorial Pacific, winter and summer extratropical North Pacific, and winter western North Atlantic).

h. Winter Arabian Sea

In contrast to summer, ENSO-related TCA during winter shows positive anomalies that are associated with increased FQ of stratocumulus and stratus and significant decrease of clear sky FQ without any systematic variations of cumulus clouds (Figs. 12a–c).

During December, ENSO-related surface mean wind anomalies over the Arabian Sea are very weak and seem not to affect the seasonal mean northeasterly monsoon circulation (Fig. 12f). Significant positive anomalies of SST and associated decrease in static stability are observed over the Arabian Sea (Figs.12d,f), suggesting a contrary relationship to the increased stratiform cloud to that discussed above. The significant decrease of clear sky FQ and absence of systematic decrease of cumulus cloud further imply that the mechanism responsible for the winter cloud variations is not associated with MBL dynamics of the type observed during summer.

As was done for summer, we did heterogeneous compositing over a grid box during the NDJ season as indicated in Fig. 12g. Among the total of 1564 days available, ship observers in the chosen grid box reported stratocumulus on 376 days, bad weather stratus on 70 days, and cumulus on 997 days. The relatively small number of days used for compositing bad weather stratus produced somewhat unstable signals but an interpretable overall pattern.

In contrast to summer, winter cumulus cloud is formed under normal climatological conditions (not shown) while stratocumulus and bad weather stratus are formed when the lower troposphere becomes destabilized and low-level positive vorticity anomalies are developed over the Arabian Sea. Thus, we speculate that winter cloud variations during positive ENSO phase over the Arabian Sea may be associated with enhanced activity of cyclonic disturbances. But it is not clear why these circulation anomalies might be more frequent or strong in positive ENSO phase, so the complete physical mechanism is unclear.

5. Summary and conclusions

In this paper, we examined seasonal marine low cloudiness and precipitation FQ variations associated with ENSO by individual cloud types. Combined analysis of SST, atmospheric flows, and static stability parameters gave insights into the physical mechanisms responsible for marine cloud variations on the interannual time scale. In addition to ENSO-related anomaly patterns in the tropical oceans, previously reported by BL and described here in greater spatial and temporal detail, significant anomalies are found in regions remote from the Tropics. Overall summaries of ENSO-related marine cloud variations are given in Table 3.

In the western (and central) equatorial Pacific, ENSO cloud anomalies are closely related to large-scale divergence changes in association with anomalous Walker and Hadley circulations. However, in the autumn eastern equatorial Pacific where the SST cold tongue is well developed, ENSO cloud anomalies are mainly due to the SST changes and associated variations of MBL static stability. In the vicinity of the equatorial SST cold tongue, the lower boundary layer tends to be more coupled to the surface moisture source while the MBL top (inversion layer) is more decoupled during positive ENSO phase. This results in the reduction of both clear sky FQ and stratocumulus amount but enhanced cumulus during the positive ENSO phase.

In the winter–spring extratropical North Pacific, anomalous atmospheric flows (anomalous upper-level jet, storm track, and mean vertical velocity) in association with the PNA wave propagation pattern are responsible for enhanced deep convective clouds south of the Aleutian Islands, a zonal band of enhanced precipitation FQ across the midlatitude North Pacific, and negative (positive) anomalies of stratocumulus (stratus and cumulus) off the California coast. In the summer extratropical North Pacific, southward shifts of low clouds are closely coupled to the southward shifts of the SST gradient zone, upper-level jet, and storm track that, in contrast to winter, is likely to be the result of residual SST change rather than atmospheric teleconnection.

In the winter western North Atlantic, positive anomalies of TCA (also LCA, stratocumulus, bad weather stratus, and precipitation FQ) during positive ENSO phase are mainly due to the equatorial shift of storm track in association with the PNA pattern. LCA anomalies in this region are remarkably large and may have a significant influence on the energy budget through CRF. In the autumn northeastern Atlantic and Western Mediterranean Sea, TCA, stratiform cloud, precipitation FQ, and related upper-level clouds significantly increase during positive ENSO phase. As discussed in Park (2004), this remote teleconnection arises partly from a quasi-stationary Rossby wave that is somewhat similar to the winter wave propagation pattern, PNA, but has shorter wavelength and is shifted westward in the North Pacific.

In the summer Arabian Sea, reduced TCA during positive ENSO phase is accompanied by decreases of stratocumulus and stratus and an increase of cumulus, associated with the weakening of southwesterly monsoon circulation and increased SST. Modulation of MBL decoupling structure is likely to be responsible for these cloud variations similar to the autumn eastern equatorial Pacific. In contrast to summer, TCA during winter shows positive ENSO anomalies with increased FQs of stratocumulus and stratus but a significant decrease of clear sky FQ without any systematic variations of cumulus
Fig. 12. As in Fig. 11 but during NDJ. (a)–(d), (f) ENSO regression maps, (e) climatologies, and heterogeneous compositing maps of NCEP–NCAR reanalysis surface wind vector and lower tropospheric static stability for (g) stratocumulus and (h) bad weather stratus. In (a)–(d), climatologies are also plotted by thin solid lines.
### Table 3. Significant ENSO anomalies of marine low cloud and precipitation FQ.

<table>
<thead>
<tr>
<th>Region</th>
<th>Season</th>
<th>TCA</th>
<th>Cloud type</th>
<th>Cloud type</th>
<th>Precipitation FQ</th>
<th>Physical mechanisms</th>
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<td>Central equatorial Pacific and northern SPCZ</td>
<td>All</td>
<td>+</td>
<td>Cb</td>
<td>Cu</td>
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<td>+ Rising motion associated with anomalous Walker and Hadley circulations</td>
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<td></td>
<td></td>
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<td>Sf</td>
<td>(mainly S.Cu)</td>
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<td>+ Surface convergence</td>
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<tr>
<td>Western equatorial Pacific and southern SPCZ</td>
<td>All</td>
<td>−</td>
<td>Cu</td>
<td>Cb</td>
<td>−</td>
<td>+ Sinking motion associated with anomalous Walker and Hadley circulations</td>
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<td></td>
<td></td>
<td></td>
<td>Clear sky</td>
<td>SI</td>
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<td>+ Surface divergence</td>
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<td></td>
<td>+ Midtropospheric static stability</td>
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<td></td>
<td>− Cold air outbreak</td>
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<tr>
<td>Eastern equatorial Pacific</td>
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<td>−</td>
<td>Cu</td>
<td>Sc</td>
<td>+</td>
<td>+ SST</td>
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<td></td>
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<td></td>
<td>(mainly L.Cu)</td>
<td>FSt</td>
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<td>− LS</td>
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<td>Cb</td>
<td>Clear sky</td>
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<td>+ SI</td>
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<td></td>
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<td></td>
<td>+ Surface convergence</td>
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<tr>
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<td>St</td>
<td>Cb</td>
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<td></td>
<td>Cu</td>
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<td>• Synoptic storm activity</td>
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<td>• Midtropospheric static stability</td>
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<td>• Upper-level jet</td>
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<td>Jul ~ Oct</td>
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<td>Sc</td>
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<td></td>
<td>+ SST</td>
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<td>St</td>
<td>Sc</td>
<td>+</td>
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<td>Cu</td>
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<td>Jul ~ Oct</td>
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<td>FSt</td>
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<td>Clear sky</td>
<td></td>
<td>+ Synoptic storm activity</td>
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<td></td>
<td>Nov ~ Jun</td>
<td></td>
<td>Sf</td>
<td>Cu</td>
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<tr>
<td>Eastern North Atlantic and Western Mediterranean Sea</td>
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<td>Sf</td>
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<td></td>
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<td>Arabian Sea</td>
<td>May ~ Sep</td>
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<td>−</td>
<td>Cu</td>
<td>SI</td>
<td>+ Weakening of summer monsoon</td>
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<td></td>
<td>Nov ~ Mar</td>
<td></td>
<td>−</td>
<td>Sf</td>
<td>Clear sky</td>
<td>• Near the coast of California and Baja California</td>
</tr>
</tbody>
</table>

* Positive anomaly during the positive ENSO phase.

* Negative anomaly during the positive ENSO phase.

* During winter.

* South of Aleutian Islands extending to the California coast.

* Near the coast of California and Baja California.
clouds. Physical mechanisms responsible for winter cloud variations are unclear, but we speculate that they might be related to enhanced activity of cyclonic disturbances over the Arabian Sea.

In summary, outside of the deep convection region, ENSO-related marine low cloud anomalies appear to be associated with either 1) documentable changes in storm track and/or 2) changes in boundary layer structure associated with changes in low-level static stability and temperature advection. In most areas, anomalous temperature advection (either by anomalous atmospheric flows and/or by anomalous SST gradient) has significant influences on the ENSO modulation of MBL cloudiness.

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APPENDIX

Construction of 3-Month Seasonal, 5° Latitude × 5° Longitude Gridded Data

Cloud FQ for a single low cloud type or group of low cloud types is defined by the fraction of observations reporting those low cloud types among the total set of observations reporting any low cloud type information. In this study, all observations without any low cloud type information were proportionally partitioned into the low cloud types excluding observations without any low cloud information. Here, we summarize the methods used for the calculation of cloud FQ, AWP, and AMT from the EECRA dataset. Hereafter, \( \langle X \rangle \) symbolizes the number of observations satisfying the condition of \( X \) in a given space and time averaged domain.

The number of total observations, \( T \), is

\[
T = \langle \text{CL} = (-1, 0, 1, 2, 3, 4, 5, 6, 7, 8, 9, 10, 11) \rangle,
\]

(A1)

where CL is low cloud code in EECRA (CL = -1: no low cloud-type information, CL = 10: sky obscuration by thunderstorm shower, CL = 11: sky obscuration by fog; see Table 1 for the others). The number of gray observations, \( G \), is defined by the number of observations without any low cloud type information:

\[
G = \langle \text{CL} = -1 \text{ and } N \neq 0 \rangle,
\]

(A2)

where \( N \) is total cloud cover. The number of independent observations, \( I \), is defined by the number of observations independent of the gray observations, i.e. that is, none of the gray observations can have low cloud types of the independent observations:

\[
I = \langle \text{CL} = (10, 11) \text{ or } N = 0 \rangle.
\]

(A3)

The number of dependent observations, \( D \), is defined by the number of observation not independent from the gray observations; that is, any of the gray observations could have low cloud types of dependent observations in real situations:

\[
D = \langle (\text{CL} = 0 \text{ and } N \neq 0) \text{ or } \text{CL} = (1, 2, 3, 4, 5, 6, 7, 8, 9) \rangle.
\]

Then,

\[
T = G + I + D.
\]

(A5)

Assume that \( C \) is the number of observation for a group of low cloud types in a given space and time-averaged domain [e.g., \( C = (\text{CL} = (3, 9, 10)) \)]. If \( C \) is the number of independent observations among the chosen group of low cloud types [e.g., \( C = (\text{CL} = 10) \); Eq. (A3)], nonbiased cloud frequency, \( \text{FQ}(C) \) for the chosen cloud group is calculated as follows. In other cases, a missing value is assigned to \( \text{FQ}(C) \):

\[
\text{FQ}(C) = \frac{1}{T} \left[ C + (C - C_{\text{I}}) \left( 1 + \frac{G}{D} \right) \right], \quad \text{if } T \neq 0
\]

and \( D \neq 0 \) \( \text{FQ}(C) = \frac{C_{\text{I}}}{T}, \quad \text{if } T \neq 0 \)

and \( D = 0 \) and \( I \neq 0 \). \( \text{FQ}(C) = \frac{\sum N_{\text{I}} \langle \text{CL} = (\text{Explicit}) \rangle}{C}, \quad \text{if } C \neq 0 \). \( \text{AWP}(C) = \frac{\sum N_{\text{I}} \langle \text{CL} = (\text{Explicit}) \rangle}{\text{FQ}(C) \text{AWP}(C)}, \quad \text{if } \text{FQ}(C) \neq \text{Missing} \)

and \( \text{AWP}(C) \neq \text{Missing} \) \( \text{AMT}(C) = 0, \quad \text{if } \text{FQ}(C) = 0 \) and \( \text{AWP}(C) = \text{Missing} \)

TCA and clear sky FQ are calculated as follows:

\[
\text{TCA} = \frac{\sum N}{T}, \quad \text{FQ( CLR)} = \frac{\langle N = 0 \rangle}{T}. \quad \text{AWP}(C) = \frac{\sum N_{\text{I}} \langle \text{CL} = (\text{Explicit}) \rangle}{\text{FQ}(C) \text{AWP}(C)}
\]

“Precipitation events” were defined by the sum of the following present weather codes following Hahn and Warren (1999): drizzle (WW = 50–59), rain (WW = 60–69), snow (WW = 70–75, 77, 79), and thunder-
storm/shower (WW = 80–99). Precipitation FQ is simply calculated as follows:

\[
FQ(\text{precipitation}) = \frac{\langle WW = (50 \sim 75, 77, 79, 80 \sim 99) \rangle}{\langle WW = (-1, 0, 1, 2, \ldots, 97, 98, 99) \rangle}
\] (A10)

In the calculation of cloud variables (FQ, AWP, AMT) and precipitation FQ, the data from the WMO’s Historical Surface Temperature (HSST) Data Project (identified by card deck number 150–156) were not used due to the absence of information about cloud type and present weather code (however, in calculation of TCA and clear sky FQ, we incorporated the HSST data). In addition, the observations executed under poor illumination conditions are prefiltered following a screening criteria developed by Hahn et al. (1995) (however, in calculation of precipitation FQ, we did not apply this criteria).

After calculating cloud variables (FQ, AWP, AMT), TCA, and FQs of clear sky and precipitation using the methods described above for the respective local daytime (0600–1800) and nighttime (1800–0600) observations, a whole day average is calculated by giving the same weighting to the respective daytime and nighttime average. If either the daytime or nighttime average value is missing, the other is used as the whole day average. Because we constructed 3-month seasonal mean, 5° × 5° gridded data, monthly sampling errors due to the preferential usage of data near full moon periods and geographical sampling bias can be reduced. Because of this, all of the systematic errors in the cloud type analysis (Hahn and Warren 1999)—night detection bias, day–night sampling bias, monthly sampling bias, clear sky bias, sky obscured bias, and geographical sampling bias—are minimized.

In order to reduce random errors due to the small number of observations, 5° × 5° grid boxes in which the observation number is less than 10 in a 3-month period were discarded and filled by linear interpolation using the nearest grid values in the zonal and meridional direction. When both east–west and north–south values were available, the final interpolation value was calculated by giving the zonal interpolation twice the weight of the meridional interpolation. This interpolation process was repeated twice. The same observation number filtering and interpolation process are also applied for the noncloud ship-observed variables.

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