Dynamic and Thermodynamic Relations of Distinctive Stratus Clouds on the Lee Side of the Tibetan Plateau in the Cold Season

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ABSTRACT

Given the large discrepancies that exist in climate models for shortwave cloud forcing over eastern China (EC), the dynamic (vertical motion and horizontal circulation) and thermodynamic (stability) relations of stratus clouds and the associated cloud radiative forcing in the cold season are examined. Unlike the stratus clouds over the southeastern Pacific Ocean (as a representative of marine boundary stratus), where thermodynamic forcing plays a primary role, the stratus clouds over EC are affected by both dynamic and thermodynamic factors. The Tibetan Plateau (TP)-forced low-level large-scale lifting and high stability over EC favor the accumulation of abundant saturated moist air, which contributes to the formation of stratus clouds. The TP slows down the westerly overflow through a frictional effect, resulting in midlevel divergence, and forces the low-level surrounding flows, resulting in convergence. Both midlevel divergence and low-level convergence sustain a rising motion and vertical water vapor transport over EC. The surface cold air is advected from the Siberian high by the surrounding northerly flow, causing low-level cooling. The cooling effect is enhanced by the blocking of the YunGui Plateau. The southwesterly wind carrying warm, moist air from the east Bay of Bengal is uplifted by the HengDuan Mountains via topographical forcing; the midtropospheric westerly flow further advects the warm air downstream of the TP, moistening and warming the middle troposphere on the lee side of the TP. The low-level cooling and midlevel warming together increase the stability. The favorable dynamic and thermodynamic large-scale environment allows for the formation of stratus clouds over EC during the cold season.

1. Introduction

Clouds are an important component of climate system and are among the largest sources of spread in the

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the surface temperature, which can further change the cloud amount by altering the static stability (Klein and Hartmann 1993; Yu et al. 2004, hereafter Yu04). Low stratus clouds in the marine boundary layer (MBL) cause the largest simulation uncertainty in the tropical ocean regions (30°S–30°N) (Bony and Dufresne 2005). However, the discrepancy associated with the stratus radiative effect over eastern China (EC) has attracted little attention. The EC stratus is important to the local climate because of the intense cloud feedback in cold season (Yu04). The representation of EC stratus in GCM is critical to the surface temperature simulation (Zhou and Yu 2006). In the following paragraph, we will show that the EC stratus is also a major source of cloud radiative effect simulation error in current GCMs.

Figure 1 displays the root-mean-square error (rmse) in the results from 16 models in the phase 5 of the Coupled Model Intercomparison Project (CMIP5) and the Clouds and the Earth’s Radiant Energy System (CERES) satellite data during the boreal cold season (November–April). The maxima centers of the SWCF are primarily located in the tropical ocean [especially the southeast Pacific Ocean (SPO)] and eastern China. The midlatitude stratus in summer also constitutes a large portion of errors, because of large stratus amount. However, the longwave cloud forcing (LWCF) maxima centers are primarily found in the tropics and the
Tibetan Plateau (TP), with an overall deviation lower than that of the SWCF. Many studies have revealed that cloud radiative forcing (CRF) associated with tropical convective activity is a major source of uncertainty in climate models (Wyant et al. 2006; Williams and Webb 2009; Ichikawa et al. 2011). In contrast, in EC and the SPO the large rmse results from the poorly simulated stratus radiative effects, which peak in both regions during this period. Although the simulation errors in two regions are similar, the dominant cloud properties and the cloud formation mechanism are quite different.

The SPO is regarded as the most persistent MBL stratocumulus deck in the world (Bretherton et al. 2004). Simulating the MBL stratocumulus in the (sub)tropics using climate models has long been recognized as difficult. The thin MBL stratocumulus typically lies under a sharp inversion and is maintained by a blend of complex physical processes that are difficult to parameterize (Bretherton et al. 2004). Many studies reported in the literature have examined the properties of MBL low clouds (e.g., George and Wood 2010; Lin et al. 2009; Chen and Cotton 1987). Stratocumulus clouds tend to occur in an environment with strong lower-tropospheric stability, large-scale subsidence, and a sufficient supply of surface moisture (Wood 2012). Some studies have investigated the response of clouds or CRF to environmental changes by sorting them into different dynamic and thermodynamic regimes (Bony et al. 2004; Eitzen et al. 2011; Williams et al. 2003; Norris and Iacobellis 2005). Weaver and Ramanathan (1997), Stowasser and Hamilton (2006), and Eitzen et al. (2011) all demonstrated that low clouds and their associated SWCF in marine regimes featuring large-scale subsidence do not change significantly with the vertical velocity at 500 or 700 hPa.

As the only continental stratus region (Klein and Hartmann 1993; Wood and Bretherton 2006), EC also experiences large simulation errors in the current models (Fig. 1). However, stratiform clouds over EC are quite different from those over the SPO. The midlevel cloud over EC (primarily nimbostratus and altostratus) peaks in the cold season (Yu et al. 2001) and the stratus amount dominates. Yu04 have shown the importance of these clouds to local climate. Stratus clouds modulate the surface air temperature, which influences the low-level static stability and relative humidity and further alters the cloud amount. Yu04 also reported that the stratus clouds over EC result from mechanical forcing and the stable stratification downstream of the TP. Moreover, the seasonal variation corresponds well to the divergence difference between the midlevel (600 hPa) and the low level (850 hPa) (Li and Gu 2006).

Although these earlier studies pointed out the relation between influencing factors and stratus, the way these factors influence cloud has not been described in detail. A comprehensive mechanism analysis is needed to better understand the formation of EC stratus. This paper offers a new physical explanation about how these factors impact cloud. In consideration of the large CRF simulation error in current climate models, this paper first presents the difference between EC and MBL stratus. Further, the way that the general circulation influences stratus, the formation mechanism of strong stability in cold season, and the connection between dynamic and thermodynamic factors are described in detail. We think that the results of this paper can help to enhance the understanding of EC stratus so as to improve the simulation of these clouds in GCMs.

This paper is organized as follows. Section 2 introduces the data and the choice of thermodynamic and dynamic agents. The comparison will be shown in section 3. In section 4, we focus on the sources of the dynamic contributed moisture and the thermodynamic stable stratification, both of which favor the formation of stratiform clouds over EC during the cold season. A summary and discussion are contained in section 5.

2. Data and method
a. Data description

We use the GCM simulator-oriented International Satellite Cloud Climatology Project (ISCCP) monthly product (2.5° × 2.5° resolution), which is analogous to the ISCCP D2 monthly averages with several minor differences (see http://climse rv.ipsl.polytechnique.fr/cfmp-obs/). Clouds in the cloud-top pressure/cloud optical thickness joint histogram at the full resolution (6 × 7 categories) are further classified into nine types according to the ISCCP cloud definitions (Rossow and Schiffer 1999). We focus on four types of optically thick stratiform clouds at middle and low levels: nimbostratus, altostratus, stratocumulus, and stratus (“stratus” refers to stratiform clouds in this paper except here). The Pathfinder Atmosphere (PATMOS) monthly cloud data (0.5° × 0.5°) from 1982 to 2007 are used to determine the water cloud amount. This dataset uses the difference between the 0.63- and 3.7-μm albedos to separate the water clouds from the snow and ice (Stowe et al. 2002). A station observational dataset from the China Meteorological Administration is used to obtain the total cloud amount from 1980 to 2005.

The CERES Energy Balanced and Filled (EBAF) monthly product at 1° × 1° resolution from 2001 to 2010 is used as the shortwave and longwave radiance fluxes at
the top of atmosphere (TOA). This product is generated using an objective algorithm to constrain the shortwave and longwave fluxes at the TOA within their range of uncertainty to adjust the average global net TOA flux and the heat storage in the earth–atmosphere system into consistency (Loeb et al. 2009). The clear-sky and all-sky TOA fluxes are further used to calculate the SWCF and LWCF according to Ramanathan et al. (1989). Net cloud radiative forcing is the algebraic sum of the SWCF and LWCF. More details can be found online (see http://ceres.larc.nasa.gov/documents/DQ_summaries/CERES_EBAF_Ed2.6r_DQS.pdf).

The European Centre for Medium-Range Weather Forecasts (ECMWF) Interim Re-Analysis (ERA-Interim; 1.5° × 1.5° resolution; Uppala et al. 2008) is used to obtain the wind field, vertical velocity, specific humidity, relative humidity, and temperature. To validate the reliability of the reanalysis data, we compared several variables with L-wave band station radiosonde observational data obtained from the China Meteorological Administration. This product has been processed with strict quality control.

Figure 2 shows the seasonal variations of the vertical cross section of relative humidity in the mid to low troposphere (500 to 1000 hPa) and the static stability (potential temperature difference between 500 and 850 hPa). Both variables are averaged in EC (27°–32°N, 103°–118°E) from January 2007 to February 2010. The distribution of relative humidity in the reanalysis closely resembles that in the observation (Figs. 2a,b). The large vertical gradient of the relative humidity between 600 and 800 hPa from November to April and the maxima center in the lower troposphere in summer are all captured in the reanalysis. The discrepancy mainly lies in the following: the low-level relative humidity in the reanalysis is lower than that observed between January and April, especially in February. Similarly, two datasets have close stability (Fig. 2c), both of which reach their maximum in the cold season, which is consistent with the inversion layers evident in the large vertical gradient of relative humidity between 500 and 850 hPa (Figs. 2a,b). The consistent results reaffirm our choice of using the reanalysis data for further studies.

b. Choice of dynamic and thermodynamic agents

The dynamic and thermodynamic factors must represent the environment that is favorable to the formation of stratus. We select the dynamic proxy to be identical to the vertical velocity at 700 hPa (\(\omega_{700}\)) to respectively represent the large-scale lifting and subsidence conditions for EC and the SPO. Additionally, a \(U\) component at 600 hPa and a \(V\) component at 850 hPa are used as the horizontal circulation indicators over EC because the midlevel divergence is mainly attributed to the zonal wind, while the meridional divergence is primarily responsible for the low-level convergence.
We use different thermodynamic proxies for the two regions to better reflect the stable stratification in the troposphere. For the SPO, an estimated inversion strength (EIS) is used according to Eitzen et al. (2011). The EIS is proposed in Wood and Bretherton (2006) as an improved representation of the boundary inversion with a higher correlation with stratiform low clouds than its predecessor, lower-tropospheric stability (LTS) (Klein and Hartmann 1993). The EIS is calculated as follows:

$$EIS = \theta_{700} - \theta_0 - \Gamma_{m}^{850} (Z_{700} - \text{LCL}),$$

in which $\theta_{700}$ and $\theta_0$ are the potential temperatures at 700 hPa and the surface, $\Gamma_{m}^{850}$ is the moist-adiabatic potential temperature gradient at 850 hPa, $Z_{700}$ is the geopotential height at 700 hPa, and LCL is surface-based lifting condensation level. Differently, because the inversion layer over EC is higher and thicker than marine low cloud regimes [Fig. 2; see Wood (2012) as a review for marine stratus], we use the potential temperature difference between 500 and 850 hPa as the thermodynamic factor to represent the static stability in accordance with previous studies on EC (Yu04; Li and Gu 2006; J. Li et al. 2005).

3. Comparing the response of stratus clouds to dynamic and thermodynamic agents between EC and the SPO

From November to April, stratiform clouds in EC and the SPO reach their largest amounts during the year (Klein and Hartmann 1993; Yu et al. 2001). As shown in Fig. 3a, the stratiform cloud fraction in both regions exceeds 60% and is the largest between 60°S and 60°N. The PATMOS data also indicate high water cloud amounts in both domains (Fig. 3b). The high coverage of stratiform clouds and high water content significantly weaken the incoming solar flux at the surface, resulting in an intensive SWCF in both regions (Fig. 3c). Moreover, because of reduced high cloud coverage and weakened role of stratus in emitting longwave radiation, the resulting strong negative net cloud radiative effect (Fig. 3d) is responsible for a decrease in the surface temperature. Cooling at the surface level enhances the low-level stability, which favors an increase in the stratus cloud amount. The positive cloud feedback in EC and the SPO has been already recognized (Yu04; Clement et al. 2009; Eitzen et al. 2011).

Although both domains have similar stratus maxima centers, strong negative NCRF, and positive cloud feedback, the interannual responses of the clouds and CRF...
to dynamic and thermodynamic factors differ. As in previous studies (Williams et al. 2003; Eitzen et al. 2011), we use a two-dimensional histogram in which cloud and CRF are binned according to the stability and vertical velocity to measure their responses to environmental changes. To remove outliers, only bins whose occurrence frequency is higher than 0.1% are shown. The periods of cloud and CRF data range from 1984 to 2007 and from 2001 to 2010, respectively. We select data only from the boreal cold season. Thus, 144 monthly samples are taken for cloud and 60 monthly samples for CRF at each grid point. The long-term monthly mean of each grid point is removed. Note that the vertical motion environment differs substantially between these two domains. The long-term monthly mean of vertical velocity in EC is primarily negative, indicating the dominant ascent motion, while the SPO is dominated by subsidence.

Figure 4 presents the result for the SPO. A stronger EIS leads to increased stratus cloud amount and more intense SWCF and NCRF, whereas \( \omega_{700} \) has little influence on the cloud amount and SWCF (NCRF). Therefore, the interannual responses of stratiform clouds and the associated SWCF (NCRF) are largely dominated by the variation in the thermodynamic, not dynamic, factor. The result confirms the findings in previous studies (Weaver and Ramanathan 1997; Stowasser and Hamilton 2006; Eitzen et al. 2011), which indicated that marine low clouds and their associated SWCF remain almost unchanged when the vertical velocity at 500 or 700 hPa varies in regimes of large-scale subsidence. Figures 4e and 4f shows the occurrence frequency for each bin; the dominance of small variation bins within \( \pm 10 \text{ hPa day}^{-1} \) in the \( \omega_{700} \) anomaly and \( \pm 1 \) K in the EIS anomaly is evident.

Figure 5 further presents the case for eastern continental China on the lee side of the TP. Unlike the SPO, stratiform clouds over EC have evident responses to both stability and vertical velocity anomalies (Fig. 5a). When the stability and \( \omega_{700} \) anomalies fall into quadrants that are favorable to an increase in cloud amount and positive stability (+stability) and negative \( \omega_{700} \) (−\( \omega_{700} \)) anomalies, the stratiform cloud amount increases and the SWCF (NCRF) strengthens (Figs. 5b,c). The unfavorable condition can be characterized as negative stability (−stability) and positive \( \omega_{700} \) (+\( \omega_{700} \)) anomalies. The cloud amount and CRF in this quadrant are weaker than the climatological mean. Moreover,
when bins fall into quadrants with only one favorable condition, the response of the cloud or CRF becomes a competing result for each variable.

The analysis also suggests that the occurrence frequency of the bins in the quadrants with two favorable or unfavorable anomalies is higher than that in the quadrants with only one favorable anomaly. The total occurrence frequencies for the four quadrants during two periods are given by the percentages in Figs. 5d and 5f. The frequency in the positive stability, $\omega_{700}$, and negative stability, $-\omega_{700}$, quadrants accounts for up to a dominant 60%, demonstrating that if the variation in $\omega_{700}$ is favorable for an increase or decrease of stratiform cloud amount, the stability will be more likely to change in the same way, altering the cloud amount over the interannual time scale. Therefore, these two factors can be coherent.

4. The sources of moist and stable stratification over the downstream of the TP

Previous studies have primarily focused on the dynamic aspect of the midlevel stratus over EC. The nimbostratus and altostratus result from the blocking and friction effects of the TP (Yu04). The midlevel cloud amount over the Sichuan basin (27°–32°N, 103°–108°E), as the maximum center of stratus, exhibits the best correlation with the local divergence difference between 600 and 850 hPa seasonally (Li and Gu 2006). Stable stratification is also important for the formation of stratus clouds (Yu04; Li and Gu 2006). Figure 5 confirms the response of stratiform clouds and the associated CRF to both $\omega_{700}$ and stability anomalies, and further suggests that the dynamic and thermodynamic factors are related. The following section will document the cloud formation mechanism associated with dynamic and thermodynamic relations. The general circulation influences stratus through regulating local moisture and static stability. The high stability formed in cold season is contributed by both midlevel warming and surface cooling. In particular, the midlevel warming is correlated with the topography in southwest China.

Abundant moisture at middle and low levels is a prerequisite for the formation of stratiform clouds. Figure 6a confirms this assertion by providing the time series of the cold-season mean cloud amount and the vertically integrated specific humidity (700–500 hPa). The cloud amount data contain both total cloud obtained from an
The dynamic circulation directly contributes to the accumulated water vapor at the middle and low levels. First, the correlation coefficient between the moisture and $\omega_{700}$ (Fig. 6b) is $-0.61$ (significant at a confidence level of 0.01), suggesting a tight coupling between the low-level large-scale lifting and the water vapor downstream from the TP. The rising motion stems from the midlevel divergence and the low-level convergence, both of which result from the midtropospheric westerly passing by the TP and the low-level meridional winds converging downstream. Thus, the variation in the horizontal wind fields can lead to further variation in the rising motion. Meanwhile, horizontal advection is another important means of transporting water vapor. Therefore, we calculate the correlation coefficients between the vertically integrated moisture over EC and 1) the 600-hPa zonal wind ($U_{600}$) and 2) the 850-hPa meridional wind ($V_{850}$) to examine the connection between the moisture and horizontal circulation.

The $V_{850}$ (Fig. 7a) at the east Bay of Bengal (region A), which is dominated by the southerly component, is positively correlated with moisture. The southern (20°–27°N) and northern (30°–40°N) areas of the Chinese mainland are mainly covered by a positive and negative correlation pattern. A small fraction (regions B and C) of the observational data (1980–2005) and the ISCCP stratiform clouds (1984–2007). The correlation coefficients between the cloud amounts from the two datasets and the specific humidity are 0.55 (station) and 0.45 (ISCCP), both of which exceed the confidence levels of 0.01 (station) and 0.05 (ISCCP), respectively, according to a Student’s $t$ test.

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clinging to the lee side of the TP is significant. The U600 (Fig. 7b) in the southwest China (region D) has a significant positive correlation with the moisture, while U600 at the northeast corner of the TP (region E) is negatively correlated. The geopotential height fields also present consistent results. We calculate the geopotential height anomaly by finding the geopotential height difference in the composite during the positive ISCCP stratiform cloud anomaly months and the climatological mean. At 850 hPa, the geopotential height decreases within a large area from the east Bay of Bengal to southwest China, contributing to the strengthening of the southwesterly winds at the mean trough front (not shown). At 600 hPa, the geopotential gradient is the largest downstream of the TP and is favorable for an increase in the westerly direction. The results confirm that moisture and stratus are not only subject to the local large-scale rising motion but are also influenced by the horizontal circulation fields at the middle and low levels.

The high moisture environment alone is not sufficient for the formation of a large stratus amount. The stable stratification, which tends to cap moisture within the middle level and generate condensation, is another favorable prerequisite. To find the relation, Fig. 8 provides the correlation coefficients between the EC-averaged stability (27°–32°N, 103°–118°E) and the horizontal temperature fields at six levels from the lower (975 hPa) to middle (500 hPa) troposphere. At the low levels (975 to 850 hPa), the positive correlation coefficients are primarily located in the east Bay of Bengal. The significant negative correlation area concentrates to the north of 24°N. The domain of the largest negative correlation is in Mongolia, which is the origin of the Siberian high, a near-surface shallow high pressure system. In the middle levels (600 and 500 hPa), the positive correlation area moves northeastward. At 500 hPa, the significant positive correlation area covers a large fraction of the east China mainland; the largest positive correlation coefficients are concentrated in southwest China (22°–28°N, 10°–115°E).

This correlation pattern is similar to that in Figs. 7a and 7b; both have significant positive correlations at the midlevel in southwest China and at the low level in the east Bay of Bengal. According to this pattern, we select a diagonal cross section (shown by the solid line in Fig. 8) to illustrate the formation mechanism for high moisture and strong stability during the cold season. The beginning and end of this diagonal line are located at 16°N, 80°E and 32°N, 120°E, respectively, which cover the east Bay of Bengal, the southwestern and northeastern
flanks of the HengDuan Mountain–YunGui Plateau (HDM-YGP) and EC downstream of the TP. The intervals at the longitude and latitude lines are 2° and 0.8°, respectively.

The left column in Fig. 9 describes the mean (from 1980 to 2009) temperature and moisture fields along this diagonal cross section to which variables are interpolated using the inverse-distance-weight method. As shown in Fig. 9a, the temperature between the northeastern and southwestern flanks of the HDM-YGP varies greatly below 700 hPa. The northeastern flank is dominated by thick low-temperature air. The vertical temperature only ranges within 10 K; thus, the potential temperature gradient from the mid to low level is large (Fig. 9b). Although the specific humidity at this side is not large compared with the southwestern flank (Fig. 9c), the maxima center of relative humidity is still evident due to the surface low temperature downstream of the plateau.
In contrast, the southwestern flank is dominated by warm moist air, which is lifted by the topography (Fig. 9a) and blown eastward by the midtropospheric westerly winds. Thus, this current can further moisten and warm the atmosphere downstream of the TP.

The right column in Fig. 9 shows the field differences between the average for the months of positive cloud anomaly and the climatological mean. During the positive cloud anomaly period, the low-level northerly anomaly and midlevel southerly anomaly carry cold and warm air, respectively, to EC (white contour in Fig. 9e), causing midlevel warming and low-level cooling (Figs. 9e,f) and enhancing the stability. The rising motion anomaly downstream of the plateau brings more water
vapor to the midlevel (Fig. 9e), moistening and saturating the air, especially between 600 and 850 hPa (Figs. 9g,h). A small region between 100° and 105°E, from 850 to 700 hPa (Figs. 9g,h), also exhibits the positive temperature anomaly. Figure 7 shows the low-level southerly components at the east Bay of Bengal and the midlevel westerly downstream of the TP as positively correlated with vertically integrated moisture over EC, suggesting that the strengthening of the wind component at these regions can increase the moisture transport. To further validate the connection with the temperature field, we calculate the correlation coefficients between the EC averaged temperature at 500 hPa (T500), 850 hPa (T850), and the diagonal vertical temperature section. The temperature field positively correlated with T850 lies primarily below 700 hPa downstream (Fig. 10a). The section positively correlated with T500 exhibits a “Γ” shape, which forms a slope from the southwestern flank of the HDM-YGP in the low level to the midlevel downstream of the TP.

Based on the previously mentioned facts, we give the following description on the cloud formation mechanism associated with dynamic and thermodynamic relations. The horizontal circulation and large-scale lifting contribute to the accumulated water vapor, which is a prerequisite for abundant stratus clouds. The stable stratification, leading to saturated air, is also connected with the circulation. Therefore, the dynamic factor not only directly contributes to the cloud by transporting moisture but also exerts an influence by regulating the thermodynamic factor. The manner in which the circulation affects the moisture and stability is as follows: At low levels, the strengthening of the southerly component from the east Bay of Bengal increases the water vapor transport. This branch of warm, moist flow is uplifted to the midlevel by the topography at the southwestern flank of the HDM-YGP. The midtropospheric westerly further advects the warm flow to the lee side of the TP, confining the stratus below the midlevels. Accompanied by the surface cold advection from the Siberian high, which primarily results in reducing the low-level temperature, the circulation directly leads to a moist and stable environment over EC. A recent study also indicated that surface cold advection seem to be more dominant in contributing the inversion layer over central China (Li et al. 2012).

The circulation can also regulate the moisture at the middle and low levels by changing the local divergence. At the low level, the southerly in southwest China and the northerly in north China favor low-level convergence. At the midlevel, the strengthening of the westerly on the lee side of the TP and the weakening of the westerly at the northeast corner of the TP intensify the midlevel divergence over EC. The increase in divergence difference between the middle and low levels further strengthens the local vertical moisture transport. The synthesis of analysis above also confirms that the stability and $\omega_{700}$ anomalies are coherent as described in section 3. The increase in the midlevel westerly speed downstream of the TP favors divergence and warm advection, causing an increase in $\omega_{700}$ and stability, respectively.

This article reveals the connection between the circulation strength and the stratiform cloud amount
during the cold season. Because winter circulation over EC is influenced by many factors such as the Siberian high (Wu and Wang 2002; D’Arrigo et al. 2005), winter Arctic Oscillation (Gong et al. 2001), and the North Atlantic Oscillation (NAO) (J. Li et al. 2005; Sung et al. 2010), certain connections may be established between the stratus amount and the circulation strength, and thereby regulating the local climate. For example, J. Li et al. (2005) demonstrated that the midtropospheric westerly over the North Atlantic and North Africa intensifies during positive NAO phases. The enhanced westerly, after passing over the TP, leads to strengthened ascending motion on the lee side of the plateau, which favors the formation of stratiform clouds and intensifies the local positive cloud feedback.

5. Summary and discussion

a. Summary

This study reveals the distinction between stratiform clouds over EC and representative marine stratus clouds. The formation of the environment over EC during the cold season with high moisture and stability is examined. The major conclusions are summarized as follows:

1) Stratus clouds over EC and the MBL regions (e.g., the SPO) cause large uncertainty in the simulation of climate models. The MBL clouds are primarily governed by the thermodynamic inversion strength. Stratus clouds over EC are particularly influenced by variations in both the dynamic circulation and thermodynamic stability. The positive (negative) stability and negative (positive) $\omega_{(0)}$ anomalies are favorable (unfavorable) to the formation of stratiform clouds.

2) During the cold season, stratiform clouds over EC result from a large quantity of accumulated water vapor at the middle and low levels. Accompanied by a capping inversion, the moist and stable stratification leads to a favorable environment for the occurrence of midlevel stratus. The formation of this ambient field is interpreted as follows: At low levels, a branch of a warm, moist flow from the east Bay of Bengal is uplifted by the topography at the southwestern flank of the HDM-YGP. The midtropospheric westerly further advects this current to the lee side of the TP, leading to moistening and warming at the midlevels and favoring divergence, which increases the rising motion downstream of the TP. The surface cold advection from the Siberian high primarily results in a cooling of the low-level temperature. Thus, the stability is increased by the combined effect of midlevel warming and low-level cooling to further contribute to stratus formation.

b. Discussion

The distinction between the stratus over EC and other MBL stratus suggests that simultaneously simulating them in the GCM for both regions is difficult (e.g., Fig. 1). The simulation of the cloud radiative feature in the models is critical to their overall performance. In particular, any deviation from the simulated cloud radiation can further lead to a spread in the simulating surface temperature changes. Zhou and Yu (2006) demonstrated that many models are unable to accurately simulate the surface temperature variations downstream of the TP due to the uncertainty in the cloud feedback process.

This work aims to provide a reference for evaluating and improving the cloud radiative performance in the climate models over EC. Based on the facts of this paper, the authors evaluated the simulation of stratus radiative effect in CMIP3 and CMIP5 models (Zhang and Li 2013). The results suggest that most models in CMIP3 and CMIP5 cannot simulate a correct response of EC stratus radiative effect to large-scale controls, and even a reasonable ambient field.

Previous research has confirmed that the simulation of midlevel stratiform clouds over EC can be improved greatly if the dynamic and thermodynamic conditions are well represented (Y. Li et al. 2005). To achieve better performance, a model should accurately reproduce the circulation fields associated with the orographic effect. The midlevel divergence confines the accumulated moisture contributed by the large-scale lifting and uplifted low-level southwesterly below the middle levels. The resulting favorable ambient field with high moisture and strong stability downstream of the TP is important to the formation of stratiform clouds.

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CORRIGENDUM

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In Zhang et al. (2013), Fig. 3 was processed incorrectly and the land values in panel (b) were mistakenly omitted. The figure appears below as it was meant to appear originally. The staff of the Journal of Climate regrets any inconvenience this error may have caused.

REFERENCE


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FIG. 3. Climatological mean during the boreal cold seasons (November–April) between 60°S and 60°N for (a) stratiform cloud amount (%), period: 1984–2007, (b) water cloud amount (%), period: 1982–2007, and (c), (d) SWCF and NCRF, respectively (W m$^{-2}$, period: 2001–10).