Wintertime Weakening of Low-Cloud Impacts on the Subtropical High in the South Indian Ocean

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ABSTRACT: To elucidate the unique seasonality in the coupled system of the subtropical Mascarene high and low-level clouds, the present study compares wintertime cloud radiative impacts on the high with their summertime counterpart through coupled and atmospheric general circulation model simulations. A comparison of a fully coupled control simulation with another simulation in which the radiative effects of low-level clouds are artificially switched off demonstrates that the low-cloud effect on the formation of the Mascarene high is much weaker in winter. Background climatology plays an important role in this seasonality of the Mascarene high reinforcement. Relative to summer, the suppression of deep convection due to low-level clouds that acts to reinforce the high is much weaker in winter. This arises from 1) seasonally lower sea surface temperature in concert with the smaller sea surface temperature reduction due to the deeper ocean mixed layer and the weaker cloud radiative effect under weaker insolation and 2) seasonally stronger subtropical subsidence associated with the Hadley circulation in winter. As verified through atmospheric dynamical model experiments, enhanced cloud-top radiative cooling by low-level clouds acts to reinforce the wintertime Mascarene high in comparable magnitude as in summer. The present study reveals that the self-sustaining feedback with low-level clouds alone is insufficient for replenishing the full strength of the wintertime Mascarene high. This implies that another internal feedback pathway and/or external driver must be operative in maintaining the wintertime high.

KEYWORDS: Indian Ocean; Subtropics; Anticyclones; Clouds; Air-sea interaction

1. Introduction

Over the subtropical ocean, surface winds are strongly influenced by an anticyclonic high pressure system known as a subtropical high. It develops under the descending branch of the zonal-mean Hadley circulation (e.g., Held and Hou 1980; Dima and Wallace 2003), constituting a zonally asymmetric anticyclone over each of the ocean basins (e.g., Rodwell and Hoskins 2001; Miyasaka and Nakamura 2005, 2010). A subtropical high plays a crucial role in shaping climatological patterns of winds, temperature, and precipitation through atmospheric and oceanic processes as well as their interactions (e.g., Nakamura 2012). Deepening our understanding of a subtropical high is therefore beneficial for various aspects of climate science.

A subtropical high is considered to constitute a positive feedback system with low-level clouds through air–sea interactions. Over the eastern portion of a subtropical high, the equatorward surface winds invoke a cool eastern boundary current, coastal upwelling, upper-ocean mixing, and surface

evaporation, acting to maintain relatively low sea surface temperature (SST) underneath (e.g., Seager et al. 2003). Suppressing deep convection and acting as an anticyclonic potential vorticity anomaly at the surface, the lowered SST reinforces the high via local air–sea interactions (e.g., Seager et al. 2003; Miyasaka and Nakamura 2005, 2010; Miyamoto et al. 2021). Here, low-level clouds prevailing over the eastern portion of the high cool the underlying ocean owing to their high albedo and thus reinforce the high (e.g., Seager et al. 2003; Miyasaka and Nakamura 2005, 2010; Miyamoto et al. 2021). They can also reinforce the high directly through cloud-top longwave cooling (e.g., Miyasaka and Nakamura 2005, 2010; Miyamoto et al. 2021). Since a subtropical high facilitates the formation of low-level clouds by enhancing midtropospheric subsidence and lowering SST (e.g., Klein and Hartmann 1993; Wood and Bretherton 2006; Wood 2012; Koshiro and Shiotani 2014; Miyamoto et al. 2018), there should be self-sustaining coupling between low-level clouds and a subtropical high.

While the role of low-level clouds over the South Pacific in shaping equatorial asymmetry of SST and precipitation under the wind–evaporation–SST (WES) feedback (Xie and Philander 1994) has been well documented (e.g., Philander et al. 1996; Ma et al. 1996; Yu and Mechoso 1999; Gordon et al. 2000; Xie et al. 2007), this climatological self-sustaining feedback with low-level clouds has recently been substantiated by Miyamoto et al. (2021) for the summertime subtropical Mascarene high over the south Indian Ocean (SIO). By artificially switching off low-cloud radiative effects in a coupled general circulation model (CGCM), Miyamoto et al. (2021) have quantitatively demonstrated that the substantial portion of the summertime Mascarene high is maintained by the low-cloud feedback. Furthermore, they

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FIG. 1. Climatological distribution of the top-of-atmosphere cloud radiative effect (CRE; W m^{-2} ; color shaded unevenly as indicated at the bottom) in (a) DJF, (b) MAM, (c) JJA, and (d) SON in the CERES-EBAF dataset. Superimposed with orange and red contours are low-cloud fraction (LCF; every 15%) based on the ISCCP dataset and zonally asymmetric sea level pressure (SLP*; every 1.5 hPa; solid and dashed lines for positive and negative values, respectively; zero contours are omitted) based on ERA5, respectively. The land area is masked. (e)–(h) As in (a)–(d), respectively, but for the CM_CTL run. In (e)–(h), black rectangular box indicates the domain where low-level clouds are made artificially transparent in the CM_NoCRE run. See the text for details.

have identified fundamental reinforcement mechanisms for the summertime Mascarene high through atmospheric dynamical model experiments. Suppression of deep convection due to lowered SST under the high albedo of low-level clouds induces a Rossby-wave response (Matsuno 1966; Gill 1980), which explains the major portion of the Mascarene high reinforcement. The cloud-top longwave cooling by low-level clouds emphasized by earlier studies is found to play a secondary role in its reinforcement. Unlike in the Pacific and Atlantic, these feedback processes are localized over the subtropical SIO, because surface westerlies along the equator suppress the equatorial asymmetric mode (e.g., Kawamura et al. 2001; Schott et al. 2009).

In winter, the Mascarene high strengthens and shifts westward, unlike the other subtropical highs (Fig. 1c; Rodwell and Hoskins 2001; Lee et al. 2013; Miyamoto et al. 2018; Xulu et al. 2020), and thus an area of large low-cloud fraction (LCF) expands across the entire subtropical SIO (Fig. 1c; Miyamoto et al. 2018). Seasonally enhanced storm-track activity along the oceanic frontal zone associated with the warm Agulhas Current system (e.g., Nakamura and Shimpo 2004) in addition to the formation of a subtropical SST front also acts to increase the wintertime subtropical LCF by enhancing cold advection and scalar wind speed (Miyamoto et al. 2018). Miyamoto et al. (2021) noted, however, that the reinforcement of the Mascarene high

by low-level clouds simulated in the CGCM is much weaker in winter than in summer. The present study aims at elucidating the mechanism for the reinforcement of the wintertime Mascarene high by low-level clouds simulated in our CGCM experiments and contrasting it with its summertime counterpart. Together with our previous works, this study will help understand the unique seasonality in the coupling of the Mascarene high with low-level clouds.

The rest of the paper is organized as follows. Section 2 describes data and model experiments. Section 3 presents the main results of this study: section 3a discusses seasonality in the ocean cooling by low-level clouds, section 3b compares wintertime and summertime radiative impacts of low-level clouds on the Mascarene high, and sections 3c and 3d discuss seasonality in low-cloud impacts on the Mascarene high via SST and in-atmosphere radiative cooling, respectively. Section 4 discusses the importance of the lowered SST in reinforcing the wintertime Mascarene high. Section 5 provides a summary and implications of this study.

2. Data and model experiments

a. Observational data

Our CGCM simulations described in the next subsection are compared with the following observational data. For sea level pressure (SLP), the ERA5 global dataset (Hersbach et al. 2020) with a 1° horizontal resolution is used for the period from 1979 to 2018. For LCF, we use the International Satellite Cloud Climatology Project (ISCCP) data (Rossow and Schiffer 1991, 1999) from July 1998 to June 2008, since a geostationary satellite was moved to over the Indian Ocean in 1998 (Evan et al. 2007). The LCF with a 2.5° grid interval has been evaluated under the random overlap assumption. We also use the Clouds and the Earth's Radiant Energy System (CERES) Energy Balanced and Filled (EBAF) edition 4.1 (NASA/LARC/SD/ASDC 2019) on a 1° grid for radiative fluxes at the top of the atmosphere (TOA) from March 2000 to February 2020. The ocean mixed layer depth (MLD) simulated in our CGCM is compared with the Mixed Layer Dataset of Argo, Grid Point Value (MILA-GPV) for the 2001–18 period. The horizontal resolution is 1° in both longitude and latitude. The MLD is defined as a depth at which potential density difference is 0.03 kg m^{-3} relative to the surface. This difference corresponds to buoyancy difference of 0.00029 m s^{-2} with typical seawater density (1026 kg m^{-3}) and the acceleration of gravity (9.8 m s^{-2}), which is close to the definition of MLD employed in our CGCM (buoyancy difference of 0.0003 m s^{-2}).

b. Coupled general circulation model experiments

The model experiments used in this study are the same as employed in Miyamoto et al. (2021), from which some of the texts in the following three subsections are derived with minor modifications. This study utilizes the Geophysical Fluid Dynamics Laboratory (GFDL) Coupled Model version 2.1 (CM2.1; Delworth et al. 2006), with the resolution of 2° latitude $\times$ 2.5° longitude for the atmosphere and land. The resolution of the ocean model is 1° in both latitude and longitude in the extratropics, while the meridional resolution equatorward of

TABLE 1. Overview of the CGCM (top two rows) and AGCM (bottom two rows) experiments.

	Radiative effects of low-level clouds over the subtropical SIO	Prescribed SST
CM_CTL	Active	—
CM_NoCRE	Inactive	—
AM_CTL	Active	Monthly climatology of CM_CTL
AM_NoCRE_SSTclim	Inactive	Monthly climatology of CM_CTL

30° increases progressively to reach $1/3^\circ$ at the equator. The atmospheric and oceanic components have 24 and 50 vertical levels, respectively.

We compare two types of CGCM experiments (Table 1) as performed by Miyamoto et al. (2021), who investigated radiative impacts of low-level clouds on the summertime Mascarene high. One is a fully coupled control experiment (CM_CTL) integrated for 100 years with a fixed radiative forcing for 1990. In the 50-yr climatology until November of the final year, the seasonality of the Mascarene high is well captured in the zonally asymmetric SLP (Fig. 1), although its intensity is slightly overestimated. CM_CTL also captures the seasonal characteristics of LCF over the subtropical SIO, such as a local maximum off the west coast of Australia in summer (Figs. 1a,e) and the basinwide enhancement toward winter (Figs. 1b,c,f,g). As described in section 1, this seasonal cycle of LCF is investigated in detail by Miyamoto et al. (2018). Consistently with the simulated LCF, CM_CTL reasonably reproduces the seasonality in TOA cloud radiative effect over the subtropical SIO (Fig. 1). However, the cooling effect over the subtropical SIO tends to be overestimated in the model throughout the year. Karlsson et al. (2008) showed that CM2.1 has the “too few, too bright” low-cloud problem like many other climate models (e.g., Nam et al. 2012).

The other experiment with CM2.1 is a “no low-cloud radiative effect” experiment (CM_NoCRE), where we set cloud fraction below the 680-hPa level to zero only for radiative transfer calculations in the model over the subtropical SIO (13° – 34°S , 30° – 120°E ; indicated with blue boxes in Figs. 1e–h). Although the cloud scheme is still active, simulated low-level clouds are substantially reduced in CM_NoCRE (Fig. 2). This is presumably because cloud-top radiative cooling is the main driver of convection that maintains boundary layer clouds (e.g., Lilly 1968; Wood 2012). Thus, this methodology effectively eliminates low-level clouds in the CGCM while adequately simulating the global hydrological cycle. CM_NoCRE was integrated for 70 years after branched off on 1 January of the 31st year in the CM_CTL simulation. In the following, the difference of CM_CTL from CM_NoCRE is referred to as an “anomaly” that represents a response of the model climate (i.e., 50-yr climatology until the last November) to radiative effects of low-level clouds. The statistical significance of the difference is assessed with the t test. As discussed in Miyamoto et al. (2021), climatic drift in either experiment is sufficiently small during the last 50 years.

FIG. 2. Annual-mean vertical profiles of model-level cloud fraction (%) simulated in the cloud scheme in CM_CTL (solid) and CM_NoCRE (dashed) averaged over the subtropical SIO where low-level clouds are made artificially transparent in the CM_NoCRE run. For clarity, the vertical axis is converted into the pressure coordinate with area-averaged surface pressure in CM_CTL.

c. Atmospheric general circulation model experiments

To extract the radiative impacts of low-level clouds without any changes in SST, we compare two atmospheric general circulation model (AGCM) experiments prescribed with the climatological-mean SST simulated in CM_CTL (Table 1). In addition to a 51-yr control experiment (AM_CTL), we run AM2.1, the atmospheric component of CM2.1, without radiative effects of low-level clouds as in CM_NoCRE (AM_NoCRE_SSTclim). As in the CGCM experiments, “anomaly” signifies the last 50-yr climatological difference of AM_CTL from AM_NoCRE_SSTclim, and the t test is used to assess its statistical significance.

d. Atmospheric dynamical model experiments

A linear baroclinic model (LBM) based on the primitive equations linearized about a specified basic state (Watanabe and Kimoto 2000) is utilized to investigate an atmospheric response within the framework of dry dynamics to diabatic heating and transient eddy fluxes. We use a version with T42 resolution in the horizontal and 20 σ levels in the vertical. The LBM is linearized about the wintertime (JJA) climatology of CM_CTL and forced with the wintertime anomalies in diabatic heating as well as anomalous convergence of heat and vorticity fluxes by submonthly eddies. An average over the last five days of a given 29-day integration is regarded as an equilibrium response to the forcing. Damping and diffusion introduced for the LBM are the same as in Miyamoto et al. (2021). Our conclusions are insensitive to the parameters for damping and diffusion (not shown).

3. Radiative impacts of low-level clouds on the Mascarene high simulated in CM2.1

a. Seasonal differences in the lowering of SST

We begin with the seasonal differences in SST anomalies induced by low-level clouds. Figure 3c shows summertime SST anomalies, while anomalies in net surface radiative flux in individual seasons are shown in Fig. 4. As described in Miyamoto et al. (2021), the negative SST anomalies are the strongest (around -4°C) in the eastern portion of the SIO in summer (Fig. 3c). These strong summertime anomalies are forced by anomalous radiative cooling in spring and summer (Figs. 4a,d),

FIG. 3. Difference (i.e., “anomaly” defined as CM_CTL – CM_NoCRE) in climatological-mean SST [shaded for every 1°C as indicated at the bottom of (c)] and surface winds (m s^{-1} ; arrows with reference in the bottom left; red and blue arrows signify increased and decreased scalar wind speed, respectively, with the 99% confidence for the difference) in (a) JJA and (c) DJF. Superimposed with green contours is climatological-mean SST (every 3°C) in CM_NoCRE with the 27°C isotherms thickened, whereas blue contours indicate the 27°C isotherms in CM_CTL. Stippling indicates the 99% confidence for the difference. (b),(d) As in (a) and (c), respectively, but for SLP [hPa; color shaded unevenly as indicated at the bottom of (d)]. Panels (c) and (d) are reproduced from Miyamoto et al. (2021) with minor modifications.

FIG. 4. Difference (i.e., “anomaly” defined as $CM_CTL - CM_NoCRE$) in the net surface radiative flux ($W\ m^{-2}$; positive values for downward flux) in (a) DJF, (b) MAM, (c) JJA, and (d) SON. The coloring convention is indicated at the bottom. Stippling indicates the 99% confidence for the difference.

which reflects the albedo effect of low-level clouds (Figs. S1a–h in the online supplemental material). A heat budget analysis for the ocean mixed layer confirms the predominance of the reduced shortwave heating for the spring-to-summer intensification of the negative SST anomalies (Miyamoto et al. 2021).

TABLE 2. Seasonal climatology of MLD (m) in MILA-GPV, CM_CTL, and CM_NoCRE for the subtropical SIO where low-level clouds are made artificially transparent in the CM_NoCRE run.

	MILA-GPV	CM_CTL	CM_NoCRE
DJF	25.9	22.5	20.8
JJA	60.4	81.6	72.8

In winter, the negative SST anomalies over the subtropical SIO are reduced to around $-2^{\circ}C$ with enhanced zonal uniformity (Fig. 3a). As shown in Fig. 4c, anomalous radiation induced by the albedo effect of low-level clouds still cools the subtropical SIO, acting to augment the zonal uniformity of the SST anomalies. However, the magnitude is substantially smaller than in spring and summer due mainly to the seasonally reduced insolation. Seasonally increased MLD in winter (Table 2) also acts to weaken their radiative impacts on the subtropical SST. Note that anomalies in middle- and high-cloud fractions (MCF and HCF, respectively) are mostly negative in all the seasons and thereby act to offset the shortwave cooling (not shown). Overall, the seasonality in the albedo effect of low-level clouds is consistent with the negative SST anomalies over the subtropical SIO.

The net anomalies in surface turbulent heat fluxes (Figs. S1i–p) tend to be downward where SST is lowered by the radiative forcing, as an indication of the thermodynamic feedback forcing by the SST response on the overlying atmosphere (Xie et al. 2010). This feedback forcing, which acts to damp the SST anomalies, prevails in the SIO in summer but weakens in winter due to the weaker SST anomalies. Atmospheric forcing on the anomalous turbulent heat flux also weakens in winter. In summer (Fig. 3c), the enhanced Mascarene high augments cold advection and scalar wind speed over the equatorward portion of the subtropical SIO between 15° and $20^{\circ}S$, while it acts conversely over its southwestern portion. This contributes to the stronger SST anomalies in the equatorward portion of the subtropical SIO, over which deep convection is strongly suppressed (section 3c; Miyamoto et al. 2021). Meanwhile, anomalous meridional Ekman transport induced by anomalous winds can advect warmer and cooler water toward its equatorward and poleward portions, respectively (Fig. S2), acting to reduce the inhomogeneity of the SST anomalies within the subtropical SIO. The corresponding atmospheric forcing weakens in winter (Fig. 3a), which is consistent with the enhanced uniformity of the SST anomalies. The following subsections discuss the weaker reinforcement of the Mascarene high in winter. Note that the reversal of climatological-mean wind direction into winter equatorward of $15^{\circ}S$ (Schott et al. 2009) results in increased wind speed by anomalous surface southeasterlies and hence equatorward expansion of the negative SST anomaly in winter (Fig. 3a).

b. Seasonal differences in the reinforcement of the Mascarene high by low-level clouds

The reinforcement of the Mascarene high by low-level clouds exhibits distinct seasonality. As shown in Fig. 3b, the wintertime SLP anomalies feature a high pressure anomaly

FIG. 5. Wintertime (JJA) difference (i.e., “anomaly” defined as $CM_CTL - CM_NoCRE$) in climatological-mean vertically integrated (a) Q_{precip} , (b) Q_{vdf} , and (c) Q_{rad} . Unit is $W\ m^{-2}$. (d)–(f) As in (a)–(c), respectively, but for summer (DJF). Stippling indicates the 99% confidence for the difference. The *uneven* coloring convention is indicated at the bottom. Black rectangular box indicates the domain where low-level clouds are made artificially transparent in the CM_NoCRE run. Note that (d)–(f) are reproduced from Miyamoto et al. (2021) with the different coloring convention.

centered at $20^{\circ}S$, $70^{\circ}E$. The induced positive SLP anomaly in Fig. 3b is located on the northeastern side of the climatological Mascarene high as the planetary-wave component centered at $35^{\circ}S$, $60^{\circ}E$ (Fig. 1g). The peak amplitude of the wintertime SLP anomaly is ~ 1.3 hPa, which is only 25% of its summertime counterpart (Fig. 3d). Interestingly, the center of the wintertime high-pressure anomaly is almost unchanged from its summertime counterpart, despite the westward expansion of LCF (Figs. 1a,c). In summary, the Mascarene high reinforcement by low-level clouds is much weaker in winter than in summer.

c. Indirect impacts of low-level clouds: Lowered SST

The aforementioned seasonality in the Mascarene high reinforcement is tied to seasonal differences in anomalous diabatic heating, especially condensation heating (large-scale condensation plus cumulus convection; Q_{precip}). Figures 5a and 5d show vertically integrated Q_{precip} anomalies for winter and summer, respectively, and the corresponding vertical profiles averaged over the subtropical SIO are shown in Figs. 6b and 6a. In summer, negative Q_{precip} anomalies are pronounced over the equatorward portion of the subtropical SIO (Fig. 5d). This Q_{precip} decrease is prominent in the free troposphere (Fig. 6a) and thus associated with suppressed deep convection. By inducing a Matsuno–Gill type response, the Q_{precip} decrease explains most of the summertime surface high pressure anomaly as verified through our LBM experiments (Miyamoto et al. 2021). Over the wintertime subtropical SIO, by contrast, negative Q_{precip} anomalies are much weaker (Fig. 5a), while the dipolar anomalies are more evident around the equator. The negative Q_{precip} anomalies above the 700-hPa level and near the surface are largely offset by anomalous Q_{precip} heating between the 700- and 900-hPa levels (Fig. 6b). The anomalous cooling in the upper and middle troposphere reflects suppressed deep convective clouds, whereas the anomalous heating in the lower troposphere is due to the increased occurrence of low-level clouds.

The anomalous Q_{precip} impacts on the wintertime Mascarene high are quantified with the LBM experiments. As shown in Fig. 7a, the LBM can reproduce the surface anticyclonic anomaly over the wintertime subtropical SIO simulated in the CGCM experiments (Fig. 3b). This anticyclonic response in the LBM is explained mostly by the diabatic heating anomalies within the subtropical SIO (Fig. 7b). Decomposition of the anticyclonic response into the individual contributions from condensation (Q_{precip}), radiation (Q_{rad}), and vertical diffusion (Q_{vdf}) anomalies over the subtropical SIO reveals that the Q_{precip} anomalies make no substantial contribution to the total response within the subtropical basin (Fig. 8a). Likewise, the impacts of the dipolar anomalies in Q_{precip} around the equator are also negligible (not shown). Thus, precipitation changes induced by low-level clouds leave no substantial net impacts on the surface Mascarene high in winter, which makes a sharp contrast to the summertime situation.

One of the reasons for the lack of distinct precipitation anomalies within the subtropical SIO is lower climatological-mean SST in winter than in summer, which is realistically simulated in CM2.1 (Fig. S4). As in Miyamoto et al. (2021), we invoke the SST threshold for active convection (e.g., Graham and Barnett 1987; Johnson and Xie 2010), which is around $27^{\circ}C$ in either experiment ($26.8^{\circ}C$ in CM_CTL and $27.2^{\circ}C$ in CM_NoCRE ; Fig. S3). Combined with the weaker negative SST anomalies in winter (Figs. 3a,c), isotherms of the convective threshold lie equatorward of $15^{\circ}S$ even in CM_NoCRE (Fig. 3a). In summer, by contrast, the threshold isotherms shift from 20° – $25^{\circ}S$ in CM_NoCRE to 15° – $20^{\circ}S$ in CM_CTL (Fig. 3c). This prevents the ITCZ from shifting poleward in CM_CTL , leading to the substantial decrease in precipitation over the summertime subtropical SIO (Fig. 5d). Another reason is seasonally enhanced midtropospheric descent in winter associated with the Hadley circulation (e.g., Lindzen and Hou 1988; Dima and Wallace 2003). As confirmed in Fig. 9, upper-tropospheric convergence and associated midtropospheric subsidence are stronger in the winter hemisphere

FIG. 6. (a) Climatological-mean vertical profiles of summertime (DJF) model-level diabatic heating ($K day^{-1}$) in CM_CTL (solid) and CM_NoCRE (dashed) averaged over the subtropical SIO where low-level clouds are made artificially transparent in the CM_NoCRE run. Red, green, and blue lines indicate Q_{precip} , Q_{vdf} , and Q_{rad} , respectively. (b) As in (a), but for winter (JJA). (c) As in (b), but for AM_CTL (solid) and AM_NoCRE_SSTclim (dashed). (d) As in (b), but for difference in lower-tropospheric Q_{vdf} ($K day^{-1}$) in the CCGM (CM_CTL - CM_NoCRE; purple line) and AGCM (AM_CTL - AM_NoCRE_SSTclim; brown line) experiments. For clarity, the vertical axis is converted into the pressure coordinate with area-averaged surface pressure in (a),(b),(d) CM_CTL and (c) AM_CTL.

than in the summer hemisphere in both CM_NoCRE and CM_CTL, acting to suppress deep convection in the wintertime subtropics. Therefore, the impacts of low-level clouds on precipitation over the wintertime subtropical SIO are weaker than in summer.

The lowered SST can also influence the Mascarene high through Q_{vdf} , which mainly represents turbulent heating within the planetary boundary layer related to surface sensible heating (the opposite sign of Figs. S1e–h). Figure 5b shows wintertime anomalies of vertically integrated Q_{vdf} . Consistent with

FIG. 7. (a) Surface pressure response (hPa) in the LBM experiment forced with the wintertime (JJA) differences (defined as CM_CTL – CM_NoCRE) in diabatic heating and flux convergence of heat and vorticity by submonthly eddies. (b) As in (a), but for the response solely to anomalous diabatic heating in the subtropical SIO. The coloring convention is indicated at the bottom.

the lowered SST (Fig. 3a), anomalous cooling is evident over the wintertime subtropical SIO. It declines toward the equator, probably because the enhanced cold advection brought about by the stronger Mascarene high acts to increase sensible heat release from the ocean (Fig. 3a). Compared with the summertime counterpart (Fig. 5e), the anomalous cooling is also weaker in winter in accordance with the smaller SST decrease (Figs. 3a,c). The LBM experiments described above show that the wintertime negative Q_{vdf} anomalies only slightly reinforce the high (Fig. 8b).

Overall, the impacts of the lowered SST on the Mascarene high are weaker in winter than in summer. Particularly, the suppression of deep convection under the lowered SST becomes so weak in winter that its impacts are roughly offset by anomalous condensation heating associated with the increased low-level clouds. This leads to the substantial weakening of the feedback from low-level clouds onto the wintertime Mascarene high in winter.

d. Direct impacts of low-level clouds: In-atmosphere radiative cooling

Low-level clouds can reinforce the Mascarene high directly by strengthening in-atmosphere radiative cooling from their tops (e.g., Miyasaka and Nakamura 2005, 2010; Miyamoto et al. 2021). As shown in Fig. 5c, a negative anomaly in vertically integrated Q_{rad} is distributed across the wintertime subtropical SIO, accounting for $\sim 10\%$ of the climatological value in CM_NoCRE. As confirmed by the LBM experiments, this enhanced radiative cooling in the subtropical SIO induces an anticyclonic response at the surface (Fig. 8c). The magnitude of the response (~ 1 hPa) is comparable to that induced by the summertime Q_{rad} anomalies [Figs. 5f and 6a; see Miyamoto

et al. (2021) for the summertime LBM response], with the center of the wintertime response shifted slightly westward in accordance with the westward expansion of the anomalous radiative cooling.

The enhanced radiative cooling is related to anomalous cloud cover (Fig. 10). As discussed by Miyamoto et al. (2021), high-level clouds (cloud-top pressure ≤ 440 hPa) that act to warm the underlying atmosphere by downward longwave radiation decreases in summer in association with reduced deep convective precipitation (Fig. 5d). The reduced HCF enhances radiative cooling over the equatorward portion of the subtropical SIO, in addition to the enhanced radiative cooling induced by the imposed increase in low-level clouds (Fig. 1e). By contrast, wintertime reduction of HCF is modest (mostly less than 3%) in the subtropical SIO (Fig. 10b), consistent with the weak Q_{precip} anomalies in winter (Fig. 5a). Meanwhile, positive LCF anomalies extend across the subtropical SIO in winter (Fig. 10a; Miyamoto et al. 2018), acting to induce the basinwide enhancement of radiative cooling. In fact, anomalous Q_{rad} concentrates in the lower troposphere below the 680-hPa level, where the cloud radiative effect is eliminated in CM_NoCRE (Fig. 6b). Thus, owing to the basinwide increase of low-level clouds, their radiative cooling acts to reinforce the wintertime Mascarene high with a comparable magnitude to the summertime counterpart.

4. Discussion: Does the lowered SST by low-level clouds matter for the wintertime reinforcement of the Mascarene high?

Our LBM experiments suggest that radiative cooling from low-cloud tops accounts for most of the total response of the

FIG. 8. As in Fig. 7b, but for individual contributions from (a) Q_{precip} , (b) Q_{vdf} , and (c) Q_{rad} .

FIG. 9. Meridional profiles of zonal-mean horizontal divergence (10^{-6} s^{-1}) in the upper troposphere averaged between the 300- and 100-hPa levels. Blue and red solid lines indicate wintertime (JJA) and summertime (DJF) climatologies in CM_CTL, respectively, whereas dashed lines indicate their counterpart in CM_NoCRE.

wintertime Mascarene high (Fig. 8c). However, this result may not necessarily degrade the importance of the lowered SST in the low-cloud feedback onto the wintertime Mascarene high. This is because the decreased Q_{precip} due to the suppressed deep convection and the decreased low-level Q_{vdf} may act to offset the increased Q_{precip} due to boundary layer clouds (Fig. 6b). To clarify this offset, we compare AM_CTL with AM_NoCRE_SSTclim, in which their radiative impacts through SST are inhibited. Figure 11 shows SLP anomalies in the AGCM experiments. The reinforcement of the wintertime Mascarene high is statistically significant, but its magnitude is about 0.4 hPa, accounting only for 30% of its CGCM counterpart (about 1.3 hPa).

Figure 12 compares vertically integrated anomalies of the individual diabatic heating components in the AGCM experiments, while the corresponding vertical profiles are shown in Fig. 6c. The anomalous cooling by Q_{rad} in the AGCM experiments (Figs. 6c and 12c) is almost the same as in the CGCM experiments (Figs. 5c and 6b). Unlike in the CGCM experiments (Figs. 5b and 6b,d), the AGCM experiments do not simulate anomalous cooling by Q_{vdf} in the lower troposphere under the fixed SST (Figs. 6c,d and 12b). Furthermore, statistically significant positive Q_{precip} anomalies are distributed across the subtropical SIO (Fig. 12a), in contrast to the slight cooling anomalies in the CGCM experiments (Fig. 5a). This

is because the Q_{precip} reduction above the 680-hPa level simulated in the CGCM (Fig. 6b) vanishes in the AGCM experiments (Fig. 6c). As a result, despite the slight offset by evaporative cooling near the surface, the Q_{precip} increase below the 680-hPa level due to the increased low-level clouds partially offsets the Q_{rad} effects on the surface high reinforcement in the AGCM experiments. Similar cancellation in reinforcing subtropical highs has recently been reported by Kawai and Koshiro (2020). Thus, our AGCM experiments suggest that the lowered SST by low-level clouds is still important in reinforcing the wintertime Mascarene high.

It should be noted that the wintertime cloud radiative effect over the subtropical SIO simulated in CM_CTL is stronger than observed (Figs. 1c,g), although the positive bias in MLD (Table 2) acts to mitigate this excessive cooling effect. Multilinear regression analysis on low-cloud reflectivity (Brient and Schneider 2016) supports the overestimated cooling by low-level clouds in winter (see text S1 in the supplemental material). Thus, the lowered SST and its feedback onto the atmosphere in winter may be overestimated in CM2.1. The wintertime relative importance between the lowered SST and the direct longwave cooling by low-level clouds may therefore include some uncertainties.

5. Concluding remarks

Extending our recent work that demonstrated radiative impacts of low-level clouds on the summertime Mascarene high with a CGCM (Miyamoto et al. 2021), the present study has assessed their radiative impacts on the wintertime Mascarene high to reveal the seasonality. The comparison of a fully coupled control run (CM_CTL) with another run (CM_NoCRE) where the radiative effects of low-level clouds are artificially switched off has demonstrated that the low-cloud effect on the Mascarene high formation is substantially weaker in winter than in summer.

For the seasonality in the reinforcement, the background climatology is found important. In winter, the suppression of deep convection by low-level clouds that acts to reinforce the high is much weaker than in summer. This arises from 1) seasonally lower SST in concert with weaker negative SST anomalies

FIG. 10. Wintertime (JJA) difference (i.e., “anomaly” defined as $\text{CM_CTL} - \text{CM_NoCRE}$) in climatological-mean (a) LCF (%) and (b) HCF (%). Stippling indicates the 99% confidence for the difference. We employ the same coloring convention as in Miyamoto et al. (2021), which is indicated at the bottom of each panel. Black rectangular box indicates the domain where low-level clouds are made artificially transparent in the CM_NoCRE run.

FIG. 11. As in Fig. 3b, but for the difference in the AGCM experiments defined as AM_CTL – AM_NoCRE_SSTclim.

due to the thicker MLD and the weaker cloud radiative effect under smaller insolation and 2) stronger subtropical subsidence associated with the Hadley circulation in winter. By contrast, the cloud-top radiative cooling by low-level clouds, which acts to reinforce the Mascarene high, is comparable between winter and summer, as verified through atmospheric dynamical model experiments.

While it is now established that the summertime self-sustaining system of the Mascarene high and low-level clouds is driven by the seasonally heated Australian continent and associated land–sea thermal contrasts across its west coast (Miyasaka and Nakamura 2010; Miyamoto et al. 2021), the maintenance mechanisms for the wintertime high is yet to be fully understood. As discussed in Miyamoto et al. (2018), seasonally enhanced storm-track activity in the SIO might externally reinforce the high by increasing low-level clouds. However, the present study has demonstrated that the wintertime feedback from low-level clouds onto the high (~ 1.3 hPa; Fig. 3b) cannot fully explain the strength of the high as the planetary-wave component (~ 6 hPa; Figs. 1c,g). Hence, another internal feedback pathway and/or external driver is presumably needed to replenish the full strength of the wintertime Mascarene high. Indeed, previous studies suggested orographic forcing (Rodwell and Hoskins 2001) and cross-equatorial teleconnection (Lee et al. 2013). Our subsequent paper (Miyamoto et al. 2021, manuscript submitted to *J. Climate*) will assess the importance of those external modulations that strengthen and shift the wintertime high.

We have indicated that the background climatology plays an essential role in the wintertime weakening of the low-cloud feedback on the subtropical Mascarene high. In other words,

the summertime low-cloud feedback is augmented by the background climatology. In summer, climatologically high SST and weak subsidence of the Hadley circulation both favor deep convection, and therefore its suppression by low-level clouds acts as their major feedback onto the summertime Mascarene high. The importance of the background climatology also provides insight into interbasin differences in the low-cloud feedback. Especially in summer, the high SST in the subtropical SIO compared with the other eastern subtropical oceans (e.g., Tozuka and Oettli 2018) acts to amplify the low-cloud feedback, despite the relatively modest low-cloud cover over the SIO (e.g., Klein and Hartmann 1993; Wood and Bretherton 2006; Wood 2012; Koshiro and Shiotani 2014; Miyamoto et al. 2018). Over the Pacific and Atlantic, the prevalence of the WES mode may invoke another feedback pathway via changes in the equatorial oceans (e.g., Philander et al. 1996; Ma et al. 1996; Yu and Mechoso 1999; Gordon et al. 2000; Xie et al. 2007; Bellomo et al. 2014, 2015; Myers et al. 2018), which is unlikely over the Indian Ocean as mentioned in section 1. The interbasin differences of the low-cloud feedback will be investigated in our future work.

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Data availability statement. All the observational data used in this study are available from the corresponding websites (ERA5: <https://www.ecmwf.int/en/forecasts/datasets/reanalysis-datasets/era5>; ISCCP: <https://climserv.ipsl.polytechnique.fr/cfmip-obs/>; CERES-EBAF: <https://ceres.larc.nasa.gov/data/>; MILA-GPV: http://www.jamstec.go.jp/ARGO/argo_web/argo/?page_id=223&lang=en).

FIG. 12. As in Figs. 5a–c, but for the difference in the AGCM experiments defined as AM_CTL – AM_NoCRE_SSTclim.

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