Maintenance Mechanisms of the Wintertime Subtropical High over the South Indian Ocean

AYUMU MIYAMOTO, a HISASHI NAKAMURA, a, b TAKAFUMI MIYASAKA, a YU KOSAKA, a BUNMEI TAGUCHI, c and KAZUAKI NISHI d

a Research Center for Advanced Science and Technology, The University of Tokyo, Tokyo, Japan
b Japan Agency for Marine-Earth Science and Technology, Yokohama, Japan
c Faculty of Sustainable Design, University of Toyama, Toyama, Japan
d Graduate School of Bioresources, Mie University, Tsu, Japan

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ABSTRACT: Climatologically the surface Mascarene high over the subtropical south Indian Ocean (SIO) shifts westward toward austral winter, and its strength as a planetary-wave component maximizes in late austral winter, unlike its counterpart over other subtropical oceans. The present study investigates the maintenance mechanisms for the wintertime Mascarene high with a linear atmospheric dynamical model (LBM) and an atmospheric general circulation model (AGCM). The LBM experiments reveal the importance of cross-equatorial tropical influences. Deep convection associated with the Asian summer monsoon acts not only to shift the Mascarene high westward as its direct influence but also to enhance midtropospheric subsidence and equatorward surface winds over the central and western portions of the subtropical SIO. The associated near-surface cold advection and subsidence promote (suppress) the formation of low-level (deep convective) clouds. The resultant enhanced radiative cooling and reduced deep condensation heating both reinforce the equatorward portion of the surface high. The LBM experiments also reveal that seasonally enhanced storm-track activity over the SIO is important for maintaining the poleward portion of the Mascarene high through eddy heat and vorticity fluxes. The AGCM experiments demonstrate that the Agulhas Current system and the associated sea surface temperature (SST) front reinforce the high by energizing the storm-track activity. The present study thus proposes that both the Asian summer monsoon and the enhanced storm-track activity maintained by the Agulhas SST front externally modulate the positively coupled system between the wintertime Mascarene high and low-level clouds to realize its unique seasonality.

KEYWORDS: Asia; Indian Ocean; Clouds; Teleconnections; Boundary currents; Monsoons; Storm tracks; Subtropical cyclones

1. Introduction

In the subtropics, the zonal-mean Hadley cell causes descent to form a subtropical surface high pressure belt (Held and Hou 1980; Lindzen and Hou 1988; Dima and Wallace 2003). As inferred from a planetary-wave (i.e., zonally asymmetric) component of sea level pressure (SLP*; hereafter a zonally asymmetric component is denoted with an asterisk; Figs. 1a, b), SLP is particularly high over the eastern subtropical oceans, manifested as a semipermanent subtropical high. The subtropical high characterizes the surface winds over each of the subtropical oceans and therefore plays an integral role in shaping the weather and climatic conditions over the ocean as well as its surrounding land.

Although the zonal-mean descent associated with the Hadley cell is stronger in local winter (i.e., JJA in the Southern Hemisphere and DJF in the Northern Hemisphere; Lindzen and Hou 1988; Dima and Wallace 2003), the subtropical highs tend to be stronger in local summer due to the predominance of their planetary-wave component (Figs. 1a, b). Many studies therefore focused on the summertime highs and suggested that local air–sea interaction is essential in their maintenance. Surface equatorward winds over the eastern subtropical oceans bring cold-air advection as well as coastal upwelling and offshore Ekman transport. In conjunction with midtropospheric subsidence, the induced low sea surface temperature (SST) suppresses deep convective heating and facilitates radiative cooling, thereby reinforcing the subtropical highs (Rodwell and Hoskins 2001; Seager et al. 2003; Liu et al. 2004; Miyasaka and Nakamura 2005, 2010; Miyamoto et al. 2021a). These equatorward winds and subsidence are suggested to be brought about by monsoonal deep convective heating (Rodwell and Hoskins 2001) and shallow land–sea thermal contrast across the west...
The present study explores the maintenance mechanisms of the wintertime Mascarene high, which have not been fully elucidated thus far, and aims at presenting an integrated view of the coupled system between the Mascarene high and low-level clouds, as summarized in Fig. 2. In elucidating the maintenance mechanisms, one of our specific focuses is on the remote influence from the tropics (green arrow in Fig. 2a), including the Asian summer monsoon (ASM; Wang and Fan 1999). Associated with the ASM, a large amount of condensation heating can induce circulation anomalies as a Matsuno–Gill-type response (Matsuno 1966; Gill 1980; Kraucunas and Hartmann 2007) that may affect the wintertime Mascarene high. Through atmospheric general circulation model (AGCM)
experiments, Lee et al. (2013) argued that the summer monsoons in the Northern Hemisphere can reinforce the Southern Hemisphere subtropical highs in austral winter. In their simulations, however, the northern summer monsoons act to weaken the Mascarene high around its center. It may be attributable to seasonally enhanced near-surface baroclinicity associated with increasing SST gradient and reduced static stability rather than to upper-tropospheric westerlies (Inatsu and Hoskins 2004; Nakamura and Shimpo 2004). Although the reinforcement of the wintertime Mascarene high by low-level clouds is found to be modest (Miyamoto et al. 2021b), Miyamoto et al. (2018) argued that the enhanced storm-track activity promotes the formation of low-level clouds by augmenting climatological-mean cold advection and scalar wind speed (blue dashed arrow in Fig. 2a), and thereby potentially drives the positive feedback system with the Mascarene high (red and purple dashed arrows in Fig. 2a). In addition to this mechanism, baroclinic eddies can dynamically force persistent circulation anomalies through eddy heat and vorticity fluxes (e.g., Lau and Nath 1991; Held et al. 2002). This study verifies that the storm-track activity over the SIO dynamically reinforces the wintertime Mascarene high (solid blue arrows in Fig. 2a).

The rest of the paper is organized as follows. Section 2 describes data and model experiments. Section 3 examines the three-dimensional structure of the wintertime Mascarene high. Through atmospheric dynamical model and AGCM experiments, sections 4 and 5 discuss external modulations of the high by the remote influence from the tropics (green arrow in Fig. 2a) and by the storm-track activity and Agulhas SST front over the SIO (solid blue arrow from “Agulhas SST front” and “Storm track” to “Subtropical high” in Fig. 2a), respectively. Section 6 provides a summary and discussion.

2. Data and model experiments

a. Observational data

Our observational analysis is based on the Japanese 55-year Reanalysis (JRA-55; Kobayashi et al. 2015; Harada et al. 2016) from 1979 through 2018. For cloud fraction and individual contributions to diabatic heating from radiation ($Q_{rad}$), vertical diffusion ($Q_{diff}$), and condensation (large-scale condensation plus cumulus convection; $Q_{precip}$), we use their model-level fields rather than their pressure-level data to better depict their near-surface structures. Vertical integration of the heating from the surface to the model-top ($\sim 0.1 \text{ hPa}$) is calculated with mass weight. The JRA-55 cloud fraction is compared with satellite observations from the GCM-Oriented Cloud–Aerosol Lidar and Infrared Pathfinder Satellite Observations (CALIPSO) Cloud Product (GOCCP) version 3 (Chepfer et al. 2010) from June 2006 through May 2017.

![Diagram](http://example.com/diagram.png)

**FIG. 2.** (a) Schematic diagram illustrating a feedback system associated with the subtropical Mascarene high and low-level clouds over the SIO in austral winter. Solid arrows indicate influences demonstrated by the present study, while dashed arrows indicate those demonstrated by previous studies. Light blue boxes signify the components of the feedback system, while light green boxes signify external agents that can modulate the feedback. (b) Geographical locations of the individual factors in (a). “H” signifies the subtropical high, whereas low-level clouds are signified in gray. Green arrows indicate upper-tropospheric divergent wind from the tropics.
b. Atmospheric dynamical model experiments

As shown in many studies (e.g., Held et al. 2002, and references therein), a linear baroclinic model (LBM) based on the primitive equations on the sphere is a useful tool for understanding the dynamical processes involved in the formation of planetary waves. Following those studies, we use a particular LBM formulated by Watanabe and Kimoto (2000, 2001). For our purpose, the model is linearized about a zonal-mean basic state to obtain a response of the dry atmosphere to zonally asymmetric diabatic heating as well as submonthly eddy heat and vorticity fluxes (derived from deviations from monthly averages). The linearity of the LBM allows us to decompose the total response into the contributions of individual forcing components.

This study employs the same methodology as in Miyamoto et al. (2021a), who investigated the planetary-wave component of the summertime (January) Mascarene high. The following texts of this paragraph are therefore derived from the paper with minor modifications. We set a resolution of T42 with 20 vertical levels. Damping and diffusion introduced into the LBM are the same as in Miyamoto et al. (2021a). Our conclusions are found insensitive to the parameters for damping and diffusion (not shown). The basic state and forcing for each calendar month are evaluated from the JRA-55 climatologies. It should be noted that the diabatic heating prescribed to the LBM is a product of the forecast model used for JRA-55 and we thus not strongly constrained by assimilated observational data. Nevertheless, our conclusions do not change if we use other reanalysis datasets such as the Japanese 25-year Reanalysis (JRA-25; Onogi et al. 2007) and the National Centers for Environmental Prediction (NCEP) Reanalyses 2 (NCEP-R2; Kanamitsu et al. 2002), which provide the individual diabatic heating components (not shown). The LBM was integrated for 29 days with a prescribed forcing to obtain an equilibrium response as the last 5-day average.

c. AGCM experiments

An AGCM for the Earth Simulator (AFES; Ohfuchi et al. 2004, 2007; Enomoto et al. 2008; Kuwano-Yoshida et al. 2010) is utilized for extracting the impacts of the Agulhas SST front. The horizontal resolution is configured at T119 (equivalently 1° grid intervals), with 56 vertical levels from the surface to approximately 0.1 hPa (Nishii et al. 2020). Daily global SST fields taken from the Optimally Interpolated Advanced Very High Resolution Radiometer (AVHRR) Pathfinder SST (OISST) data produced by the National Oceanic and Atmospheric Administration (NOAA) with 0.25° resolution (Reynolds et al. 2007) are prescribed at the bottom boundary of AFES.

We conducted the following set of experiments: the control experiment (AFES_CTL) with the original OISST and the sensitivity experiment (AFES_SMTH) with the spatially smoothed OISST. The smoothed SST fields were created with a Gaussian-type smoothing filter as follows. At each grid point \((x, y)\), SST was horizontally averaged with Gaussian weighting \(W_{k,l}\):

\[
W_{k,l} = \exp \left\{ -\frac{(x_{k,l} - x)^2 + (y_{k,l} - y)^2}{a^2} \right\},
\]

where \(k\) and \(l\) are longitudinal and latitudinal grid indices, respectively, of the SST field, and \(a\) is set to 600 km. This procedure was performed over the extratropics (30.125°–55.125° of latitude in each hemisphere) with ±5° latitudinal buffer zones. As shown in Fig. 3, the smoothing reduces the prominent meridional SST gradient across the Agulhas SST front around 40°–45°S. In both experiments, AFES was integrated for 32 years from 1982 through 2013. Each of the experiments has 15 ensemble members, all of which are used to calculate the climatological-mean fields.

3. Three-dimensional structure of the wintertime Mascarene high

a. Observations

This section elucidates the three-dimensional structure of the wintertime Mascarene high in August, when the surface high as the planetary-wave component is the strongest in climatology (Figs. 1b,c). As discussed in section 1, the surface high forms in austral winter under the descending branch of the seasonally enhanced Hadley cell (Figs. 4a,b; Lindzen and Hou 1988; Dima and Wallace 2003). More importantly, the
Mascarene high peaks in late winter even as the planetary-wave component, which is centered at 35°S, 60°E (Figs. 1b,c), to the west of its summertime counterpart (30°S, 90°E; Figs. 1a,c). Concomitant with the wintertime enhancement and westward shift of the surface high, surface southerlies prevail over the SIO equatorward of 35°S (Fig. 4a). Cyclonic planetary vorticity advected by the enhanced southerlies is balanced with vortex-tube shrinking by horizontal divergence near the surface (the Sverdrup balance; Rodwell and Hoskins 2001), and thus dynamically consistent with basinwide midtropospheric subsidence (Fig. 5a).

The basinwide midtropospheric subsidence is linked to upper-tropospheric circulation. A meridional section of relative vorticity for the western SIO reveals a meridional vorticity dipole peaking at the 300-hPa level above the surface high, with cyclonic and anticyclonic lobes around 28° and 40°S, respectively (Fig. 6). This kind of dipolar structure is typically observed for the summertime subtropical highs (Miyasaka and Nakamura 2005, 2010). The meridional dipole is also hinted at in a horizontal map of 200-hPa vorticity, in association with the northwest–southeast tilt of the node that constitutes a zonal dipole over the poleward portion of the surface high as well (Fig. 5b). Across this zonal gradient of vorticity, the upper-tropospheric zonal-mean westerlies (Fig. 6) yield anticyclonic vorticity advection, which balances with vortex-tube stretching by upper-tropospheric horizontal convergence (Figs. 4b and 6) associated with the midtropospheric subsidence (Fig. 5a). In addition, the stretching effect is also offset by the absolute vorticity advection by poleward divergent winds (Figs. 4b and 6) that takes place slightly equatorward of the subtropical jet axis located at 30°S.

A striking difference from the summertime situation is strong anticyclonic vorticity around 20°S east of 60°E (Fig. 5b). It accompanies prominent poleward wave-activity flux in the upper troposphere (red arrows in Fig. 5b), suggestive of remote influences from the tropics. This anticyclonic vorticity, together with the meridional vorticity dipole discussed above, corresponds to the subtropical-jet entrance and exit at 30°S and the subpolar-jet exit in the SIO (Fig. 4c; Nakamura and Shimpo 2004). At these jet entrance and exit, poleward and

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**FIG. 4.** Climatological-mean August distributions of (a) SLP (color shaded for every 4 hPa) and surface winds (arrows), (b) 200-hPa horizontal divergence (color shaded for every 10⁻⁶ s⁻¹) and divergent winds (arrows), and (c) 200-hPa zonal wind (color shaded for every 7 m) and ageostrophic winds (arrows) based on JRA-55. The reference for the arrows is indicated on the right of (c).

**FIG. 5.** As in Fig. 4, but for zonally asymmetric fields of (a) 700-hPa 𝜔 (hPa day⁻¹); the uneven coloring convention as indicated at the bottom), (b) 200-hPa relative vorticity (contoured for ±2, ±6, ±10, ... × 10⁻⁶ s⁻¹; solid and dashed lines for anticyclonic and cyclonic polarities, respectively), (c) variance of 250-hPa meridional wind by submonthly eddies (\(\mathbf{V^2} \mathbf{V}^*\); color shaded for every 20 m² s⁻²), and (d) 925-hPa poleward heat flux by submonthly eddies (\(-\mathbf{V^T} \mathbf{T}^*\); color shaded for every 2 K m s⁻¹). In (b), superimposed with red arrows is Plumb’s (1985) wave-activity flux (m² s⁻²; reference on the right), while shading is applied where the zonally asymmetric SLP exceeds 2 hPa. Blue dashed lines in (b) designate the 60° and 80°E meridians, between which we take the meridional section in Fig. 6. In (c), gray lines are 32 m s⁻¹ isolines of 250-hPa total westerly wind speed. In (d), green hatches indicate where the meridional SST gradient is greater than 1.4°C (100 km)⁻¹.
equatorward ageostrophic winds $u_y$ are required, respectively, for the momentum balance of zonal geostrophic winds $u_g$ (i.e., $u_g (\partial u_g / \partial x) \approx f v_y$ with the Coriolis parameter $f$; Fig. 4c), as discussed for the summertime high (Miyasaka and Nakamura 2010). These ageostrophic winds over the SIO are consistent with the enhanced upper-level convergence (Fig. 4b) and subsidence underneath (Fig. 5a). Thus, the upper-tropospheric circulations are dynamically linked with the midtropospheric subsidence and surface Mascarene high.

b. Reproducibility of the wintertime Mascarene high in the LBM experiments

Before discussing details based on the LBM experiments, we examine the validity of our LBM in reproducing the Mascarene high. Figures 1d and 1e show January and August surface pressure responses, respectively, in the LBM experiments where both diabatic heating and submonthly eddy forcing (convergence of heat and vorticity fluxes) are imposed north of 65°S (LBM_GLB). No forcing is imposed around Antarctica to avoid an unrealistically large response over the steep orography. Compared with their observational counterparts (Figs. 1a,b), the LBM reproduces the maritime high reasonably well, despite the slight underestimation of its westward shift toward winter. As shown in Fig. 7a, the basinwide enhancement of subsidence over the wintertime subtropical SIO is also well captured.

Figure 8c shows the August response of upper-tropospheric streamfunction in LBM_GLB. Compared with the observational counterpart (Fig. 8a), the anticyclone centered at 15°S equatorward of the subtropical jet is well reproduced, though slightly displaced equatorward. This feature can also be seen in the vorticity field, accompanied by a cyclone to the south (Fig. 7b). However, another anticyclone southwest of Madagascar that constitutes the meridional vorticity dipole over the western SIO is not evident in LBM_GLB. This is likely related to the absence of Rossby-wave propagation from the subpolar South Atlantic (Figs. 7b and 8c). This wave train with a cyclone southeast of the anticyclone tends to be

FIG. 6. Meridional section of the zonally asymmetric component of climatological-mean relative vorticity (contoured in black for $\pm 1, \pm 3, \pm 5, \ldots \times 10^{-6}$ s$^{-1}$; solid and dashed lines for anticyclonic and cyclonic polarities, respectively) averaged between 60° and 80°E for August. Superimposed with red contours is zonal-mean zonal wind (contoured for every 10 m). Blue arrows indicate climatological-mean meridional component of ageostrophic wind (m s$^{-1}$) and $\omega$ velocity (hPa day$^{-1}$) with reference on the right.

FIG. 7. LBM responses of (left) 700-hPa $\omega$ (hPa day$^{-1}$; reference at the bottom) and (right) 200-hPa relative vorticity (contoured for $\pm 2, \pm 6, \pm 10, \ldots \times 10^{-6}$ s$^{-1}$; solid and dashed lines for anticyclonic and cyclonic polarities, respectively). Shown are the responses to (a),(b) all forcing in LBM_GLB and to (c),(d) diabatic heating and (e),(f) submonthly eddy forcing in LBM_SIO. In (b), (d), and (f), shaded areas are where the surface pressure response exceeds 4 hPa, whereas red arrows indicate Plumb's (1985) wave-activity flux (m$^2$ s$^{-2}$; reference on the right).
equivalent-barotropic (Figs. 1b and 8a) and therefore it might act to shift the surface Mascarene high westward. A hemispheric map of 200-hPa streamfunction based on JRA-55 reveals that this wave train originates from the Antarctic coast in the South Atlantic (Fig. 8b). Thus, the absence of the thermal, submonthly eddy and orographic forcing around Antarctica might contribute to the eastward displacement of the wintertime surface high in LBM_GLB. Otherwise, the LBM seems to capture the processes responsible for the maintenance of the wintertime surface Mascarene high.

4. Reinforcement of the Mascarene high by local diabatic cooling and its external modulation by the remote influences from the tropics

a. Impacts of diabatic cooling over the SIO

This section demonstrates that remote influences from the tropics can drive local self-sustaining feedback between the wintertime Mascarene high and diabatic processes over the SIO. We begin with the local diabatic processes over the SIO based on JRA-55. Figure 9 shows maps of the August diabatic heating after being integrated vertically, and the corresponding zonal sections across the subtropical SIO are shown in Figs. 10a–c. Unlike in summer, a shallow land–sea $Q^\text{rad}$ contrast across the west coast of Australia vanishes in austral winter under reduced insolation (Figs. 9b and 10a). Since the land–sea thermal contrast is essential for anchoring the subtropical highs over the eastern subtropical oceans (Miyasaka and Nakamura 2005, 2010; Miyamoto et al. 2021a), the diminished land–sea thermal contrast and associated atmosphere–ocean–land feedback system are consistent with the westward shift of the Mascarene high observed in austral winter.

The vertically integrated wintertime $Q^\text{rad}$ is characterized by a cooling anomaly ($\sim 15$ W m$^{-2}$) spreading almost entirely over the subtropical SIO (Fig. 9c). The cooling maximizes in the lower troposphere above a weak near-surface heating anomaly (Fig. 10b). As shown in Fig. 10d, maximum cloud fraction in JRA-55 is found between these cooling and heating, suggesting that low-level clouds exert strong longwave cooling at their top and weak heating in the subcloud layer (Lilly 1968). Although quantitative comparison is impossible owing to their different vertical resolutions, the cloud distribution in JRA-55 seems to correspond well with satellite observations (Fig. 10e).

The absence of high-top clouds (Figs. 10d,e) due to suppressed deep convection further augments longwave cooling underneath. Under the enhanced subsidence (Figs. 5a and 6), vertically integrated $Q^\text{precip}_r$ also exhibits anomalous cooling ($\sim 40$ W m$^{-2}$) over the subtropical SIO (Fig. 9d). Except within the low-cloud layer, the cooling (or suppressed heating) is observed throughout the troposphere over the subtropical SIO (Fig. 10c). The $Q^\text{rad}$ and $Q^\text{precip}_r$ cooling acts to reinforce the surface high (Matsumo 1966; Gill 1980).

Figure 11a shows the surface pressure response in the LBM to diabatic heating/cooling around the SIO (20°–120°E, 15°–65°S, as indicated by black rectangles in Fig. 9; hereafter the experiment with forcing prescribed only within this domain is referred to as LBM_SIO). Consistent with the local $Q^\text{rad}$ and $Q^\text{precip}_r$ cooling, there is a high pressure response over the subtropical SIO, accounting for the

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**FIG. 8.** (a) Climatological-mean August distributions of zonally asymmetric 200-hPa streamfunction ($\times 10^6$ m$^2$ s$^{-1}$) around the Indian Ocean and (b) from the hemispheric view. (c) The 200-hPa streamfunction response ($\times 10^6$ m$^2$ s$^{-1}$) of the LBM experiments in which diabatic heating and submonthly eddy forcing in August are prescribed north of 65°S (LMG_LBL). (d) As in (c), but for the experiments with ASM + NWIO diabatic heating (LBM_ASM + LBM_NWIO). The coloring convention for all the panels is indicated at the bottom. Superimposed with red and blue arrows in each panel is Plumb’s (1985) wave-activity flux [m$^2$ s$^{-2}$; reference on the right of (b)] for the Northern and Southern Hemispheres, respectively.
equatorward portion of the high pressure response in LBM_GLB (Fig. 1e). As shown in Fig. 7c, the local cooling can explain a substantial fraction of the midtropospheric subsidence across the subtropical SIO in LBM_GLB (Fig. 7a). Concomitantly, the moderate cyclonic vorticity with the northwest–southeast tilt is reproduced in the 200-hPa level over the subtropical SIO (Fig. 7d; the vorticity response is almost the same as in the 300-hPa level; figure not shown), acting to reinforce the zonal gradient of vorticity. With this zonal gradient thus generated, the anticyclonic vorticity advection by the zonal-mean westerlies can balance with the vorticity-tube stretching associated with the subsidence over the subtropical SIO (Fig. 7c).

The above experiments demonstrate that local diabatic cooling can reinforce the equatorward portion of the winter-time Mascarene high. This suggests the importance of local feedback between the Mascarene high and diabatic processes in austral winter, because the enhanced cooling due to the suppressed deep convective activity and increased low-level clouds is caused by the midtropospheric subsidence and near-

FIG. 9. (a) Climatological-mean August distribution of zonally asymmetric vertically integrated diabatic heating ($Q_{\text{vdf}}^* + Q_{\text{rad}}^* + Q_{\text{precip}}^*$). (b)–(d) As in (a), but for $Q_{\text{vdf}}$, $Q_{\text{rad}}$, and $Q_{\text{precip}}$, respectively, around the SIO. Unit is W m$^{-2}$. The uneven coloring convention is indicated at the bottom. The red, light blue, and black rectangles indicate the domains where forcing is prescribed in LBM_ASM, LBM_NWIO, and LBM_SIO, respectively. Vertical integration is performed from the surface to the JRA-55 model-top (∼0.1 hPa) with mass weight.

FIG. 10. Zonal sections for 25°S of August climatologies of zonally asymmetric (a) $Q_{\text{vdf}}^*$ (K day$^{-1}$), (b) $Q_{\text{rad}}^*$ (K day$^{-1}$), (c) $Q_{\text{precip}}^*$ (K day$^{-1}$), (d) cloud fraction (%) based on JRA-55, and (e) CALIPSO cloud fraction (%). The model-level JRA-55 data have been interpolated vertically onto the LBM sigma coordinate. The coloring convention is indicated at the bottom of each panel.
surface cold advection by equatorward winds associated with the high. Then one may wonder what can trigger this local feedback. One minor but possible factor is the enhanced storm-track activity in austral winter over the midlatitude SIO (Nakamura and Shimpo 2004), which facilitates the formation of low-level clouds over the subtropical SIO (Miyamoto et al. 2018). However, the low-cloud impacts on the wintertime high are found to be modest (Miyamoto et al. 2021b), as mentioned in section 1. Another possibility is remote influence from the tropics, as discussed in detail in the next subsection.

b. External modulation by the remote influences from the tropics

In austral winter (or boreal summer), convective precipitation and associated diabatic heating are enhanced around the ASM region (Fig. 9a). Concomitantly, the strong cross-equatorial Somali jet off the east coast of Africa (Fig. 4a; Rodwell and Hoskins 1995, 1996) invokes coastal upwelling and turbulent heat loss, acting to lower SST in the Arabian Sea (de Boyer Montégut et al. 2007) and thereby suppressing precipitation locally (Fig. 9a), as reviewed by Schott et al. (2009). To evaluate whether this diabatic heating/cooling can influence the Mascarene high, we force the LBM separately with the diabatic heating in August over the following two domains: 1) the ASM region, the tropical eastern Indian Ocean and western Pacific (LBM_ASM; surrounded with a red line in Fig. 9a), and 2) the northwestern Indian Ocean (NWIO) (LBM_NWIO; surrounded with a light blue line in Fig. 9a). We have confirmed that diabatic forcing in other tropical regions exerts only minor impacts over the SIO.

Figure 8d shows the combined 200-hPa streamfunction responses in LBM_ASM and LBM_NWIO. Though slightly underestimated, an anticyclonic response in each of the hemispheres resembles its counterpart in LBM_GLB (Fig. 8c), implying the importance of the tropical influence. To the southeast of the anticyclonic response over the SIO, there is a weak cyclonic response, indicative of Rossby-wave propagation as evident from poleward wave-activity fluxes (Fig. 8d).

The combined response discussed above is then decomposed into the individual contributions from the ASM and NWIO domains. Figures 12a–c show the response in LBM_ASM. The heating around the ASM region well reproduces the anticyclonic 200-hPa streamfunction response in either hemisphere in LBM_GLB (Fig. 12a), reminiscent of a Matsuno–Gill-type Rossby wave response (Matsuno 1966; Gill 1980; Kraucunas and Hartmann 2007). As revealed by the Rossby wave source analysis (Sardeshmukh and Hoskins 1988) of 200-hPa vorticity (see the appendix), the upper-tropospheric anticyclonic response over the SIO is primarily forced by absolute vorticity advection by cross-equatorial southward divergent wind response from the monsoonal convection (Fig. A1d with divergent wind response indicated by arrows) and is further reinforced by the stretching effect by the climatological Hadley cell descent (Fig. A1e), leading to the enhanced response to the heating concentrated in the opposite hemisphere. This Rossby wave propagates southeastward to produce an upper-tropospheric cyclonic response over the midlatitude SIO (Fig. 12a). The upper-tropospheric Rossby waves accompany enhanced midtropospheric subsidence over the western and central portions of the subtropical SIO (Fig. 12c), which is dynamically consistent with equatorward surface winds underneath (Fig. 12b). The location of the subsidence corresponds to the horizontal distribution of the upper-tropospheric vorticity response as discussed in the appendix. Further decomposition of the forcing domain reveals that diabatic heating over the ASM region, the tropical Indian Ocean and western Pacific all make nonnegligible contributions to the LBM_ASM response (not shown).

The circulation response to the ASM heating can modulate the wintertime Mascarene high. As a direct impact on surface pressure, the ASM heating acts to shift the surface high westward as inferred from the low pressure response centered at 30°S over the eastern subtropical SIO (Fig. 12b). More importantly, the enhanced subsidence over the western subtropical SIO suppresses condensation heating associated with deep convective clouds (Figs. 9d and 10c; Takayabu et al. 2010).
Concurrently, the subsidence facilitates low-cloud formation with the aid of near-surface cold advection by the equatorward winds, resulting in enhanced cloud-top radiative cooling (Figs. 9c and 10b). The anomalous cooling thus induced acts to reinforce the surface high as a "diabatic enhancement" (Rodwell and Hoskins 1996, 2001). Meanwhile, the LBM response accompanies ascent around the west coast of southern Australia (Fig. 12c), which may promote precipitation locally around 120°E (Fig. 9d).

The diabatic cooling in the NWIO associated with the ASM also exerts remote influences on the Mascarene high. The upper-tropospheric response in LBM_NWIO (Fig. 12d) is similar to its counterpart in LBM_ASM but is shifted westward with the opposite signs, as supported by the Rossby wave source analysis (see appendix). It accompanies midtropospheric subsidence and equatorward surface winds over the western portion of the subtropical SIO (Fig. 12f). The total subsidence response to the ASM and NWIO diabatic heating/cooling is comparable to that to the SIO heating/cooling (Fig. 7c). As in LBM_ASM, the subsidence and equatorward surface winds (Fig. 12e) act to augment diabatic cooling over the western subtropical SIO that reinforces the high. Concomitantly, the high pressure response over the westernmost subtropical SIO (Fig. 12e) acts to shift the surface high westward as its direct impact.

In summary, the diabatic heating over the ASM region as well as the tropical Indian Ocean and western Pacific acts not only to shift the wintertime surface Mascarene high westward but also to bring near-surface cold advection and midtropospheric subsidence that promote (suppress) the formation of low-level (deep convective) clouds (Figs. 11b, 12c, f). This remotely forced diabatic enhancement is included in the stronger diabatic cooling over the central and western portions of the subtropical SIO (Fig. 9a), which is partially offset by the orographic rainfall around Madagascar associated with the strengthened Mascarene high. This diabatic cooling accounts for the equatorward portion of the Mascarene high (Fig. 11a). Indeed, the combined response to the local and remote monsoonal diabatic heating/cooling (LBM_SIO + LBM_ASM + LBM_NWIO) exhibits a surface high pressure response that is stronger in the western subtropical SIO (Fig. 11c).

5. External modulation of the Mascarene high by storm-track activity enhanced along the Agulhas SST front

a. Dynamical forcing by the storm-track activity

Through the LBM and AGCM experiments, this section demonstrates the importance of the storm-track activity.
enhanced along the Agulhas SST front over the SIO in maintaining the wintertime surface Mascarene high. Figures 5c and 5d show zonally asymmetric fields of storm-track activity extracted as submonthly fluctuations in the upper and lower troposphere, respectively. As a clear indication of the storm-track core over the SIO, meridional wind fluctuations in the upper troposphere maximize along the subpolar eddy-driven westerly jet around 50°S (Figs. 4c and 5c), and poleward eddy heat flux in the lower troposphere peaks along the Agulhas SST front around 45°S (Fig. 5d).

Effects of the submonthly eddy forcing on the Mascarene high are evaluated through the LBM experiments (LBM_SIO). Note that subweekly eddies defined as 8-day high-pass-filtered fluctuations yield almost the same responses (Fig. S2 in the online supplemental material). As shown in Fig. 11d, submonthly eddies over the SIO induce a basinwide surface high pressure response centered at 40°S, 80°E. Its magnitude exceeds 9 hPa, fully accounting for the poleward portion of the high pressure response in LBM_GLB (Fig. 1e). Between this high pressure response and a low pressure response to the south, lower-tropospheric westerlies are enhanced at 50°S. Similarly, upper-tropospheric westerlies are accelerated around 55°S along the subpolar (or polar-front) jet and decelerated at 35°S along the subtropical jet, as inferred from the vorticity response (Fig. 7f). These westerly responses reflect the role of baroclinic eddies in transporting westerly momentum from the subtropical jet toward the “eddy-driven” subpolar jet and then downward to the surface. The wintertime circulation response to the eddy forcing is much stronger than its summertime counterpart (Miyamoto et al. 2021a) due to the seasonal enhancement of storm-track activity (Inatsu and Hoskins 2004; Nakamura and Shimpo 2004). Indeed, we have confirmed that this stronger response does not arise from the seasonal difference in the basic state (not shown). The eddy forcing also enhances subsidence over the northwestern portion of the induced high but only poleward of 30°S (Fig. 7e).

b. Impacts of the Agulhas SST front simulated in AFES

Since the Agulhas SST front is known to energize the storm-track activity over the SIO as mentioned in section 1, we hypothesize that the SST front acts to reinforce the wintertime high by activating baroclinic eddies. To verify this hypothesis, we compare AFES_CTL with AFES_SMTH, in which the frontal SST gradient prescribed as the lower-bconomy condition of the AGCM has been smoothed out artificially (Fig. 3b). We focus on the July–September (JAS) mean field rather than the August mean to obtain a more robust response. As in JRA-55, AFES_CTL is found to reproduce the enhanced storm-track activity over the SIO around 45°S in both the upper and lower troposphere extracted as day high-pass-filtered fluctuations (Figs. 13a,b,d,e). These features are almost the same as described in section 5a for the August submonthly variability (Figs. 5c,d). Though slightly stronger than in JRA-55, the wintertime Mascarene high is also well reproduced in AFES_CTL (Figs. 13g,h).

Differences of AFES_CTL from AFES_SMTH highlight the impacts of the sharpening of the frontal SST gradients on the storm-track activity. As shown in Fig. 13f, near-surface poleward eddy heat flux along the Agulhas SST front is significantly enhanced, suggestive of more rapid and/or frequent development of baroclinic eddies. By contrast, another peak in the near-surface storm-track activity along the sea ice edge around 60°S is found to be insensitive to the SST smoothing (Figs. 13d–f). In the upper troposphere, significant enhancement of eddy activity is found slightly downstream on the poleward side of the storm-track core (Fig. 13c). Although SST gradient has also been smoothed in other ocean basins in AFES_SMTH, only a slight weakening is found in wave-activity injection from the South Atlantic into the SIO as inferred from the anomalous Eliassen–Palm (EP) flux (Fig. 13c). These responses suggest the importance of the local impacts of the Agulhas SST front on the storm-track activity.

The stronger storm-track activity over the SIO yields enhanced eddy transport of westerly momentum toward the storm-track core near the surface, thus inducing an anticyclonic SLP+ response on its equatorward side. As shown in Fig. 13i, a high pressure response is indeed significant around 45°S with enhanced westerlies around 55°S, which is consistent with the enhanced upper-tropospheric convergence of eddy momentum flux as indicated by the anomalous EP flux (Fig. 13c) and stronger low-level poleward eddy heat flux (Fig. 13f). Similar enhancement and poleward shift of storm-track activity and subpolar jet by the frontal SST gradient were also found in AGCM experiments for the North Atlantic (O’Reilly et al. 2017) and in “aquaplanet” settings (Nakamura et al. 2008; Ogawa et al. 2012). In the westernmost portion of the basin, by contrast, the SLP+ response is negative on the equatorward side of the Agulhas SST front. This cyclonic response may be due to diabatic heating resulting from enhanced heat and moisture supply from the warm Agulhas Current system (Liu et al. 2007; Miyamoto et al. 2018; Masunaga et al. 2020). Thus, the Agulhas SST front reinforces the surface Mascarene high, especially over the central and eastern portions of the SIO.

One may notice that the SLP+ response in the AFES experiments (~1 hPa; Fig. 13i) is weaker than the planetary component of the climatological-mean Mascarene high (~6 hPa; Fig. 13h), in contrast to the LBM experiments (Figs. 1e and 11d). Besides the shallow heating over the warmer side of the SST front mentioned above, there may be other potential factors as discussed below. Even after the smoothing, a “frontal zone” with the moderately enhanced SST gradient still exists around 45°S in AFES_SMTH (Fig. 3b), which can contribute to the enhanced storm-track activity over the SIO. In addition to the local surface boundary conditions, enhanced baroclinicity in the free troposphere associated with the upper-level planetary waves further enhances the storm-track activity over the SIO (Inatsu and Hoskins 2004; Nakamura and Shimpo 2004). As suggested by the LBM response shown in Fig. 8d, the diabatic heating associated with the ASM can remotely strengthen the upper-tropospheric westerlies and thereby free-tropospheric baroclinicity in the western midlatitude SIO and the eastern subtropical SIO. This remote ASM influence may also contribute
positively to the formation of the wintertime storm-track core over the western midlatitude SIO, acting to reinforce the wintertime Mascarene high. As found in observations (Fig. 13a; Nakamura and Shimpo 2004), the incoming flux of synoptic-wave activity from the South Atlantic simulated in AFES_SMTH also acts to energize the storm-track activity over the SIO by a comparable magnitude as in AFES_CTL (Figs. 13b,c). These factors may lead to the modest responses of the storm-track activity and SLP*$ to the modiﬁed frontal SST gradient. Still, these AFES experiments highlight the connection between the Agulhas SST front and the wintertime surface Mascarene high via the enhanced storm-track activity.

6. Concluding remarks

a. Summary

Over the SIO, unlike over the other subtropical oceans, the surface subtropical Mascarene high climatologically shifts westward toward austral winter, and its strength as a planetary-wave component maximizes in late austral winter. Together with our earlier works that have shown that the wintertime Mascarene high and low-level clouds constitute a positive feedback system (Miyamoto et al. 2018, 2021b), the present study has revealed its external modulations through a series of model experiments to present its more comprehensive view (Fig. 2).

The LBM experiments have revealed that the remote influence from the tropics, including the ASM, can induce diabatic cooling (or suppressed diabatic heating) over the subtropical SIO and therefore reinforce the equatorward portion of the high (green arrow in Fig. 2a; Fig. 11c). Deep convective activity over the ASM region as well as the tropical Indian Ocean and western Paciﬁc induces an upper-tropospheric Rossby wave response with enhanced cross-equatorial southward divergent winds. This remote inﬂuence from the tropics acts not only to directly shift the surface Mascarene high westward but also to enhance midtropospheric subsidence and equatorward surface winds over the western and central portions of the subtropical SIO (Fig. 11b). The enhanced subsidence statically stabilizes and dries the free troposphere, acting to suppress deep convection over the wintertime subtropical SIO. Concurrently, the subsidence promotes low-cloud formation with the aid of the equatorward winds that yield cold advection. These cloud changes strengthen zonally asymmetric diabatic cooling over the subtropical SIO, acting to reinforce the equatorward portion of the surface Mascarene high (Fig. 11a).
The role of the storm-track activity energized by the Agulhas SST front is also highlighted. The LBM experiments have shown that the storm-track activity enhanced over the midlatitude SIO maintains the poleward portion of the wintertime Mascarene high through eddy heat and vorticity fluxes (solid blue arrow from “Storm track” to “Subtropical high” in Fig. 2a; see Fig. 11d). The AGCM experiments further indicate that the Agulhas SST front contributes to the enhanced storm-track activity and acts to reinforce the high through the net dynamical forcing of the storm-track activity on the mean flow (solid blue arrow from “Agulhas SST front” to “Subtropical high” in Fig. 2a). In addition to increasing low-level clouds (Miyamoto et al. 2018; dashed blue arrow in Fig. 2a), the storm-track activity and the Agulhas SST front reinforce the Mascarene high in austral winter, contributing to its unique seasonality.

b. Discussion

In addition to the SIO, the wintertime storm-track activities are prominent in the North Atlantic and Pacific, namely the Gulf Stream and Kuroshio–Oyashio Extension regions, respectively (e.g., Nakamura et al. 2004, 2010; Kwon et al. 2010; O’Reilly et al. 2017) despite the midwinter suppression over the North Pacific (Nakamura 1992). One may wonder if the active storm tracks in boreal winter might shift and strengthen the subtropical highs. In the wintertime Northern Hemisphere, however, large-scale orography and land–sea thermal contrasts in the extratropics force planetary waves with basin-scale surface low pressure systems (Fig. 1a; Held et al. 2002; Nigam and Chan 2009; Nakamura et al. 2010). Although net forcing from transient eddies acts to reinforce the high in the North Atlantic, the wintertime subtropical highs in the Northern Hemisphere essentially form over the eastern portions of the individual basins as a downstream element of these planetary waves (Nakamura et al. 2010).

The wintertime subtropical highs in the South Atlantic and Pacific also reside over the eastern portions of the individual basins (Fig. 1b), and low-level clouds are prevalent on the eastern sides of the highs even in austral winter (Klein and Hartmann 1993; Miyamoto et al. 2018). As shown in previous studies (Rodwell and Hoskins 2001; Xu et al. 2004; Richter and Mechoso 2004, 2006), the South American and African orography facilitates equatorward surface winds and low-level clouds over the eastern South Pacific and Atlantic, respectively. Since the resultant ocean cooling leads to enhanced land–sea thermal contrasts, the orographic forcing may drive the local atmosphere–ocean–land feedback system suggested for the summertime subtropical highs (Miyasaka and Nakamura 2005, 2010; Miyamoto et al. 2021a) even in austral winter. However, such a coupled system as above is unlikely to form around the Australian continent, because the orography is much less steep, and the landmass exists only poleward of 15°S, where the wintertime insolation is relatively weak. The concentration of the tropical diabatic heating into the ASM region (Fig. 9a) may further strengthen the interbasin differences, although the African precipitation may influence the South Atlantic subtropical high in austral winter (Richter et al. 2008). It should be noted that the prominent wintertime storm-track activity in the eastern South Atlantic (Nakamura and Shimpo 2004) may act to maintain the wintertime subtropical high locally. Further investigation is required to fully understand the wintertime subtropical highs in the South Pacific and Atlantic.

Since anticyclonic surface wind stress curl forces a subtropical oceanic gyre, the Mascarene high may indirectly strengthen the Agulhas SST front as additional feedback. Despite some debate on the relationship between the surface winds and the Agulhas Current system (e.g., Beal et al. 2011), recent studies have demonstrated that enhanced trade winds and midlatitude westerlies can strengthen the Agulhas Return Current in regional and global ocean model experiments (Durgadoo et al. 2013; Loveday et al. 2014). Thus, the influence of the Mascarene high on the Agulhas SST front might constitute another feedback loop, which should be addressed in future studies.

The monsoonal swirling of surface winds from the SIO into the north Indian Ocean across the equator (Fig. 4a) forces the southward ocean Ekman heat transport, acting to reduce the cross-equatorial SST gradient (Levitus 1987; Schott et al. 2009). This may weaken the amplitude of the seasonal cycle of the monsoon as negative feedback within the ocean–land–atmosphere coupled system (Webster 2006). Although it is well known that the diabatic heating associated with the ASM can directly force these cross-equatorial winds (Rodwell and Hoskins 1995, 1996), the Mascarene high reinforcement by the ASM might further enhance this negative feedback through the augmented southeasterly trade winds.

Last, while our focus is mainly on the maintenance, the Mascarene high undergoes a prominent seasonal transition. It reaches a maximum in the western portion of the basin in late winter after the autumn minimum, and then shifts eastward, realizing its summertime form. Full investigation of the seasonal march of the Mascarene high and the associated factors relevant to its maintenance will lead to deeper understanding of the formation and seasonal evolution of the high. It will also be interesting to test whether the proposed forcing-feedback system involving the Mascarene high and low-level clouds will change in tandem under global warming. Indeed, previous works identified that strengthening of the land–sea heating contrast (Li et al. 2013) and a poleward shift of storm-track activity (Fahad et al. 2020) may affect the Mascarene high in the future. These topics will be pursued in our future work.

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FIG. A1. The 200-hPa Rossby wave source in LBM_ASM and its decomposition. (a) Total Rossby wave source \( S \) (color shaded for every \( 2 \times 10^{-11} \, \text{s}^{-2} \)), (b) Vorticity advection by rotational winds [the sum of first and second terms on the right-hand side in (A1); color shaded for every \( 2 \times 10^{-11} \, \text{s}^{-2} \)], (c)–(f) \( S1, S2, S3, \) and \( S4 \) [individual contributions designated in (A2); color shaded for every \( 2 \times 10^{-11} \, \text{s}^{-2} \)], respectively. In (a), (b), (c), and (f), contours indicate the LBM response of 200-hPa relative vorticity (every \( 2 \times 10^{-6} \, \text{s}^{-1} \); solid and dashed lines for positive and negative values, respectively). In (c) and (d), contours indicate the basic-state field of 200-hPa absolute vorticity (every \( 2 \times 10^{-6} \, \text{s}^{-1} \); solid and dashed lines for positive and negative values, respectively). Superimposed with arrows in (b) is the basic-state (i.e., zonal-mean) field of 200-hPa rotational wind (m s\(^{-1}\)), whereas the LBM response and basic-state fields of 200-hPa divergent winds (m s\(^{-1}\)) are shown with arrows in (d) and (f), respectively.

APPENDIX

Rossby Wave Source Analysis for the Remote Influence from the Tropics

To understand the mechanisms of the remote influence from the Asian summer monsoon, we conduct the Rossby wave source analysis (Sardeshmukh and Hoskins 1988) for the upper-tropospheric response in LBM_ASM. The linearized vorticity equation with no vertical velocity may be expressed as

\[
\frac{\partial \zeta^*}{\partial t} = -v_\phi^* \cdot \nabla \zeta + \nabla \cdot \nabla \zeta^* + S, \tag{A1}
\]

where \( \zeta, \xi, \) and \( v_\phi \) denote relative vorticity, absolute vorticity, and rotational winds, respectively. In (A1), the brackets signify zonal averaging, and the source term \( S \) may be decomposed into \( S1 \) (stretching of zonal-mean absolute vorticity with anomalous divergence), \( S2 \) (advection of zonal-mean absolute vorticity by anomalous divergent winds), \( S3 \) (stretching of anomalous vorticity with zonal-mean divergence), and \( S4 \) (advection of anomalous vorticity by zonal-mean divergent winds):

\[
S = \frac{-(\zeta \nabla \cdot v_\phi^*) - v_\phi^* \cdot \nabla (\zeta)}{S1} - \frac{\zeta \nabla \cdot [v_\phi]}{S2} - \frac{\nabla \cdot [v_\phi]}{S3} - \frac{\nabla \cdot [v_\phi^*]}{S4}, \tag{A2}
\]

where \( v_\phi \) denotes divergent wind. Without \( S, \) (A1) is equivalent to the equation for barotropic Rossby waves. Thus, \( S \) represents the forcing of those waves as well as their modifications along the baroclinic westerlies.

Figure A1a shows 200-hPa Rossby wave source \( S \) in LBM_ASM with the response of relative vorticity superimposed. Over the subtropical SIO around 15°S, anticyclonic (positive) \( S \) coincides with the anticyclonic vorticity response. The anticyclonic \( S \) is mainly due to \( S2 \) (Fig. A1d), which features the advection by cross-equatorial divergent winds from the ASM regions acting on the zonal-mean absolute vorticity gradient. In addition, \( S3 \) also makes a positive contribution around the anticyclonic vorticity (Fig. A1e). The zonal-mean horizontal convergence associated with the descending branch of the Hadley circulation stretches the anticyclonic vortex tube locally, acting to reinforce the anticyclonic vorticity anomaly forced by the divergent wind response (S2). Thus, the zonal-mean Hadley circulation can be a factor for the strong response over the SIO despite the heating biased into the
Northern Hemisphere. Such an enhanced response to off-equatorial heating in the opposite hemisphere has been indicated by Kraucunas and Hartmann (2007) with a simplified model configuration. We note that the LBM response to the August tropical heating is found to weaken substantially over the SIO when the basic state is replaced with the January fields (Fig. S1). This result indicates the importance of the zonal-mean basic state in the wintertime Southern Hemisphere for the strong response to the Northern Hemisphere monsoonal heating.

As shown in Fig. A1c, $S_1$ is positive and negative in the eastern and western subtropical SIO around $30^\circ S$, respectively, exhibiting good correspondence to the 700-hPa $\omega$ response (Fig. 12c). As shown in Fig. A1b, these upper-tropospheric vorticity tendencies generated through $S_1$ are largely offset by anomalous vorticity advection with rotational winds, primarily with the upper-tropospheric zonal-mean subtropical jet crossing the anticyclonic and cyclonic vorticity responses. In this manner, vorticity balance is retained for the upper-tropospheric stationary response along the baroclinic westerlies, which is linked to the 700-hPa $\omega$ and surface pressure responses (Figs. 12b,c).

The mechanism of the LBM response in LBM_NWIO (Figs. 12d–f) is similar to its counterpart in LBM_ASM. Figure A2 shows the 200-hPa Rossby-wave source with relative vorticity response in LBM_NWIO. Equatorward divergent winds forced by the NWIO cooling produce a 200-hPa cyclonic vorticity response centered at $15^\circ S$ over the east coast of Africa (Figs. A2a,d) accompanied by an anticyclonic response around $30^\circ S$ over the SIO as the southeastward Rossby-wave propagation (Figs. 12d and A2a). To the east of the anticyclonic response, the zonal-mean westerlies yield anticyclonic vorticity advection centered at $30^\circ S$, 70°E (Fig. A2b). This vorticity tendency tends to be offset by vortex-tube stretching (Fig. A2c) in association with enhanced midtropospheric subsidence (Fig. 12f) and equatorward surface winds (Fig. 12e).

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