Structure and maintenance mechanisms of the Mascarene High in austral winter

Yuan Zhao1 | Zhiping Wen2,3 | Xiuzhen Li1,4 | Ruidan Chen1,4 | Guixing Chen1,4

1School of Atmospheric Sciences, Sun Yat-sen University, Southern Marine Science and Engineering Guangdong Laboratory (Zhuhai), Zhuhai, China
2Department of Atmospheric and Oceanic Sciences/Institute of Atmospheric Sciences, Fudan University, Shanghai, China
3Jiangsu Collaborative Innovation Center for Climate Change, Nanjing, China
4Guangdong Province Key Laboratory for Climate Change and Natural Disaster Studies, Sun Yat-sen University, Zhuhai, China

Correspondence
Xiuzhen Li, School of Atmospheric Sciences Sun Yat-sen University, Zhuhai 519082, China.
Email: lixiuzhen@mail.sysu.edu.cn

Funding information
Guangdong Province Key Laboratory for Climate Change and Natural Disaster Studies, Grant/Award Number: 2020B1212060025; National Key Research and Development Programs of China, Grant/Award Number: 2019YFC1510400; National Natural Science Foundation of China, Grant/Award Numbers: 41775043, 42088101, 42175019

Abstract
The Mascarene High (MH), is a key component of the Asian-Africa-Australia monsoon system in austral winter (JJA), spanning over the South Indian Ocean (15°–35°S, 15°–110°E). Its three-dimensional structures and maintenance mechanisms are examined in this study. It is a low-level subtropical high dominating the southern Africa and South Indian Ocean, characterized by a north-westward tilt with height, which is attributed to its spatially inhomogeneous thermal structure. Large-scale subsidence characterizes the main body of the MH, with the stronger subsidence to the east than to the west. Diagnosis using the complete form of the vertical vorticity tendency equation shows that the anticyclonic structure of the MH, which can be described by the distribution of meridional wind, is maintained mainly by the vertical gradient of diabatic heating, change in static stability, and friction dissipation. In particular, a combination of sensible heating and longwave radiative cooling results in a vertical decreasing gradient of diabatic heating in the lower troposphere. It generates the stronger southerlies over the subtropical South Indian Ocean than over the southern Africa. Meanwhile, over the South Indian Ocean, the increasing static stability as a result of the downward transport of a more stable atmosphere partly offsets the effect of the vertical gradient of diabatic heating, and southerlies still prevail there. Over the southern Africa, topographic friction dissipation induces northerlies, balancing the effect of the vertical gradient of diabatic heating with a stronger magnitude, and northerlies prevail.

KEYWORDS
Diabatic heating, friction dissipation, maintenance mechanisms, Mascarene High, static stability

1 INTRODUCTION

The Mascarene High (MH), also termed the southern Indian Ocean subtropical high, is a high-pressure area embedded in the austral subtropical high-pressure belt throughout the year. It can influence the formation of the Indian Ocean subtropical dipole (IOSD) through the surface heat flux and wind speed anomalies (Behera and Yamagata, 2001; Morioka et al., 2010; 2012). During austral summer, the trade winds out of the MH plays an important role in importing moisture transported from the warm Southwest Indian Ocean and the Agulhas
Current toward the Angola Low, affecting rainfall over southern Africa (Harrison, 1984; Cook, 2000; Tyson and Preston-Whyte, 2000; Manatsa et al., 2014; Xulu et al., 2020). The variation of MH also largely determine the movement of tropical cyclones developed over the Southwest Indian Ocean (Miyamoto et al., 2018; Xulu et al., 2020). More importantly, as a major member of the Asian summer monsoon system during austral winter (JJA, Krishnamurti and Bhalme, 1976), the MH can also modulate the Indian summer monsoon and thus the precipitation over East Asia, with its outflow crossing the equator over the western Indian Ocean and then veering toward South Asia (Xue et al., 2004; Han et al., 2017). In a future warmer climate, associated with the expansion of tropical regions and southward expansion of the Hadley Cell and storm tracks, the poleward migration of the MH is expected (Waugh et al., 2015; Cherchi et al., 2018; Grise et al., 2018; Xulu et al., 2020). Several studies have found the interannual variability of the MH is associated with Indian Ocean SSTs, ENSO, and the Antarctic Oscillation (Behera and Yamagata, 2001; Fauchereau et al., 2003; Hermes and Reason, 2005; Morioka et al., 2015; Xulu et al., 2020). Hence, it is of great importance to improve our knowledge about the activity of the MH in austral winter.

Compared to the Northern Hemisphere, there is less continental land over the Southern Hemisphere. Lack of observations over oceans, circulation characteristics over the Southern Hemisphere cannot be clearly disclosed unless satellite data are employed. The features and mechanisms of austral systems are usually disclosed by examining the consistency and difference with their counterparts in the Northern Hemisphere (Wu and Liu, 2003; Liu et al., 2004; Miyasaka and Nakamura, 2010). In the Northern Hemisphere, surface subtropical anticyclones reside over the eastern portions of the subtropical ocean basins throughout the year and show a strong seasonal variability in intensity and meridional location—that is, the highest intensity and northernmost location occur in summer. The vertical motion inside the subtropical highs is diverse, characterized by strong subsidence over the eastern portion and ascending motion to the west in summer (Miyasaka and Nakamura, 2005, 2010; Wu et al., 2009). The axis of the subtropical high tends to tilt toward the warm sector, that is, equatorward and westward with height based on the geostrophic relationship. However, the MH exhibits a seasonality distinct from its counterpart in the Northern Hemisphere and other ocean basins. It resides over the eastern part of the South Indian Ocean in austral summer, while strengthens and shifts westward in winter, covering the southern Africa and the South Indian Ocean (Miyasaka and Nakamura, 2010; Miyamoto et al., 2018; Vidya et al., 2020). As a result, whether the spatial and thermal structures of the MH in austral winter, such as vertical motion and axis displacement, show features similar or opposite to its counterpart in boreal summer in the Northern Hemisphere remains unknown.

Traditionally, subtropical highs are regarded as a result of a sinking branch of the Hadley circulation (Rodwell and Hoskins, 1996; Reboita et al., 2019). However, the stronger subtropical highs observed in summer in the Northern Hemisphere seem to conflict with this argument, as the zonally symmetric Hadley circulation predicts much weaker subtropical subsidence in summer (Hoskins, 1996; Rodwell and Hoskins, 1996). Hence, the theory of the Hadley circulation does not seem to fully explain the existence of subtropical highs (Hoskins, 1996; Liu et al., 2001; Rodwell and Hoskins, 2001; Li et al., 2019, 2020). Further studies have shown that subtropical cell-type high-pressure systems instead of a zonally uniform high-pressure belt are detected over subtropical regions, as a manifestation of climatological summertime stationary waves maintained by global diabatic heating (Ting, 1994), especially monsoonal heating (Hoskins and Rodwell, 1995; Rodwell and Hoskins, 1996; Chen et al., 2001). Rodwell and Hoskins (2001) further concluded that the combined effects of topography and monsoonal heating to the east are of primary importance for the formation of surface subtropical highs and subsidence aloft, with reinforcement by local cooling over the eastern ocean.

By simplifying the vorticity equation, Wu and Liu (1999) further emphasized the crucial role of vertical inhomogeneity in diabatic heating (Wu et al., 1999; Liu et al., 1999a; 1999b; 2001; Wu and Liu, 2003; Liu et al., 2004). Further studies by Wu and Liu (2003) and Liu et al. (2004) proposed a summertime subtropical quadruple heating pattern and illustrated the key role of vertical increases (decreases) in diabatic heating in motivating the poleward (equatorward) flow underneath in both hemispheres. In contrast, Miyasaka and Nakamura (2005, 2010) emphasized that the near-surface thermal contrast between ocean and land could accurately reproduce summertime subtropical highs in their modelling study, and the local positive feedback between low-level longwave radiative cooling, subtropical highs, and marine stratus cloud over the eastern ocean could play an important role in maintaining summertime subtropical highs (Klein and Hartmann, 1993; Norris, 1998; Norris and Klein, 2000; Wu et al., 2009). Furthermore, a study by Lee et al. (2013) highlighted the interhemispheric influence of northern summer monsoons on southern subtropical anticyclones. Enhanced summer monsoonal convection produces subsidence over the Southern Hemisphere via interhemispheric meridional overturning circulation and increases sea level pressure locally and remotely by motivating stationary
barotropic Rossby waves. Hence, the formation and maintenance of the subtropical highs are complicated. Previous studies have widely investigated the mechanisms of the summertime subtropical high, especially in the Northern Hemisphere. However, the maintenance mechanism of the MH in the southern Indian Ocean and the relative importance of the different processes involved have not been fully understood and deserve more attention.

This study aims to investigate the spatial structure and maintenance mechanisms of the wintertime MH, with emphasis on the relative importance of different physical processes in its maintenance. The causes of the key processes will be investigated in detail. The rest of this paper is organized as follows. In Section 2, the data and methods applied in this study are presented. The climatological dynamic and thermal structure of the MH are analysed in Section 3. The maintenance mechanisms of the MH are demonstrated in Section 4, focusing on the physical processes associated with external and internal forcing. Conclusions and discussion follow in Section 5.

\[ \frac{\partial \zeta}{\partial t} + \mathbf{V} \cdot \nabla \zeta + \beta \mathbf{v} = (1 - K)(f + \zeta) \frac{\partial \theta}{\partial P} - (f + \zeta) \frac{1}{\theta} \frac{\partial \theta}{\partial t} + 1 \frac{\partial F}{\partial \zeta} \cdot \nabla \theta + f + \zeta \frac{\partial Q}{\partial \zeta} - \frac{1}{\theta} \frac{\partial \theta}{\partial z} \frac{\partial Q}{\partial x} + 1 \frac{\partial \theta}{\partial z} \frac{\partial Q}{\partial y} + \frac{1}{\theta} P_e \frac{d}{dt} \left( \frac{1}{\theta} \right) - \frac{1}{\theta} \frac{d}{dt} \left( C_D \right) \]  

(1)

\[ \text{DATA AND METHODS} \]

\[ \text{2.1 Data} \]

The monthly mean reanalysis data used in this study are obtained from the Japanese 55-year Reanalysis (JRA-55), the global reanalysis constructed by the Japan Meteorological Agency (JMA; Kobayashi et al., 2015; Harada et al., 2016), from 1979 to 2017, with a horizontal resolution of 1.25° longitude × 1.25° latitude. This data set extends from 1,000 to 1 hPa, with 37 vertical pressure levels. Compared with other reanalysis data, JRA-55 has improved the representation of climate variability in the tropics and the temporal consistency throughout the reanalysis period (Chen et al., 2014; Harada et al., 2016). It is selected also by considering diabatic heating rates data in different levels is provided which is crucial to investigate the vertical structure of diabatic heating in Section 4.2. The variables employed in this study are geopotential height, horizontal wind fields, temperature, and vertical velocity. Monthly heating rate data provided by JRA-55 are also used to investigate the horizontal and vertical distributions of diabatic heating. Furthermore, we also used monthly mean reanalysis data extracted from the National Centers for Environmental Prediction (NCEP)/Department of Energy (DOE) Reanalysis 2 (NCEP-2) and the European Centre for Medium-Range Weather Forecasts (ECMWF) Interim Reanalysis Data (ERA-Interim) to verify the major features. The austral winter is defined as the mean of June–August (JJA).

\[ \text{2.2 Methods} \]

In previous studies (Wu et al., 1999; Liu et al., 1999a; 1999b; 2001; 2004), the complete form of the vertical vorticity tendency equation has proven to be a powerful tool to diagnose the formation and maintenance mechanisms of subtropical anticyclones. The specific information on the derivation of the equation is provided in Appendix A. Deduced from the Ertel potential vorticity theory, the equation can be written as:
represent changes in the internal thermodynamic structure of the atmosphere (Wu and Liu, 1999; Wu, 2001). Hence, by applying the complete form of the vertical vorticity tendency equation, the relative contributions of dynamical and thermodynamical processes to the tendency of vertical vorticity can be evaluated.

Previous studies have confirmed that the formation and maintenance of subtropical highs are attributed mainly to spatial inhomogeneity in diabatic heating (Liu et al., 2001; Rodwell and Hoskins, 2001; Wu and Liu, 2003; Liu et al., 2004). In this study, we use the apparent heat source $Q$ to analyse the diabatic heating associated with the maintenance of the MH in the complete form of the vertical vorticity tendency equation, which is computed as a residual in the thermodynamic equation by referring to Yanai et al. (1973) and Luo and Yanai (1984):

$$Q = C_p \left[ \frac{\partial T}{\partial t} + \nabla \cdot \mathbf{V} T + \left( \frac{p}{p_0} \right) K \frac{\partial \theta}{\partial p} \right]$$

where $T$ is temperature, $\theta$ is potential temperature; $\mathbf{V}$ is horizontal velocity; $\omega$ is vertical velocity; $K = R/C_p$, $R$ and $C_p$ are the gas constant and specific heat capacity in dry air at constant pressure; and $p_0 = 1,000$ hPa. To further compare the contributions of different types of diabatic heating, the latent heating rate ($LH$; which includes the convective heating rate and large-scale condensation heating rate), sensible heating rate ($SH$; also termed the vertical diffusion heating rate), and radiative heating rate ($RH$) derived from JRA-55 are also employed.

It should be noted with caution that $Q$ computed as a residual in the thermodynamic equation instead of the total heating rate ($Q_T = LH + SH + RH$) derived from JRA-55 is preferred in the calculation of $V_6$–$V_8$ in Equation (1) for better equation closure. In fact, the apparent heat source ($Q$) shows high consistency with the total heating rate ($Q_T$) in spatial distribution, but with a slightly smaller magnitude (not shown).

3 | DYNAMIC AND THERMAL STRUCTURES OF THE MASCARENE HIGH

To depict the spatial structures of the MH, Figure 1 shows the climatology of wintertime geopotential height along with horizontal wind fields from near the surface to 500 hPa over the South Indian Ocean. Clearly, the MH is a cell-type anticyclonic high-pressure system with a well-defined core region, constrained in the lower troposphere below 700 hPa. Its activity centre lies over the western South Indian Ocean at around 30°S near the surface, with its outflow rotating anticyclonically. With the westward shift of the MH and absence of the Angola Low in austral winter (Howard and Washington, 2018; Miyamoto et al., 2018; Xulu et al., 2020), abundant moisture originating from the South Indian Ocean is advected by the south-easterly outflow of MH, then converges over the northern Africa and South Asia. As a result, strong...
precipitation is detected there, whereas droughts appear over the southern Africa (not shown). The activity centre of the MH tilts north-westward with height, with two split centres lying over the southern Africa and the western South Indian Ocean at 850 hPa, and one core over the southern Africa at around 20°/C14S at 700 hPa. At 500 hPa, it transforms into a zonally uniform belt across the subtropical region, with strong westerlies prevailing to the south.

The vertical tilt of the MH may be related to its thermal structure due to the thermal wind relation (Holton and Hakim, 2012). To clarify this issue, the vertical distribution of temperature along with geopotential height in both the meridional and zonal directions is examined. The meridional-vertical sections of temperature along with the geopotential height deviation from the equatorial value averaged between 20°/C14 and 110°/C14E are shown in Figure 2a. The geopotential height characterized by a subtropical high-pressure centre is detected only in the lower troposphere below 700 hPa, consistent with the result shown in Figure 1. The MH tilts slightly equatorward with height to the high-temperature region in low latitudes, which is consistent with other subtropical highs characterized by an axis tilting toward the warm sector with height. A zonal-vertical section of the zonal deviation of temperature and geopotential height averaged between 20° and 110°E is shown in Figure 2b. The geopotential height deviation is zonally standardized (divided by the standard deviation at a certain latitude) so that the values at different levels are comparable. A positive zonal deviation of geopotential height is observed between 20° and 60°E, indicating the location of the MH centre over the southern Africa and the western South Indian Ocean in the lower troposphere. In the upper troposphere, though the cell-type anticyclone is deformed into a zonal belt, a ridge can still be found overlying the MH. Together, a large zonal deviation in temperature is found in the lower troposphere, reflecting a strong land-sea thermal contrast, with higher temperature over the southern Africa and Madagascar and lower temperature over the subtropical South Indian Ocean basin. The zonal deviation of temperature overturns at around 60°E. As a result, the centre of the MH tilts westward with height to the higher temperature over the land region in the lower troposphere. In the upper level, an opposite zonal gradient of temperature is found, which might be why the upper ridge tilts eastward with height (Figure 2b).

To further explore the dynamic structure of the MH, the vertical motion associated with the subtropical high over the South Indian Ocean is examined (Figure 3a). Large-scale subsidence characterizes the main body of the MH, with the strongest subsidence east of the southern Africa and the central South Indian Ocean, respectively. That is, the maximum subsidence lies east of the centres of the MH depicted by the geopotential height and wind fields. Such a double-centre structure of descending motion might be attributed to the zonal inhomogeneity of the local meridional circulation expanding between tropics and subtropics. As is shown in Figures 3b,c, two local meridional circulation cells with their descending branches over the western (20°–50°E) and eastern (60°–110°E) parts of the MH are distinct from each other. In the western part, a typical double-loop structure over the Southern and Northern Hemispheres are detected, characterized by tropical ascending motion associated with North African monsoonal heating (0°–20°N) and two subtropical descending branches off the equator (Figure 3b). The equatorial ascent is much

![Figure 2](wileyonlinelibrary.com)
weaker than the subtropical descent, illustrating that the meridional circulation might play only a partial role in the subsidence over the western MH. Instead, over the eastern part of the MH, the meridional circulation is a single loop, characterized by strong ascending motion over the equator and subtropical Northern Hemisphere and descending motion over the subtropical Southern Hemisphere (Figure 3c). Such a strong meridional circulation is triggered by South Asian monsoonal convective heating, the strongest global heating source. As Lee et al. (2013) point out, the boreal summer monsoonal convection plays an important role in either maintaining or strengthening the southern subtropical anticyclones via interhemispheric meridional overturning circulations. Hence, zonal asymmetry appears in the dynamic structure of the MH, which might imply a zonal asymmetry in the maintenance of the MH; this will be discussed in detail in Section 4.

A comparison based on JRA-55, NCEP-2, and ERA-Interim shows quite similar spatial distributions of the MH (not shown), indicating that the conclusions associated with the characteristics of the MH in JRA-55 are reliable. In summary, the MH shows spatial inhomogeneity in dynamic and thermal conditions, with its ridge tilting equatorward and westward with height to a warmer section. The west-to-east discrepancy in the equator-subtropics meridional circulation cell associated with the MH indicates that different processes might be occurring in its maintenance dynamics.

4 | MAINTENANCE MECHANISMS OF THE MASCARENE HIGH

4.1 | Physical processes involved in maintaining the Mascarene High

The important roles of topography, summertime diabatic heating, and subsidence of the interhemispheric meridional overturning circulation in the formation and maintenance of global subtropical anticyclones have been emphasized in previous studies (Rodwell and Hoskins, 2001; Wu and Liu, 2003; Liu et al., 2004; Miyasaka and Nakamura, 2005, 2010; Lee et al., 2013; Reboita et al., 2019). Compared with the South Atlantic and Pacific subtropical highs that reside over the east portion of the ocean (Miyamoto et al., 2018; Reboita et al., 2019), the MH covers the southern Africa continent and the South Indian Ocean and shows spatial inhomogeneous structure in austral winter. Moreover, the MH lies over the large-scale subsidence branched of the meridional circulation that associated with North African and South Asian monsoonal heating. Therefore, the distinct characteristic of the MH implies that its maintenance mechanism may be diverse from other subtropical highs.

The complete form of the vertical vorticity tendency equation (Equation (1)) is a preferable tool for diagnosing the formation and maintenance mechanisms of the MH, as it includes not only dynamic elements but also external and internal forcings (Wu and Liu, 1999; Liu et al., 2001; Wu, 2001). The spatial distributions of the
different physical processes involved in the vertical vorticity tendency equation at 850 hPa are shown in Figure 4. The level of 850 hPa is selected because the MH is strong at this level and its outflow plays an important role in the Asian summer monsoon. The vertical gradient at 850 hPa is calculated as the centred finite difference between the values at 825 and 875 hPa. Because it is the maintenance of the climatological mean of the MH that is being examined, the tendency of the vorticity is neglected, as is the horizontal advection of the relative vorticity because it is weak over the main body of the MH over the subtropical South Indian Ocean (Figure 4a). That is, the change in vorticity is controlled mainly by the advection of geostrophic vorticity (\( \beta \nu \)), and it is balanced by the frictional dissipation (V5), vertical gradient of diabatic heating (V6), and change in static stability (V9) over the region where the MH lies. The effect of vertical velocity (V3) and heating itself (V4), the horizontal gradient of diabatic heating (V7–V8), and the mapping from horizontal vorticity components to vertical vorticity components (V10) are around one order smaller and are negligible in the maintenance of the MH. Hence, by applying the scale analysis technique, Equation (1) can be simplified as:

\[
\begin{align*}
\text{(a) relative vorticity advection, } & \mathbf{V} \cdot \nabla \zeta \\
\text{(b) geostrophic vorticity advection, } & \beta \nu \\
\text{(c) effect of vertical velocity, } & (1 - K)(f + \zeta)\frac{\partial \theta}{\partial t} \\
\text{(d) effect of heating itself, } & - (f + \zeta) \frac{4 \pi}{R} \mathbf{\nabla} \cdot \mathbf{\nu} \\
\text{(e) frictional dissipation effect, } & \frac{1}{R} \mathbf{\nabla} \cdot \mathbf{\nu} \\
\text{(f) vertical, } & \frac{1}{R} \mathbf{\nabla} \cdot \frac{1}{\rho} \frac{\partial \theta}{\partial z} \\
\text{(g) zonal, } & - \frac{1}{R} \mathbf{\nabla} \cdot \frac{1}{\rho} \frac{\partial \nu}{\partial x} \\
\text{(h) meridional gradient of diabatic heating, } & \frac{1}{R} \mathbf{\nabla} \cdot \frac{1}{\rho} \frac{\partial \nu}{\partial y} \\
\text{(i) change in static stability, } & \frac{1}{\alpha \rho} \frac{\partial \zeta}{\partial t} \\
\text{(j) mapping from horizontal vorticity to vertical vorticity at 850 hPa, } & \text{[Colour figure can be viewed at wileyonlinelibrary.com]}
\end{align*}
\]
In Equation (3), $\beta \nu$ can be regarded as the joint result of the vertical gradient of diabatic heating, frictional dissipation, and change in static stability. This is different from the simple Sverdrup balance in Liu et al. (2004), in which only the effect of the vertical gradient of diabatic heating is emphasized when the maintenance of the western Pacific subtropical high is studied, based on the premise that frictional dissipation and the role of changes in the internal thermodynamic structure of the atmosphere are negligible. However, in this study focusing on the MH over the southern Africa and the subtropical South Indian Ocean, these two terms are non-negligible, and their roles in maintaining the MH deserve more exploration.

As shown Figure 4f, the vertical gradient of diabatic heating is positive and spatially uniform over the subtropical South Indian Ocean basin, with its centre lying over the east-central subtropical South Indian Ocean. Positive values can also be found over the southern African continent, indicating that the vertical gradient of diabatic heating contributes positively to $\beta \nu$ and is favourable for the maintenance of northerlies over the western and eastern parts of the MH. The spatial distribution of frictional dissipation (Figure 4e) is large over the southern African continent due to the orographic effect, which is negative and contributes to the maintenance of northerlies over the southern Africa, counteracting the effect of the vertical gradient of diabatic heating. Over the South Indian Ocean, the frictional dissipation is weak, and the effect of the vertical gradient of diabatic heating is offset mainly by the change in static stability (Figure 4i). The change in static stability exhibits a zonally elongated negative centre over the subtropical South Indian Ocean, similar to that of the vertical gradient of diabatic heating but with opposite sign. It is weak over the continent, which might be related to the weak static stability when the atmosphere is unstable. Hence, a west-to-east discrepancy over the southern Africa and the subtropical South Indian Ocean characterizes the maintenance of $\beta \nu$ and thus the structure of the MH.

To show the west-to-east discrepancy more quantitatively, the regional mean of the key terms in Equation (3) is calculated over the southern Africa (continental region, $35^\circ$–$15^\circ$S, $15^\circ$–$35^\circ$E), with prevailing northerlies on the western side of the MH, and over the subtropical South Indian Ocean (ocean region, $35^\circ$–$15^\circ$S, $60^\circ$–$110^\circ$E), with prevailing southerlies (Figure 5). Over the continental region, negative $\beta \nu$ is the residual of the positive vertical gradient of diabatic heating ($-1 \times 10^9$) and negative frictional dissipation ($-1.8 \times 10^9$). That is, the northerlies over the southern Africa are maintained mainly by the balance of northerlies related to the potential effect of topographic friction or tropospheric mixing and the relatively weaker southerlies attributed to the vertical decrease in diabatic heating with height. In contrast, over the eastern side of the MH, positive regional $\beta \nu$ is the residual of the vertical gradient of diabatic heating ($>2 \times 10^9$) and the changes in static stability ($<-1.5 \times 10^9$), with the regional mean of frictional dissipation negligible in the troposphere. That is, the southerlies over the subtropical South Indian Ocean are stimulated by a strong vertical decrease in diabatic heating with height in the lower troposphere, which is partly offset by the change in static stability in the atmosphere.

Hence, the northerlies over the continental region west of the MH can be regarded as the balance of the friction effect and the relatively weaker vertical decrease in diabatic heating with height, whereas the southerlies over the ocean region east of the MH are maintained by the effect of a stronger vertical decrease in diabatic heating and the change in static stability. It is easy to understand the greater friction effect over the continent than the ocean, but it is still unclear why a west-to-east discrepancy appears in the vertical structure of diabatic heating and change in static stability over the subtropical South Indian Ocean. The physical processes involved in these two terms will be investigated in detail.

### 4.2 Inhomogeneity in the vertical structure of diabatic heating

Previous studies have demonstrated a quadruple heating pattern over subtropical regions, and the relative
importance of different types of diabatic heating in the formation and maintenance of summertime subtropical anticyclones over both hemispheres have been proposed (Wu and Liu, 2003; Liu et al., 2004). In the summertime, the eastern ocean is characterized by strong longwave radiative cooling, whereas the western and eastern continent are dominated mainly by sensible heating and convective heating, and the western ocean is characterized by two forms of dominant heating, with longwave radiative cooling prevailing over convective heating (Wu and Liu, 2003). However, for the Southern Hemisphere and in the winter season, extremely strong monsoonal convective heating is absent, and the dominant diabatic heating pattern and its associated effects on the maintenance of subtropical highs might change.

To illustrate, a vertical-zonal section of different types of heating rates and their net effect–total heating rate averaged over 35°–15°S are presented in Figure 6 to examine their spatial distributions and relative importance. The total heating is large only in the lower troposphere, characterized by a strong vertical gradient. Diabatic heating is constrained near the surface and decreases with height to diabatic cooling above 850 hPa, with its centre detected over the subtropical South Indian Ocean at around 800 hPa (Figure 6a). Of the four different kinds of diabatic heating rates, because the magnitude of the solar radiative heating rate in the atmosphere is much smaller (not shown), only the characteristics of the sensible, latent, and longwave radiative heating rates are examined in Figure 6b–d, respectively. Clearly, the net diabatic heating near the surface is attributed mainly to sensible heating that is positive and extends to around 700 hPa over the southern African continent, with its maximum at around 850 hPa due to the local high altitudes. In contrast, strong sensible heating can be found only near the surface and decreases abruptly with height then turns into weak sensible cooling above 850 hPa over the subtropical South Indian Ocean. The diabatic cooling in the free atmosphere is contributed mainly by the longwave radiative cooling, the minimum of which lies over the subtropical South Indian Ocean at around 800 hPa. In contrast to the sensible heating and longwave radiative heating rate, the intensity of the latent heating rate is weaker over the region of interest; it acts as cooling near the surface and turns into heating with height over the subtropical South Indian Ocean.

Liu et al. (2004) and Wu and Liu (2003) have concluded that summer monsoonal convective heating to the east of continents in the upper troposphere plays an important role in the formation and maintenance of subtropical highs in the Northern Hemisphere, whereas latent heating in winter seems negligible over the southern Africa and the subtropical South Indian Ocean due to the absence of strong convection. In addition, the centre of longwave radiative cooling is detected over the central South Indian Ocean instead of the eastern portion of the ocean, which might be related to westward displacement of the centre of the MH in austral winter. This is different from the quadruple heating pattern, which maintains the seasonal peak of the subtropical high in the Northern Hemisphere in summer. Hence, the net diabatic heating within the MH is distinct and shows a negative vertical gradient in the lower troposphere as a combined result of sensible heating near the surface and longwave radiative cooling in the lower troposphere.

As mentioned above, though diabatic heating shows a decreasing trend with height over both the southern Africa and the South Indian Ocean, an apparent west-to-east discrepancy appears, and the contribution of the different kinds of heating are different in these two regions. To clarify, the vertical profiles of the regional average of the different kinds of heating rates over the southern Africa (35°–15°S, 15°–35°E) and the subtropical South Indian Ocean (35°–15°S, 60°–110°E) are shown in Figure 7. Over the continent, because the precipitation over the southern Africa in winter is scarce, latent heating is negligible. The profile of the total diabatic heating is determined by sensible heating and longwave radiative cooling, which show a similar decreasing trend with height. As a result, the total diabatic heating is positive near the surface and decreases slightly with height; it becomes negative above 850 hPa and reaches its minimum at around 700 hPa (Figure 7a). In contrast, latent heating is non-negligible over the subtropical South Indian Ocean. The total heating is determined by all three kinds of heating, transitioning from heating near the surface to cooling in the lower troposphere and reaching a minimum at around 800 hPa (Figure 7b). It shows a much sharper vertical decreasing trend in the lower troposphere than that over the southern Africa as a joint result of stronger decreasing trends in both sensible heating and longwave radiative cooling. The latent heating shows a contrary vertical profile, but it is weak and can only partly compensate for the other two kinds of heating. Hence, the vertical decreasing gradient of diabatic heating is much stronger over the ocean than over the continent.

As it was mentioned in Section 4.1 that the effect of the vertical decrease in diabatic heating spanning from the southern Africa to the South Indian Ocean is favourable for the maintenance of southerlies, we will now explain its relevant mechanisms. Referring to Liu et al. (2001), considering the short-term evolution, in a statically stable atmosphere (θe > 0) in the Southern Hemisphere (f < 0), the heating that decreases with
height in the lower troposphere will generate a positive anticyclonic vorticity source as:

\[
\frac{\partial \zeta}{\partial t} + \frac{f + \zeta}{\theta_z} \frac{\partial Q}{\partial z} > 0
\]

Thus, the vertical decreasing gradient of diabatic heating over the key region is conducive to triggering a low-level anticyclonic circulation in the short-term evolution. However, from a long-term evolution perspective, \(\frac{\partial \zeta}{\partial t}\) can be negligible, and the geostrophic vorticity advection \((\beta v)\) from high latitudes by the southerlies can balance the positive vorticity source forced by diabatic heating following the Sverdrup balance (Wu et al., 1999; Liu et al., 1999a; 1999b; Wu and Liu, 2003; Liu et al., 2004). Hence, the vertical decrease in diabatic heating over the
key region tends to generate southerlies in the long-term evolution, with that over the ocean larger than that over the continent.

### 4.3 Change in static stability

As it was shown in Figures 4 and 5 that a change in static stability plays an important role in the maintenance of the MH by offsetting the effect of the vertical gradient of diabatic heating over the subtropical South Indian Ocean, to further explore the physical processes involved in the change in static stability, it is decomposed as:

\[
\frac{1}{α}P_{E} \frac{d}{dt} \left( \frac{1}{θ_{z}} \right) = -\frac{1}{α} P_{E} \left( \frac{1}{θ_{z}^2} \right) \frac{dθ_{z}}{dt} = -\frac{1}{α} P_{E} \left( \frac{1}{θ_{z}} \right) \left[ \frac{∂θ_{z}}{∂t} + \nabla \cdot (\nabla θ_{z}) \right] = -\frac{1}{α} P_{E} \left( \frac{1}{θ_{z}} \right) \left[ \frac{u}{θ_{z}} + \frac{v}{θ_{z}} \right] + w \frac{∂θ_{z}}{∂z}
\]

(5)

The change in static stability (Lagrangian time derivative of $θ_z$) can be decomposed into the time rate of change of $θ_z$ and the three-dimensional advection of $θ_z$ (denoted by $∂θ_z/∂t$ and $V \cdot \nabla θ_z$). Considering the long-term evolution, the local tendency of $θ_z$ can be negligible, and the change in static stability is determined by the advection of static stability. The advection of $θ_z$ can further be divided into horizontal advection ($u dθ_z/dx + v dθ_z/dy$) and vertical advection ($w dθ_z/dz$), which are caused by the spatially inhomogeneous distribution of static stability and atmospheric motion. Figure 8 shows the distribution of the change in static stability and its vertical and horizontal advection components. The spatial distribution of the change in static stability is characterized by a widespread and relatively uniform negative value over the subtropical South Indian Ocean. That is, the static stability dominated by the vertical advection of $θ_z$ decreases with time over the ocean, which plays an important role in the maintenance of the MH as discussed in Section 4.1. In contrast, the scattered distribution of the change in static stability over the southern Africa, which has a negligible effect on $βν$, is caused mainly by the weak horizontal advection of $θ_z$.

Furthermore, scale analysis reveals that though the magnitude of vertical velocity ($w$, e$^{-1}$) in Equation (5) is an order less than the horizontal wind velocity ($u$, $v$), the vertical gradient of static stability ($∂θ_z/∂z$) with an order of e$^{-6}$ is much larger than the horizontal gradient ($∂θ_z/∂x$, $∂θ_z/∂y$) with orders of e$^{-10}$–e$^{-11}$. Hence, the vertical advection of static stability is much larger than the horizontal advection; that is, the change in static stability is caused mainly by the vertical advection of static stability.

![FIGURE 8](image_url) Spatial distributions of (a) change in static stability, (b) vertical advection change in static stability, and (c) horizontal advection change in static stability (units: 10$^{-10}$ s$^{-2}$) [Colour figure can be viewed at wileyonlinelibrary.com]

Furthermore, the vertical velocity and vertical gradient of static stability at 850 hPa are shown in Figure 9. The widespread subsidence and positive vertical gradient of static stability are detected over the subtropical South Indian Ocean with similar spatial distribution. That is, the static stability increases with height in the lower troposphere where descending motion dominates. As was shown in Figure 3c, the descending branch of the interhemispheric meridional overturning circulation is induced by strong monsoonal convection over South Asia (Lee et al., 2013). Hence, air with higher static stability is transported downward to lower static stability surroundings over the subtropical South Indian Ocean, and the static stability of the air parcels is decreased, further maintaining the MH. Here, two questions remain: Why is the static stability characterized by a vertical increasing gradient at 850 hPa? How does the change in static stability influence the maintenance of the MH?

Since the thermodynamic structure of the atmosphere can be modified by the forcing of external diabatic heating, the vertical distribution of static stability and its
association with diabatic heating are explored in Figure 10. Static stability enhances with altitude over the key region and reaches its maximum at around 800 hPa, where the strongest diabatic cooling due mainly to longwave radiative cooling appears. As diabatic cooling stabilizes the air, the atmosphere there becomes the most stable, showing a much larger positive vertical gradient of static stability over the South Indian Ocean than over the southern Africa. Hence, as a response to the forcing of external heating, the static stability increases with height over the subtropical South Indian Ocean in the lower troposphere.

Another important question is how the change in static stability influences the maintenance of the structure of the MH over the ocean. Considering the short-term evolution, the downward transport of air with larger static stability could stabilize the low-level atmosphere and then weaken the anticyclonic vorticity source forced by the vertical gradient of diabatic heating, referred to in Equation (4). In other words, the increasing stability of the atmosphere is unfavourable for the further development of the disturbance vortex source, which partly counteracts the effect of the vertical gradient of diabatic heating. From a long-term evolution perspective, the geostrophic vorticity advection (βυ) from low latitudes by the northerlies should be excited to balance the negative vorticity source forced by the stabilization of the low-level atmosphere. Hence, over the subtropical Indian Ocean, the vertical increasing diabatic cooling tends to stimulate southerly wind, which is partly compensated by northerly wind motivated by the change in static stability. That is, the southerly geostrophic vorticity advection on the east side of the MH is a result of the counterbalance between the vertical gradient of diabatic heating and the change in static stability.

5 | CONCLUSIONS AND DISCUSSION

This study has investigated the dynamic and thermal structures of the climatological MH and its maintenance mechanisms in austral winter based on monthly reanalysis data from JRA-55. The MH has a cell-type anticyclonic structure and spans from the southern Africa to the subtropical South Indian Ocean in the lower troposphere below 700 hPa. Its activity centre tilts equatorward and westward with height due to its thermal structure, which shows a warmer distribution over the north and west of the MH. The dynamic structure of the MH is characterized by large-scale subsidence and shows a west-to-east discrepancy: a stronger descending motion to the east is the sinking branch of the single-loop Hadley circulation triggered by South Asian monsoonal convective heating; a weaker descending motion to the west results from the southern branch of the double-loop Hadley circulation associated with much weaker African monsoonal convective heating.

By applying the complete form of the vertical vorticity tendency equation, the maintenance mechanism of the anticyclonic structure of the MH is simplified as the causes of the spatial distribution of advection of geostrophic
vorticity ($\beta v$). It is a joint result of the vertical gradient of diabatic heating, frictional dissipation, and the change in static stability. It shows a west-to-east discrepancy over the southern Africa and the subtropical South Indian Ocean. The northerlies over the southern Africa can be regarded as the balance of the strong friction effect and the relatively weaker vertical gradient of diabatic heating, whereas the southerlies over the subtropical South Indian Ocean are maintained by the effect of the stronger vertical gradient of diabatic heating and the change in static stability. A schematic diagram is presented in Figure 11. Diabatic heating within the MH, as a combination of sensible heating and longwave radiative cooling, shows a vertical decreasing gradient in the lower troposphere. According to the Sverdrup balance (Liu et al., 2004), diabatic heating with a vertical decreasing gradient tends to generate southerlies with stronger magnitude over the subtropical South Indian Ocean than over the southern Africa. Over the southern Africa, strong topographic friction dissipation balances the effect of the vertical gradient of diabatic heating, and northerlies prevail. Over the South Indian Ocean, the strongest diabatic cooling dominated by longwave radiative cooling appears at 800 hPa and makes the atmosphere there the most stable. By the strong descending motion of the Hadley circulation triggered by the South Asian monsoonal convection, air with higher static stability is transported downward to lower layers. It can stabilize the low-level atmosphere and partly counteract the effect of the vertical gradient of diabatic heating. Whereas the effect of vertical gradient of diabatic heating is a bit stronger than the change in static stability over the South Indian Ocean, and thus southerlies prevail there. The dynamic and thermal processes relevant to the maintenance of the MH reveal an important aspect of the winter climate over the southern Africa and the subtropical South Indian Ocean.

This study has emphasized the role of longwave radiative cooling in dominating the vertical distribution of diabatic heating. It is still unclear why the strongest longwave radiative cooling appears over the South Indian Ocean at around 800 hPa. In austral winter, longwave radiative cooling characterizes the entire subtropical South Indian Ocean, whereas it is constrained over the eastern South Indian Ocean in summer (Wu and Liu, 2003; Liu et al., 2004; Miyasaka and Nakamura, 2010). A similar seasonal discrepancy can be found in the spatial displacement of the MH (Miyamoto et al., 2018), where a feedback might occur. As suggested by previous studies, a positive feedback between marine boundary layer clouds, longwave radiative cooling, and subtropical highs appears over the eastern portion of the ocean basin, maintaining the summertime subtropical highs (Klein and Hartmann, 1993; Norris and Klein, 2000; Miyasaka and Nakamura, 2005; Wu et al., 2009). Marine boundary layer clouds over the eastern portion of the ocean basin can induce strong longwave radiative cooling aloft (Norris, 1998; Wu et al., 2009; Wei et al., 2017), which plays an important role in maintaining the subtropical highs (Liu et al., 2004; Miyasaka and Nakamura, 2005). In return, the equatorward flow along the eastern flank of the subtropical highs can generate cold air advection from high latitudes and coastal upwelling along the east coast of the ocean. This cools the local sea surface temperature and increases the static stability, which favours the growth of marine boundary layer clouds (Klein and Hartmann, 1993; Norris and Klein, 2000; Wei et al., 2017). For the South Indian Ocean, Miyamoto et al. (2018) have shown that in contrast to the summertime when low clouds are maximized off the west coast of Australia, the low clouds in wintertime are distributed more zonally across the basin. This is consistent with the seasonal variability in the displacement of longwave radiative cooling. It implies that strong longwave radiative cooling over the subtropical South Indian Ocean might be related to the radiative effect of low-level clouds. However, the feedback between low clouds, longwave radiative cooling, and the MH in winter is still unclear. Furthermore, the displacement of low clouds and the centre of the MH seems unique among other ocean basins where the strongest signals appear in the eastern portion. All these problems deserve further investigation to disclose the nature and origin of the MH.

ACKNOWLEDGEMENTS
This work is jointly supported by the National Natural Science Foundation of China (No. 42088101) and
National Key Research and Development Programs of China (No. 2019YFC1510400), National Natural Science Foundation of China (No. 41775043 and 42175019), and Guangdong Province Key Laboratory for Climate Change and Natural Disaster Studies (Grant 2020B1212060025).

AUTHOR CONTRIBUTIONS
Yuan Zhao: Conceptualization; formal analysis; methodology; writing – original draft; writing – review and editing. Zhiping Wen: Conceptualization. Xiuzhen Li: Conceptualization; funding acquisition; methodology; writing – review and editing. Ruidan Chen: Conceptualization. Guixing Chen: Conceptualization.

ORCID
Xiuzhen Li ◁ https://orcid.org/0000-0002-4672-0455
Ruidan Chen ◁ https://orcid.org/0000-0001-6207-8805

REFERENCES
Behera, S.K. and Yamagata, T. (2001) Subtropical SST dipole events in the southern Indian Ocean. Research Letters, 28, 327–330. https://doi.org/10.1029/2000GL011451.

Chen, G., Iwasaki, T., Qin, H. and Sha, W. (2014) Evaluation of the warm-season diurnal variability over East Asia in recent reanalyses JRA-55, ERA-Interim, NCEP CFSR, and NASA MERRA. Journal of Climate, 27, 5517–5537. https://doi.org/10.1175/JCLI-D-14-00005.1.

Chen, P., Hoerling, M.P. and Dole, R.M. (2001) The origin of subtropical anticyclones. Journal of the Atmospheric Sciences, 58, 1827–1835. https://doi.org/10.1175/1520-0469(2001)058<1827:TOOTSA>2.0.CO;2.

Cherchi, A., Ambrizzi, T., Behera, S., Freitas, A.C.V., Morioka, Y. and Zhou, T. (2018) The response of subtropical highs to climate change. Current Climate Change Reports, 4, 371–382. https://doi.org/10.1007/s40641-018-0114-1.

Cook, K.H. (2000) The south Indian convergence zone and interannual rainfall variability over southern Africa. Journal of Climate, 13, 3789–3804. https://doi.org/10.1175/1520-0442(2000)013<3789:TSICZA>2.0.CO;2.

Fauchereau, N., Trzaska, S., Richard, Y., Roucou, P. and Camberlin, P. (2003) Sea-surface temperature co-variability in the southern Atlantic and Indian Oceans and its connections with the atmospheric circulation in the Southern Hemisphere. International Journal of Climatology, 23, 663–677. https://doi.org/10.1002/joc.905.

Grise, K.M., Davis, S.M., Staten, P.W. and Adam, O. (2018) Regional and seasonal characteristics of the recent expansion of the tropics. Journal of Climate, 31, 6839–6856. https://doi.org/10.1175/JCLI-D-18-0060.1.

Han, X., Wei, F. and Chen, X. (2017) Influence of the anomalous patterns of the Mascarene and Australian highs on precipitation during the prerainy season in South China. Advances in Meteorology, 2017, 1–12. https://doi.org/10.1155/2017/6408029.

Harada, Y., Kamahori, H., Kobayashi, C., Endo, H., Kobayashi, S., Ota, Y., Onoda, H., Onogi, K., Miyaoka, K. and Takahashi, K. (2016) The JRA-55 reanalysis: Representation of atmospheric circulation and climate variability. Journal of the Meteorological Society of Japan, 94, 269–302. https://doi.org/10.2151/jmsj.2016-015.

Harrison, M.S.J. (1984) A generalized classification of South African summer rain-bearing synoptic systems. Journal of Climate, 4, 547–560. https://doi.org/10.1002/joc.3370040510.

Hermes, J.C. and Reason, C.J.C. (2005) Ocean model diagnosis of interannual coevolving SST variability in the South Indian and South Atlantic Oceans. Journal of Climate, 18, 2864–2882. https://doi.org/10.1175/JCLI3422.1.

Holton, J.R. and Hakim, G.J. (2012) An introduction to dynamic meteorology, 5th edition. Burlington, MA: Elsevier Academic Press, https://doi.org/10.1016/C2009-0-63394-8.

Hoskins, B.J. (1996) On the existence and strength of the summer subtropical anticyclones. Bulletin of the American Meteorological Society, 77, 1287–1292.

Hoskins, B.J. and Rodwell, M.J. (1995) A model of the Asian summer monsoon. Part I: the global scale. Journal of the Atmospheric Sciences, 52, 1329–1340. https://doi.org/10.1175/1520-0469(1995)052<1329:AMOTAS>2.0.CO;2.

Howard, E. and Washington, R. (2018) Characterizing the synoptic expression of the Angola low. Journal of Climate, 31, 7147–7165. https://doi.org/10.1175/JCLI-D-18-0017.1.

Klein, S.A. and Hartmann, D.L. (1993) The seasonal cycle of low stratiform clouds. Journal of Climate, 6, 1587–1606. https://doi.org/10.1175/1520-0442(1993)006<1587:TSICLC>2.0.CO;2.

Kobayashi, S., Ota, Y., Harada, Y., Ebita, A., Moriya, M., Onoda, H., Onogi, K., Kamahori, H., Kobayashi, C., Endo, H., Miyaoka, K. and Kiyotoshi, T. (2015) The JRA-55 reanalysis: General specifications and basic characteristics. Journal of the Meteorological Society of Japan, 93(1), 5–48. https://doi.org/10.2151/jmsj.2015-001.

Krishnamurti, T.N. and Bhalme, H.N. (1976) Oscillations of a monsoon system. Part I. Observational aspects. Journal of the Atmospheric Sciences, 33, 1937–1954. https://doi.org/10.1175/1520-0469(1976)033<1937:OOAMS>2.0.CO;2.

Lee, S.K., Mechos, C.R., Wang, C. and Neelin, J.D. (2013) Interhemispheric influence of the northern summer monsoons on southern subtropical anticyclones. Journal of Climate, 26, 10193–10204. https://doi.org/10.1175/JCLI-D-13-00106.1.

Li, X., Wen, Z., Chen, D. and Chen, Z. (2019) Decadal transition of the leading mode of interannual moisture circulation over East Asia-western North Pacific: bonding to different evolution of ENSO. Journal of Climate, 32, 289–308. https://doi.org/10.1175/JCLI-D-18-0356.1.

Li, X., Wen, Z. and Huang, W.R. (2020) Modulation of South Asian jet wave train on the extreme winter precipitation over Southeast China: comparison between 2015/16 and 2018/19. Journal of Climate, 33, 4065–4081. https://doi.org/10.1175/JCLI-D-19-0678.1.

Liu, Y., Liu, H., Liu, P. and Wu, G. (1999a) The effect of spatially nonuniform heating on the formation and variation of subtropical high. Part II: land surface sensible heating and East Pacific subtropical high. Acta Meteorologica Sinica, 57, 385–396 (in Chinese).

Liu, Y., Wu, G., Liu, H. and Liu, P. (1999b) The effect of spatially nonuniform heating on the formation and variation of subtropical high. Part I: spatially nonuniform heating on the formation and variation of subtropical high. Acta Meteorologica Sinica, 57, 525–538 (in Chinese).

Liu, Y., Wu, G., Liu, H. and Liu, P. (2001) Condensation heating of the Asian summer monsoon and the subtropical anticyclone in
the eastern hemisphere. *Climate Dynamics*, 17, 327–338. https://doi.org/10.1007/s003820000117.

Liu, Y., Wu, G. and Ren, R. (2004) Relationship between the subtropical anticyclone and diabatic heating. *Journal of Climate*, 17, 682–698. https://doi.org/10.1175/1520-0442(2004)017<0682:RTSAAA>2.0.CO;2.

Luo, H. and Yanai, M. (1984) The large-scale circulation and heat sources over the Tibetan Plateau and surrounding areas during the early summer of 1979. Part II: heat and moisture budgets. *Monthly Weather Review*, 112, 966–989. https://doi.org/10.1175/1520-0493(1984)112<0966:ITSSCA>2.0.CO;2.

Manatsa, D., Morioka, Y., Behera, S.K., Matarira, C.H. and Yamagata, T. (2014) Impact of Mascarene High variability on the East African “short rains.”. *Climate Dynamics*, 42, 1259–1274. https://doi.org/10.1007/s00382-013-1848-z

Miyamoto, A., Nakamura, H. and Miyasaka, T. (2018) Influence of the subtropical high and storm track on low-cloud fraction and itsseasonality over the Southern Indian Ocean. *Journal of Climate*, 31, 4017–4039. https://doi.org/10.1175/JCLI-D-17-0229.1.

Miyasaka, T. and Nakamura, H. (2005) Structure and formation mechanisms of the Northern Hemisphere summertime subtropical highs. *Journal of Climate*, 18, 5046–5065. https://doi.org/10.1175/JCLI3599.1.

Miyasaka, T. and Nakamura, H. (2010) Structure and mechanisms of the Southern Hemisphere summertime subtropical anticyclones. *Journal of the Southern Hemisphere*, 23, 2115–2130. https://doi.org/10.1175/2009JCLI3008.1.

Morioka, Y., Takaya, K., Behera, S.K. and Masumoto, Y. (2015) Local SST impacts on the summertime mascarene high variability. *Journal of Climate*, 28, 678–694. https://doi.org/10.1175/JCLI-D-14-00133.1.

Morioka, Y., Tozuka, T., Masson, S., Terray, P., Luo, J.J. and Yamagata, T. (2012) Subtropical dipole modes simulated in a coupled general circulation model. *Journal of Climate*, 25, 4029–4047. https://doi.org/10.1175/JCLI-D-11-00396.1.

Morioka, Y., Tozuka, T. and Yamagata, T. (2010) Climate variability in the southern Indian Ocean as revealed by self-organizing maps. *Climate Dynamics*, 35, 1075–1088. https://doi.org/10.1007/s00382-010-0843-x.

Norris, J.R. (1998) Low cloud type over the ocean from surface observations. Part II: geographical and seasonal variations. *Journal of Climate*, 11, 383–403. https://doi.org/10.1175/1520-0442(1998)011<0383:LCTOOT>2.0.CO;2.

Norris, J.R. and Klein, S.A. (2000) Low cloud type over the ocean from surface observations. Part III: relationship to vertical motion and the regional surface synoptic environment. *Journal of Climate*, 13, 245–256. https://doi.org/10.1175/1520-0442(2000)013<0245:LCTOOT>2.0.CO;2.

Reboita, M.S., Ambrizzi, T., Silva, B.A., Pinheiro, R.F. and da Rocha, R.P. (2019) The South Atlantic subtropical anticyclone: present and future climate. *Frontiers in Earth Science*, 7, 1–15. https://doi.org/10.3389/feart.2019.00008.

Rodwell, M.J. and Hoskins, B.J. (1996) Monsoons and the dynamics of deserts. *Quarterly Journal of the Royal Meteorological Society*, 122, 1385–1404. https://doi.org/10.1256/qmsqj.53407.

Rodwell, M.J. and Hoskins, B.J. (2001) Subtropical anticyclones and summer monsoons. *Journal of Climate*, 14, 3192–3211. https://doi.org/10.1175/1520-0442(2001)014<03192:SAASM>2.0.CO;2.

Rodwell, M.J. and Hoskins, B.J. (1996) Monsoons and the dynamics of deserts. *Quarterly Journal of the Royal Meteorological Society*, 122, 1385–1404. https://doi.org/10.1256/qmsqj.53407.

Rodwell, M.J. and Hoskins, B.J. (2001) Subtropical anticyclones and summer monsoons. *Journal of Climate*, 14, 3192–3211. https://doi.org/10.1175/1520-0442(2001)014<03192:SAASM>2.0.CO;2.

Rodwell, M.J. and Hoskins, B.J. (1996) Monsoons and the dynamics of deserts. *Quarterly Journal of the Royal Meteorological Society*, 122, 1385–1404. https://doi.org/10.1256/qmsqj.53407.

Rodwell, M.J. and Hoskins, B.J. (2001) Subtropical anticyclones and summer monsoons. *Journal of Climate*, 14, 3192–3211. https://doi.org/10.1175/1520-0442(2001)014<03192:SAASM>2.0.CO;2.
APPENDIX A.

Referring to the previous studies (Wu et al., 1999; Liu et al., 1999a, 1999b; Liu et al., 2001), the complete form of the vertical tendency equation can be deduced as follows:

The Ertel potential vorticity can be written as:

\[ P_E = \alpha \zeta_a \cdot \nabla \theta \]  \hspace{1cm} (A1)

where \( \zeta_a \) denotes the three dimensional absolute vorticity, \( \theta \) is potential temperature, and \( \alpha \) is specific volume, \( \nabla = \frac{\partial}{\partial x} \hat{i} + \frac{\partial}{\partial y} \hat{j} + \frac{\partial}{\partial z} \hat{k} \). Equation (A1) can also be rewritten as:

\[ \alpha(f + \zeta) = \frac{P_E}{\theta_z} - \alpha \nabla \cdot \nabla_h \theta \]  \hspace{1cm} (A2)

\( \theta_z = \frac{\partial \theta}{\partial z} \) is static stability, \( \vec{V} \) denotes the three dimensional velocity vector, \( \nabla_h = \frac{\partial}{\partial x} \hat{i} + \frac{\partial}{\partial y} \hat{j} \), \( \nabla \theta_z \) denotes the tilt of the isentropic surface. The thermodynamic parameter \( C_D = \alpha \nabla \times \vec{V} \cdot \nabla \alpha \theta \) represents the mapping of horizontal vorticity to vertical vorticity per unit mass. According to the theory of slantwise vorticity development (SVD) proposed by Wu and Liu (1999), when a parcel slides down a slantwise isentropic surface and the thermal parameter \( C_D \) is decreasing, its vertical vorticity develops. Thus, Equation (A2) can be written as:

\[ \frac{\partial \zeta}{\partial t} + \vec{V} \cdot \nabla \zeta + \beta \nu = (1 - K)(f + \zeta) \frac{\alpha}{P} - (f + \zeta) \frac{1}{\theta_z} \frac{d \theta}{dt} + \frac{1}{\theta_z} F_{\zeta} \cdot \nabla \theta + \frac{f + \zeta}{\theta_z} \frac{\partial Q}{\partial z} - \frac{1}{\theta_z} \frac{\partial v}{\partial z} \frac{\partial Q}{\partial x} + \frac{1}{\theta_z} \frac{\partial u}{\partial z} \frac{\partial Q}{\partial y} + \frac{1}{\alpha} \frac{d}{dt} \left( \frac{F_{\zeta}}{\theta_z} \right) - \frac{1}{\alpha} \frac{d}{dt} (C_D) \]  \hspace{1cm} (A6)

Let \( dA/dt \) be the total derivative of \( A \) in the inertial frame \( \left( \frac{dA}{dt} = \frac{\partial A}{\partial t} + \vec{V} \cdot \nabla A \right) \), Equation (A3) can be written as:

\[ \frac{d[\alpha(f + \zeta)]}{dt} = \frac{d}{dt} \left( \frac{P_E}{\theta_z} - C_D \right) \]  \hspace{1cm} (A4)

And Equation (A4) also can be:

\[ (f + \zeta) \frac{d\alpha}{dt} + \alpha \frac{d(f + \zeta)}{dt} = \frac{d}{dt} \left( \frac{P_E}{\theta_z} - C_D \right) \]  \hspace{1cm} (A5)

From the equation of state for an ideal gas (\( P = RT \)), (A5) can be written as:

\[ \frac{\partial \zeta}{\partial t} + \vec{V} \cdot \nabla \zeta + \beta \nu = (1 - K)(f + \zeta) \frac{\alpha}{P} - (f + \zeta) \frac{1}{\theta_z} \frac{d \theta}{dt} + \frac{1}{\alpha \theta_z} \frac{d}{dt} \left( \frac{P_E}{\theta_z} - C_D \right) \]  \hspace{1cm} (A6)

\( P \) represents the air pressure. \( K = R/C_p, R \) and \( C_p \) are the gas constant and specific heat capacity in dry air at constant pressure. \( \beta = \frac{d \theta}{dt} \) Considering the definition of Ertel potential vorticity (A1), Equation (A7) describing the variation of potential vorticity can be gained.

\[ \frac{dP_E}{dt} = a \vec{F}_{\zeta} \cdot \nabla \theta + \alpha \zeta_a \cdot \nabla Q \]  \hspace{1cm} (A7)

As a result, Equation (A6) can be rewritten as:

\[ a(f + \zeta) = \frac{P_E}{\theta_z} - C_D \]  \hspace{1cm} (A3)

where \( \vec{F}_{\zeta} \) denotes the friction dissipation, calculated as the residual from the three dimensional vorticity equation (Holton and Hakim, 2012), \( Q \) is diabatic heating rate. This is the complete form of the vertical vorticity tendency Equation (A8), in which, the relative contributions of dynamical and thermodynamical processes to the tendency of vertical vorticity can be evaluated.