Responses of the Summertime Subtropical Anticyclones to Global Warming

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ABSTRACT

Subtropical anticyclones dominate the subtropical ocean basins in summer. Using the multimodel output from phase 5 of the Coupled Model Intercomparison Project (CMIP5), the future changes of the subtropical anticyclones as a response to global warming are investigated, based on the changes in subsidence, low-level divergence, and rotational wind. The subtropical anticyclones over the North Pacific, South Atlantic, and south Indian Ocean are projected to become weaker, whereas the North Atlantic subtropical anticyclone (NASA) intensifies, and the South Pacific subtropical anticyclone (SPSA) shows uncertainty but is likely to intensify. Diagnostic analyses and idealized simulations suggest that the projected changes in the subtropical anticyclones are well explained by the combined effect of increased tropospheric static stability and changes in diabatic heating. Increased static stability acts to reduce the intensity of all the subtropical anticyclones, through the positive mean advection of stratification change (MASC) over the subsidence regions of the subtropical anticyclones. The pattern of change in diabatic heating is dominated by latent heating associated with changes in precipitation, which is enhanced over the western North Pacific under the “richest get richer” mechanism but is reduced over subtropical North Atlantic and South Pacific due to a local minimum of SST warming amplitude. The change in the diabatic heating pattern substantially enhances the subtropical anticyclones over the North Atlantic and South Pacific but weakens the North Pacific subtropical anticyclone.

1. Introduction

The atmospheric circulation over subtropical oceans is dominated by basin-scale anticyclones in summer, associated with subsidence, low-level divergence, and

anticyclonic wind curl (Fig. 1). The subtropical anticyclones are responsible for the formation of monsoons, subtropical deserts, and Mediterranean-type climate. Given its profound climatic impact, the formation mechanism for the subtropical anticyclone has been thoroughly studied (Ting 1994; Chen et al. 2001; Rodwell and Hoskins 2001; Seager et al. 2003; Liu et al. 2004). It was hypothesized that subtropical anticyclones are contributed to both by the descending branch of the Hadley circulation (Namias 1972; Dima and Wallace 2003) and

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FIG. 1. The mean state for boreal summer in Northern Hemisphere and for austral summer in Southern Hemisphere. (a) 925-hPa wind (m s^{-1}), (b) 925-hPa relative vorticity (10^{-6} s^{-1}), (c) 925-hPa divergence (10^{-6} s^{-1}), and (d) 700-hPa vertical velocity ($10^{-2} \text{ Pa s}^{-1}$). The vectors in (a) and the shading in (b)–(d) are based on the observation. In (b)–(d), the contours represent the MMM simulation. In each panel, the contour interval is the same as the shading interval indicated by the color bar, and the negative contours are dashed.

by zonal asymmetric diabatic heating associated with the land–sea thermal contrast (Rodwell and Hoskins 2001; Liu et al. 2004; Miyasaka and Nakamura 2010).

The long-term change of tropical atmospheric circulation under anthropogenic greenhouse gas (GHG) forcing has been widely studied. Based on coupled model simulations and theoretical prediction, GHG forcing may reduce the intensity of the Walker circulation (e.g., Held and Soden 2006; Lu et al. 2007; Vecchi and Soden 2007; Ma et al. 2012) and lead to a poleward expansion of Hadley circulation (Lu et al. 2007; Tao et al. 2016), but its effect on the intensity of Hadley circulation is under debate (Ma and Xie 2013; Su et al. 2014; Lau and Kim 2015; Tao et al. 2016). The decadal change of tropical circulation over the past few decades does not exactly follow the multimodel simulated response. During the past decades, the Hadley circulation has expanded poleward but its intensity change is uncertain (Hu and Fu 2007; Nguyen et al. 2013), and the intensity change of the Walker circulation is also under debate (L’Heureux et al. 2013; DiNezio et al. 2013;

McGregor et al. 2014; Bellomo and Clement 2015; Ma and Zhou 2016). The uncertainty in the decadal change for the past decades may result from the limitations in data (Held and Soden 2006) and metrics (DiNezio et al. 2013). The inconsistency between the observed decadal change and model-simulated response suggests that internal variability originating from the air–sea coupled climate system may play a key role in the observed decadal change (DiNezio et al. 2013; Garfinkel et al. 2015).

Compared with the tropical circulation, the possible response of the subtropical anticyclones to GHG forcing has received less attention. The western North Pacific subtropical high (WNPSH) is projected to become weaker in the midtroposphere but stays generally unchanged in the lower troposphere (He and Zhou 2015; He et al. 2015). However, the nature of the WNPSH differs considerably from the core regions of the subtropical anticyclones (Liu and Wu 2000). Li et al. (2012, 2013) indicated an intensification of the subtropical anticyclones under GHG forcing, based on the increase of the streamfunction near the centers of the subtropical

anticyclones. However, the streamfunction ψ is obtained by solving the Poisson equation with relative vorticity ζ , that is, $\nabla^2\psi = \zeta$. Based on its definition, the streamfunction at any grid point depends on global relative vorticity, and the change in local streamfunction may result from changes in either local wind or remote winds. As shown by the example in Fig. S1 of the supplemental material, a small change in zonal wind over the Antarctic could induce a substantial change in streamfunction over the subtropical North Pacific, without any change in wind over North Pacific. In fact, local rotational wind is determined by the horizontal gradient of streamfunction but not related to the absolute magnitude of streamfunction. Therefore, the increase of streamfunction near the center of the anticyclone may not indicate intensification of the anticyclone. It requires a metric based on local circulation to investigate the responses of the subtropical anticyclones under GHG forcing.

The mechanism for the responses of the tropical and subtropical circulation to GHG forcing involves atmospheric internal dynamics and ocean–atmosphere coupled processes. With the increase of water vapor content, the static stability of troposphere increases through moist adiabatic adjustment under GHG forcing (Knutson and Manabe 1995). The increased static stability induces an additional adiabatic heating (cooling) over the climatological descending (ascending) zones, and reduces the mean state descending (ascending) motion (Ma et al. 2012). Besides stabilization, the pattern of change in atmospheric diabatic heating is also important for the changes in atmospheric circulation (Li et al. 2012; Ma et al. 2012; Qu and Huang 2016). The pattern of change in diabatic heating is usually dominated by latent heating associated with precipitation (Ma et al. 2012), and the change in precipitation pattern is modulated by the SST warming pattern, characterized by enhanced (suppressed) precipitation over the region with relatively strong (weak) SST warming (Xie et al. 2010; Chadwick et al. 2013; Ma and Xie 2013).

This study aims at answering the following two questions. 1) Do the subtropical anticyclones intensify under GHG forcing as revealed by local wind and other variables derived directly from local wind? 2) What is the fundamental mechanism responsible for the responses of the subtropical anticyclones to GHG forcing? We address these questions based on the output of 30 models from phase 5 of the Coupled Model Intercomparison Project (CMIP5; Taylor et al. 2012) and test our hypothesized mechanism by using a linear baroclinic model (Watanabe and Kimoto 2000). The rest of this article is organized as follows. The model, data, and methods are introduced in section 2, and the

observational and model-simulated present-day climatology of the subtropical anticyclones is documented in section 3. The projected changes of the subtropical anticyclones by the CMIP5 models are addressed in section 4, and a possible mechanism for the projected future changes is discussed in section 5. Finally, a summary is presented in section 6.

2. Model, data, and methods

CMIP5 collects a large group of models that perform experiments under identical external forcing (GHG, aerosols, etc.). In the historical experiment, the coupled models are forced by the observed historical external forcing from 1850 to 2005, and the 1950–99 period of the historical experiment is adopted here as a baseline of present-day climate. The representative concentration pathway 8.5 (RCP8.5) experiment is adopted as a projection of future climate. The RCP8.5 is a high concentration pathway toward 8.5 W m^{-2} radiative forcing in the year of 2100, equivalent to 1370 ppm CO_2 concentration (van Vuuren et al. 2011). In total 30 available models are adopted for analysis, and the names of the models are listed in Table S1 of the supplemental material. Only the monthly output of the first ensemble member (r1i1p1) of each model is analyzed, and all the model data are bilinearly interpolated onto a common $1^\circ \times 1^\circ$ horizontal grid.

To clearly demonstrate the present-day climatology and to evaluate the performance of the models in simulating the present-day climate, the historical experiment of the model is compared with the reanalysis data for the period of 1950–99. The reanalysis dataset adopted in this study is from the National Centers for Environmental Prediction (NCEP) and National Center for Atmospheric Research (NCAR) (Kalnay et al. 1996), referred to herein as “observation.” Following Li et al. (2012, 2013), we focus on June–August (JJA) for the Northern Hemisphere and December–February (DJF) for the Southern Hemisphere throughout this study.

The mean state of the 2050–99 period in RCP8.5 experiment is compared with the mean state of the 1950–99 period in the historical experiment, to address the projected future change. Intermodel consistency of the projected change by the multimodel mean (MMM) is evaluated against the uncertainty caused by model bias and internal variability. The intermodel consistency is calculated as the percentage of models that agree on the sign of change with the MMM. For a vector field (e.g., wind), the intermodel consistency is defined as the average of the intermodel consistencies between the zonal and meridional components. Under the assumption of independence among models, a threshold of 68%

intermodel consistency is equivalent to 95% statistical significance as determined by the Student's t test (Power et al. 2012). Since 30 models are adopted here, we use a slightly stricter threshold of 70% intermodel consistency. The projected change is considered as significant if at least 21 of the 30 models agree on the sign of change.

A linear baroclinic model (LBM) is adopted to test the response of atmospheric circulation to prescribed heating. The LBM is a primitive equation model developed by Watanabe and Kimoto (2000), and it was previously used by Ma et al. (2012) and Qu and Huang (2016) to diagnose the responses of Walker circulation and South Asian high to prescribed heating under global warming scenarios. In this work, the LBM is run with triangular truncation at wavenumber 42 (T42) in the horizontal with 20 vertical levels in sigma coordinates. The forcing field diagnosed from the output of CMIP5 models is interpolated vertically from pressure levels to sigma levels of the LBM and interpolated horizontally from the $1^\circ \times 1^\circ$ grid to a Gaussian grid for T42 of LBM. We linearize this model at the JJA (DJF) mean state of the historical experiment simulated by the MMM, when we prescribe forcing on the Northern (Southern) Hemisphere.

3. Present-day climatology

As shown in Fig. 1a and many previous studies (e.g., Rodwell and Hoskins 2001; Miyasaka and Nakamura 2010), the subtropical ocean basins are regulated by anticyclones in summer. In boreal summer (JJA), there are two subtropical anticyclones in the Northern Hemisphere, including the North Pacific subtropical anticyclone (NPSA) and the North Atlantic subtropical anticyclone (NASA). In austral summer (DJF), there are three subtropical anticyclones in the Southern Hemisphere, including the South Pacific subtropical anticyclone (SPSA), the South Atlantic subtropical anticyclone (SASA), and the south Indian Ocean subtropical anticyclone (SISA). As seen from Fig. 1a, the centers of these anticyclones are located at about 30°N and 30°S , and the scopes for the subtropical anticyclones range from 5° to 50°N and from 5° to 50°S .

Since the subtropical anticyclones are characterized by anticyclonic wind curl, subsidence and low-level divergence in the troposphere, we measure the subtropical anticyclones with 925-hPa relative vorticity (Vot925), 925-hPa divergence (Div925), and 700-hPa vertical velocity (W700). For relative vorticity and divergence, we focus on 925 hPa in order to be comparable with Li et al. (2012, 2013). We use the central difference formula to calculate relative vorticity and divergence, so that the calculated vorticity and divergence depend on only local

(not global) wind. We measure the subsidence of subtropical anticyclones at 700 hPa since the strongest subsidence is located at near 700 hPa (Rodwell and Hoskins 2001; see also the contours in Fig. 4).

As shown in Fig. 1b, negative Vot925 dominates the subtropical North Pacific and subtropical North Atlantic in boreal summer, while positive Vot925 dominates the subtropical South Pacific, subtropical South Atlantic, and subtropical south Indian Ocean in austral summer. The negative Vot925 in Northern Hemisphere and positive Vot925 in Southern Hemisphere are both associated with anticyclonic rotational wind. Unlike the anticyclonic relative vorticity ranging across the subtropical ocean basins, the 925-hPa divergence and 700-hPa subsidence are located at the eastern parts of the subtropical oceans (Figs. 1c,d). Comparing the contours with the shading in Figs. 1b–d, it is clear that the mean state features of the subtropical anticyclones in the observation are well captured by the MMM simulation, adding to our confidence to further explore the model projected changes.

4. Projected changes of the subtropical anticyclones

The MMM projected changes in 925-hPa wind, Vot925, Div925, and W700 under RCP8.5 are displayed in Fig. 2. Characteristic contours of the climatology in the historical and RCP8.5 experiments are shown in Figs. 2b–d, for the convenience to compare the projected change with the mean state and to check the possible location change of the subtropical anticyclones. The Vot925 contours of $-6 \times 10^{-6} \text{ s}^{-1}$ in the Northern Hemisphere and $+6 \times 10^{-6} \text{ s}^{-1}$ in the Southern Hemisphere are shown in Fig. 2b. The $1 \times 10^{-6} \text{ s}^{-1}$ contours of Div925 are shown in Fig. 2c, and the $2 \times 10^{-2} \text{ Pa s}^{-1}$ contours of W700 are shown in Fig. 2d.

Over the subtropical North Pacific, an anomalous cyclone is seen, characterized by westerly wind anomaly within 20° – 30°N and easterly wind anomaly along 45°N (Fig. 2a), suggesting a weakened NPSA. Figure 1a of Shaw and Voigt (2015) shows that the increase of streamfunction at the center of NPSA is weaker than around it, and our result is consistent with theirs in the sense of horizontal gradient of streamfunction. Consistent with the change in wind, positive Vot925 anomalies are seen within 25° – 40°N over the subtropical North Pacific (Fig. 2b), opposing the climatological negative Vot925. The projected changes in Div925 and W700 over the subtropical eastern North Pacific are characterized by negative anomalies (Figs. 2c,d), opposite to their positive climatology. The decreased subsidence, low-level divergence, and anticyclonic rotational wind

FIG. 2. The MMM projected change calculated as the difference between the 2050–99 mean state in the RCP8.5 simulation and the 1950–99 mean state in the historical experiment. (a) 925-hPa wind (m s^{-1}), (b) 925-hPa relative vorticity (10^{-6} s^{-1}), (c) 925-hPa divergence (10^{-6} s^{-1}), and (d) 700-hPa vertical velocity ($10^{-2} \text{ Pa s}^{-1}$). The shaded areas in (a) and the stippled areas in (b)–(d) indicate that the intermodel consistency is no less than 70%. The scope of the subtropical anticyclone over each ocean basin for each variable (Vot925, Div925, and W700) is marked in (b)–(d) for the historical (black contours) and RCP8.5 (white contours) experiments, which is defined as the region with the corresponding variable exceeding a threshold. The threshold is $\pm 6 \times 10^{-6} \text{ s}^{-1}$ for Vot925 (negative for Northern Hemisphere and positive for Southern Hemisphere), $1 \times 10^{-6} \text{ s}^{-1}$ for Div925, and $2 \times 10^{-2} \text{ Pa s}^{-1}$ for W700.

consistently suggest a weakened NPSA under the warming climate. Comparing the characteristic contours between the RCP8.5 and historical experiments in Figs. 2b–d, it is clear that the location of the NPSA generally stays unchanged under GHG forcing.

Over the subtropical North Atlantic, however, the projected change in wind is characterized by an anomalous anticyclone near the eastern coast of North America, accompanied by a cyclonic curl off the western coast of North Africa (Fig. 2a), suggesting a northwestward shift of the NASA. Within the region enclosed by the $-6 \times 10^{-6} \text{ s}^{-1}$ contour of Vot925, the Vot925 increases on the southeastern flank and decreases on the northwestern flank (Fig. 2b). The Div925 and W700 are projected to decrease on the southeast flank but increase on the northwest flank of NASA (Figs. 2c,d). These features suggest a northwestward shift of NASA, which is confirmed by comparing the white contours with the black contours in Figs. 2b–d. Shaw and Voigt (2015)

concluded there was a westward shift of the NASA based on the changes of streamfunction, and many previous studies claimed a poleward expansion of the Hadley circulation under GHG forcing (e.g., Lu et al. 2007; Tao et al. 2016). Our result about the northwestward shift of NASA is consistent with these previous studies. However, it requires further quantitative evaluation on the intensity change of the NASA.

Over the South Pacific, the changes in low-level wind are characterized by a southeasterly wind anomaly from the southern tip of South America to the equatorial central Pacific, with an anomalous cyclone at the eastern subtropical South Pacific and an anomalous anticyclone at the central subtropical South Pacific (Fig. 2a). As seen from the changes in Vot925, Div925, and W700, the SPSA seems to be weakened on its eastern and northern flanks but enhanced on its western and southern flanks (Figs. 2b–d). Comparing the characteristic contours between the RCP8.5 and historical experiment

simulations in Figs. 2b–d, it is clear that the SPSA shifts southwestward, and its intensity change still requires further quantitative evaluation.

Over the South Atlantic and South Indian Oceans, the subtropical anticyclones show a common tendency toward reduced intensity and southward displacement. In Fig. 2a, easterly wind anomalies are seen across South Pacific and south Indian Ocean at about 40°S, accompanied by anomalous cyclones within 10°–40°S and anomalous anticyclones within 40°–55°S. Across the subtropical ocean basins around 30°S, the changes in Vot925, Div925, and W700 are mostly opposite to the climatology except at the poleward edges of SASA and SISA (Figs. 2b–d). The characteristic contours of SASA and SISA for RCP8.5 experiment are displaced slightly southward compared with the historical experiment (Figs. 2b–d). All of these projected changes suggest reduced intensity and a slightly poleward shift of the SASA and SISA, consistent with the poleward expansion of the Hadley cell (Lu et al. 2007; Tao et al. 2016).

To quantitatively evaluate the change in intensity of each subtropical anticyclone, we adopt three indices based on the regional integrated quantities for each anticyclone. Supposing $\mathbf{V} = (u, v)$ is the wind vector, $\zeta = \partial v/\partial x - \partial u/\partial y$ is relative vorticity, and $\delta = \partial u/\partial x + \partial v/\partial y$ is divergence, the Stokes law can be written as

$$\oint_{\partial\Omega} \mathbf{V} \cdot d\mathbf{r} = \int_{\Omega} \zeta dS \quad \text{and} \quad (1)$$

$$\oint_{\partial\Omega} \mathbf{V} \cdot d\mathbf{n} = \int_{\Omega} \delta dS. \quad (2)$$

Here, Ω is a domain on the sphere, and $\partial\Omega$ is the boundary of Ω . Also, $\mathbf{r}$ is the unit tangent vector along $\partial\Omega$, and $\mathbf{n}$ is the outward unit normal vector. The Stokes law states that the intensity of rotational wind along the boundary of the domain equals the domain integrated vorticity, and the intensity of divergent flow across the boundary of the domain equals the domain integrated divergence. For each subtropical anticyclone, the Vot925 index is defined as the regional integrated Vot925 enclosed by the characteristic contour over the corresponding ocean basin shown in Fig. 2b. For example, the Vot925 index for NPSA in the historical experiment (RCP8.5) scenario is defined as the regional integrated Vot925 over the domain enclosed by the black (white) contour over North Pacific in Fig. 2b. Similarly, the Div925 index for each anticyclone in the historical experiment (RCP8.5) simulation is defined as the regional integrated Div925 over the domain enclosed by the black (white) contour in Fig. 2c over the corresponding ocean basin, and the W700 index for each anticyclone in the historical experiment (RCP8.5)

simulation is defined as the regional integrated W700 over the domain enclosed by the black (white) contour in Fig. 2d over each ocean basin. Slightly different domains are used in calculating the indices for the historical experiment and RCP8.5 simulations in order to avoid the impact of the location change of the anticyclone on the intensity index. For example, a pure location change of the anticyclone does not alter the above intensity indices, but it may result in a spurious change in the intensity index if the intensity index is defined over an identical domain for the historical experiment and RCP8.5 runs. The sensitivity of the results to the choice of the characteristic contour was tested by using slightly different characteristic contours, and the results are robust (figure not shown).

The percentages of changes in these three indices for the five anticyclones are shown in Figs. 3a–c, and the projected changes scaled by the tropical (30°S–30°N) averaged surface warming are shown in Figs. 3d–f. The three indices consistently show an increased intensity of NASA and reduced intensity of NPSA, SASA, and SISA under RCP8.5, agreed upon by more than 70% of the models. The NPSA has the greatest amplitude of reduced intensity. These quantitative results are consistent with the visual impression from Fig. 2. For SPSA, the Vot925 and Div925 indices show an intensification agreed upon by more than 70% of the individual models, but the MMM projected change in W700 index is slightly smaller than 0, indicating uncertainty. The projected changes in the arithmetic average of the five subtropical anticyclones are negative and agreed upon by more than 70% of the models, and the reduction of the W700 indices is generally stronger than Vot925 and Div925 indices for most subtropical anticyclones and their arithmetic average. The magnitude of the MMM projected change (increase or decrease) hardly exceeds 5% K^{−1} of warming for all the five subtropical anticyclones, as agreed upon by the three indices (Figs. 3d–f).

The above evidence based on multiple variables corroborates an intensification of the NASA, reduced intensity in the NPSA, SASA, and SISA, and possibly intensification of the SPSA despite uncertainty. To check the robustness of the results to the choice of pressure level, Fig. 4 shows the vertical profile of MMM projected changes in the subtropics, in terms of the 25°–35°N average for the Northern Hemisphere and the 35°–25°S average for the Southern Hemisphere. It is clear that the projected changes are generally homogeneous from 850 to 1000 hPa for relative vorticity and divergence, and the projected changes in vertical velocity generally agree in sign at all pressure levels through the troposphere. The changes in 1000-hPa wind and 500-hPa vertical velocity (Fig. S2 in the supplemental material)

FIG. 3. The projected percentage change of the (a) Vot925, (b) Div925, and (c) W700 indices for the five subtropical anticyclones and their arithmetic average (%). (d)–(f) As in (a)–(c), but scaled by the amplitude of tropical mean (30°S – 30°N) surface warming ($\% \text{ K}^{-1}$). The thick blue bar indicates the MMM, and the thin black error bar indicates the 30%–70% range among the individual models.

show similar patterns to the changes in 925-hPa wind and 700-hPa vertical velocity. Therefore, it is reasonable to measure the low-level circulation at 925 hPa, and to measure the vertical motion by using W700.

Many previous studies suggested increasing aridity on the eastern flanks of subtropical anticyclones and expansion of the subtropical dry zones under GHG forcing (Held and Soden 2006; Seager et al. 2010; Cai et al. 2012). The fact of drying subtropics does not contradict the reduced intensity of the three of the five subtropical anticyclones. Precipitation is regulated by both a thermodynamic factor (moisture) and a dynamic factor (circulation). Previous studies concluded that the drying subtropics results from the thermodynamic factor (Seager et al. 2010), while the effect of the dynamic process (i.e., the changes in circulation) is to moisten the subtropical dry zones (e.g., Seager et al. 2010; Bony et al. 2013). Therefore, our conclusion of the reduced intensity of these three subtropical

anticyclones does not contradict the changes in precipitation and aridity.

5. Mechanism for the responses of the subtropical anticyclones

The zonal asymmetric diabatic heating is responsible for the formation of the subtropical anticyclones (Rodwell and Hoskins 2001; Liu et al. 2004). As argued by Li et al. (2012, 2013), the zonal asymmetric diabatic heating pattern intensifies under the warming climate, in favor of enhanced subtropical anticyclones. However, as seen in section 4, three of the five subtropical anticyclones are projected to become weaker. There must be some other mechanism that acts to weaken the subtropical anticyclones. It was proposed that the increased static stability of troposphere acts to weaken the Hadley and Walker cells (Ma et al. 2012; Seo et al. 2014) and the South Asian monsoon circulation in the upper

FIG. 4. The longitude–height profile of the MMM projected changes in (a),(d) relative vorticity (10^{-6} s^{-1}); (b),(e) divergence (10^{-6} s^{-1}); and (c),(f) vertical velocity ($10^{-2} \text{ Pa s}^{-1}$). The shading is the difference between the RCP8.5 and historical experiments, and the contours are the mean state in the historical experiment with negative contours dashed. The meridional average between 25° and 35°N in JJA is shown in (a)–(c), and the meridional average between 25° and 35°S in DJF is shown in (d)–(f).

troposphere (Ma and Yu 2014). The increased static stability is a robust feature of the troposphere in tropics and midlatitudes under the warming climate (Knutson and Manabe 1995; Frierson 2006). It will be interesting to investigate how the increased static stability impacts the subtropical anticyclones in combination with the changes in diabatic heating.

a. Diagnostic analyses

By taking the long-term average, the temperature equation under pressure coordinate can be written as

$$-\mathbf{V} \cdot \nabla T + S\omega + Q = 0, \quad (3)$$

where $\mathbf{V}$, T , S , ω , and Q represent horizontal wind, temperature, static stability, vertical velocity, and diabatic heating, respectively. Equation (3) indicates that the horizontal temperature advection $-\mathbf{V} \cdot \nabla T$, adiabatic

heating $S\omega$, and diabatic heating Q should be in balance for the mean state. By making difference between future climate and present-day climate, we obtain

$$-\mathbf{V}' \cdot \nabla \bar{T} + \bar{S}\omega' - \bar{\mathbf{V}} \cdot \nabla T' + S'\bar{\omega} - \mathbf{V}' \cdot \nabla T' + S'\omega' + Q' + \text{Res} = 0. \quad (4)$$

Here the variables with an overbar stand for the mean state of the historical experiment, and the variables with a prime stand for the differences between the RCP8.5 and historical experiments. The changes of temperature and stability (T' and S' , respectively) could either act as a forcing on the circulation or be forced by the circulation. For example, increased static stability could either be forced by the enhanced descending motion through dry adiabatic subsidence or reduce the intensity of the descending motion by

enhancing the adiabatic heating in descending region (Ma et al. 2012). Therefore, it is necessary to break T and S' into zonal mean and zonal deviation to separate local dynamic feedback from the large-scale thermal forcing (i.e., $T' = T'_m + T'_e$ and $S' = S'_m + S'_e$). Thus Eq. (4) becomes

$$-\mathbf{V}' \cdot \nabla \bar{T} + \bar{S}'\omega' - \bar{\nabla} \cdot \nabla T'_m - \bar{\nabla} \cdot \nabla T'_e + S'_m \bar{\omega} + S'_e \bar{\omega} - \mathbf{V}' \cdot \nabla T' + S'\omega' + Q' + \text{Res} = 0. \quad (5)$$

There are 10 terms in Eq. (5). The first two terms indicate the effect of changes in circulation ($-\mathbf{V}' \cdot \nabla \bar{T} + \bar{S}'\omega'$). The following four terms represent the changes in thermodynamic states ($-\bar{\nabla} \cdot \nabla T'_m - \bar{\nabla} \cdot \nabla T'_e + S'_m \bar{\omega} + S'_e \bar{\omega}$), followed by higher-order terms due to changes in both thermodynamic and dynamic variables ($-\mathbf{V}' \cdot \nabla T' + S'\omega'$), change in diabatic heating Q' , and the residual term (Res). Here, $S'_m \bar{\omega}$ is induced by changes in the zonal mean static stability and is equivalent to the mean advection of stratification change (MASC) proposed by Ma et al. (2012). Note that $-\bar{\nabla} \cdot \nabla T'_m$ is induced by changes in zonal mean temperature gradient (since the zonal mean is taken, this term is actually the effect of changed zonal mean meridional temperature gradient). The terms $-\bar{\nabla} \cdot \nabla T'_m$ and $S'_m \bar{\omega}$ can be considered as the thermal forcing on the circulation due to changes in zonal mean thermal states. The terms $-\bar{\nabla} \cdot \nabla T'_e$ and $S'_e \bar{\omega}$ represent the effect of changes in local thermal states, but they include the dynamic feedback of circulation changes on local thermal states. Diagnosis based on Eq. (5) helps to identify how the balance between the dynamic fields, thermodynamic fields, and diabatic heating is achieved under GHG forcing, and may give us some clue as to the mechanism for the circulation changes.

Based on CMIP5 model output, the 10 terms in Eq. (5) are calculated for each model separately, and the MMM of the 30 models is obtained. The vertical averages within 300–925 hPa for the 10 terms in Eq. (5) are shown in Fig. 5. Here, the diabatic heating Q and its change Q' are computed following Yanai et al. (1973). For the dynamic terms (Figs. 5a,b), $\bar{S}'\omega'$ dominates over the subtropics near 30°N and 30°S, while $-\mathbf{V}' \cdot \nabla \bar{T}$ has a substantial contribution on the poleward flanks of subtropical anticyclones. The dynamic changes are balanced mainly by MASC and Q' . The MASC term (Fig. 5e) is positive over the descending regions of the subtropical anticyclones, offsetting a large fraction of $\bar{S}'\omega'$ (Fig. 5b). By definition, the effect of positive MASC is to enhance the ascending motion or to weaken the descending motion (Ma et al. 2012) and thus to weaken the subtropical anticyclones. The Q' term over the subtropical Northern Hemisphere is characterized by a zonal one-wave pattern, with increased Q' over the

western North Pacific and decreased Q' over North Atlantic; Q' also decreases substantially over the subtropical South Pacific (Fig. 5i). Compared with MASC and Q' , the other terms contribute less to balance the dynamic changes in the subtropics (Figs. 5c,d,f–h), and the residual term is small (Fig. 5j).

To assess the contribution of changes in latent heating LH' to Q' , we computed the latent heating for the historical and RCP8.5 experiments based on the apparent moisture sink, following Yanai et al. (1973). The 300–925-hPa-averaged LH' and the change in residual heating RE' (calculated as $Q' - \text{LH}'$) are shown in Figs. 6a and 6b, respectively. The spatial pattern of LH' resembles Q' , including the enhanced heating over western North Pacific and decreased heating over North Atlantic and South Pacific (Fig. 6a). Compared with LH' , RE' is smaller and generally opposite in sign with LH' in the subtropics (Fig. 6b). The spatial pattern of LH' resembles the change in precipitation (Fig. 6c). Therefore, the pattern of projected changes in diabatic heating is dominated by the moist processes associated with latent heat release (Ma et al. 2012), and possibly the cloud-related radiative effects (Ceppi and Hartmann 2016).

As noted by Zhang and Li (2017), the substantially increased rainfall over the western North Pacific under GHG forcing conforms to the “richest get richer” mechanism, as the most abundant rainfall in the subtropical Northern Hemisphere is located at western North Pacific. Besides, the SST warming pattern plays a crucial role in the response of rainfall pattern (Xie et al. 2010; Ma and Xie 2013). Figure 6d shows the MMM projected change in SST, with the zonal mean warming removed to highlight the spatial pattern of change. It is clear that the SST warming over the subtropical North Atlantic is much weaker than North Pacific and the zonal mean, which accounts for the decreased rainfall over North Atlantic and the zonal one-wave pattern of the change in precipitation shown in Fig. 6c. This interbasin difference in SST warming amplitude under GHG forcing may originate from the weakened Atlantic meridional overturning circulation (Leloup and Clement 2009; Lee et al. 2011), or the enhanced wind–evaporation–SST feedback over North Atlantic (Xie et al. 2010). The suppressed precipitation over the subtropical central South Pacific is also associated with a local minimum of SST warming (Figs. 6c,d). These results suggest that the pattern changes in the subtropical precipitation and diabatic heating are modulated by the mean state rainfall and SST warming pattern, and the relatively weak warming over subtropical North Atlantic and South Pacific may be responsible for the enhanced NASA and SPSA.

FIG. 5. The magnitude of the 10 terms in Eq. (5) as diagnosed from the MMM of the 30 models (K day^{-1}). The region where the projected sign of change is agreed upon by more than 70% of the models is stippled.

b. Idealized model simulations

To examine the effect of MASC and Q' on the subtropical anticyclones, we run LBM simulations forced by the three-dimensional pattern of MASC and Q' , respectively. The LBM is forced with MASC and Q' within the latitudinal bands of 10° – 45° N and 10° – 45° S from 925 to 300 hPa, and the vertical structure of MASC and Q' is

shown in Fig. 7. It is evident that MASC generally keeps its sign in the vertical, and it reaches its maximum at the upper troposphere (Figs. 7a,b). The enhancement of Q' over western North Pacific maximizes in the upper troposphere near 300 hPa, whereas the decreases of Q' over the subtropical North Atlantic and South Pacific maximize in the midtroposphere (Figs. 7b,d). The LBM simulations forced by MASC and Q' for the Northern

FIG. 6. The MMM projected change in (a) latent and (b) residual heating (K day^{-1}), (c) precipitation (mm day^{-1}), and (d) SST pattern (K). The zonal mean warming of SST is removed in (d). The region where the sign of projected change is agreed upon by at least 70% of the models is stippled.

and Southern Hemispheres are all integrated for 50 days to obtain steady responses.

The LBM simulated responses of wind and relative vorticity at 925 hPa are shown in Figs. 8a–c. To quantitatively evaluate the LBM simulated response of the intensity of the subtropical anticyclones, we add the pattern of LBM responses of Vot_{925} shown in Figs. 8a–c to the mean state of the historical experiment simulated by the MMM of CMIP5 models, and obtain a “synthetic future climate” to mimic the RCP8.5 experiment. The Vot_{925} indices for the subtropical anticyclones are calculated based on this synthetic future climate by integrating the Vot_{925} over the domains enclosed by the white contours in Fig. 2b. The percentage change of the Vot_{925} indices in synthetic future climate relative to the historical experiment are shown in Fig. 8d.

The circulation response to MASC at the subtropics is characterized by anomalous cyclones and cyclonic vorticity over the five subtropical oceans (Fig. 8a). The response pattern is generally opposite to their climatology although displaced slightly equatorward (see Figs. 1a,b). The quantitative result shows that the effect of MASC is to reduce the intensity of all the subtropical anticyclones (blue bars in Fig. 8d). Thus MASC is the mechanism

connecting the increased zonal mean static stability and weakened subtropical anticyclones.

In response to Q' forcing, there is an anomalous cyclone over subtropical western North Pacific and an anomalous anticyclone over subtropical North Atlantic (Fig. 8b). These circulation anomalies are Rossby wave responses to the positive heating anomalies over the western North Pacific and negative heating anomaly over the North Atlantic, respectively (Gill 1980). As a response to Q' forcing, a southeasterly wind anomaly appears from the southern tip of South America to equatorial central Pacific, associated with an anomalous cyclone (anticyclone) on its east (west) (Fig. 8b). The quantitative result shows that change in diabatic heating substantially weakens the NPSA but enhances the NASA and SPSA, while its effect is weak on the SASA and SISA (red bars in Fig. 8d).

The sum of the LBM responses to MASC and Q' , as shown in Fig. 8c, well reconstructs the MMM projected changes shown in Fig. 2. The sum of LBM responses to MASC and Q' is characterized by anomalous cyclones over the subtropical North Pacific, subtropical South Atlantic, and subtropical south Indian Ocean (Fig. 8c), resembling the projected changes by CMIP5 models

FIG. 7. The longitude–height profile of (a),(c) MASC and (b),(d) Q' (K day^{-1}). The meridional average between 10° and 45°N in JJA is shown in (a),(b), and the meridional average between 10° and 45°S in DJF is shown in (c),(d). The region where the sign of the MMM projected change is agreed upon by more than 70% of the models is stippled.

(Figs. 2a,b). The sum of responses to MASC and Q' is characterized by an anomalous cyclone (anticyclone) on the southeast (northwest) flank of NASA, and an anomalous cyclone (anticyclone) over the subtropical eastern (central) South Pacific. The combined effect of MASC and Q' captures the decreased intensity in the NPSA, SASA, and SISA, and increased intensity in the NASA and SPSA (black bars in Fig. 8d). The LBM-simulated decrease in NPSA intensity is the strongest among the five subtropical anticyclones, consistent with the CMIP5 model projection.

As demonstrated by LBM simulations, the effect of MASC is to reduce the intensity of all the subtropical anticyclones. The effect of Q' also plays an important role in reducing the intensity of the NPSA and increasing the intensity of the NASA and SPSA. The combined effect of MASC and Q' is to weaken the NPSA, SASA, and SISA but to enhance the NASA and SPSA. The intensity of the NPSA decreases the most, because of the cooperation of MASC and Q' .

6. Summary

In this study, we investigated the responses of the summertime subtropical anticyclones to global warming by comparing the RCP8.5 experiment with the historical experiment of the 30 CMIP5 models. To find out the mechanisms responsible for the projected future changes, diagnostic analysis based on the thermodynamic equation was performed, and the atmospheric circulation response

to prescribed forcing was tested with an LBM. The major conclusions are summarized as follows.

- 1) Based on the projected changes in 925-hPa wind, 925-hPa relative vorticity, 925-hPa divergence, and 700-hPa vertical velocity, the CMIP5 model simulation under the RCP8.5 scenario projects reduced intensity for the subtropical anticyclones over the North Pacific, South Atlantic, and south Indian Ocean. The North Atlantic subtropical anticyclone is projected to intensify, and the South Pacific subtropical anticyclone is also likely to intensify despite uncertainty. The intensity changes of the subtropical anticyclones are generally less than $5\% \text{ K}^{-1}$, as projected by the MMM of CMIP5 models. Besides intensity change, the subtropical anticyclones shift poleward slightly under GHG forcing except for the NPSA. The NASA and SPSA shift both poleward and westward.
- 2) Under GHG forcing, the projected changes in the subtropical anticyclones are dominated by the combined effect of increased tropospheric static stability and changes in diabatic heating. With increased static stability forced by GHG-induced global warming, the subsidence over the subtropics could produce an extra adiabatic heating, that is, the mean advection of stratification change (MASC). Diagnoses and LBM simulations suggest that the increased static stability acts to weaken all of the subtropical anticyclones through MASC. The projected weakening of SASA

FIG. 8. LBM responses of 925-hPa wind and relative vorticity (10^{-6} s^{-1}) to prescribed (a) MASC and (b) Q' ; (c) the sum of responses (a) and (b). (d) The percentage change in the Vot925 indices of the subtropical anticyclones based on the synthetic future climate relative to the mean state in the historical experiment, where the synthetic future climate is obtained by adding the LBM responses to the mean state in the historical experiment to mimic the mean state of the RCP8.5 experiment. The blue, red, and black bars stand for the intensity responses of subtropical anticyclones to MASC, Q' , and their sum, respectively. The details are elaborated in the main text.

and SISA is mainly contributed by the increased static stability, with little contribution from the change in diabatic heating.

- 3) The change in diabatic heating contributes substantially to the projected changes in the NPSA, NASA, and SPSA. The enhanced diabatic heating over western North Pacific acts to weaken the NPSA, and its combined effect with increased static stability explains the strongest reduction in NPSA intensity among the five subtropical anticyclones. The reduced diabatic heating over the subtropical North Atlantic and South Pacific overwhelms the weakening effect of increased static stability, and finally enhances the NASA and SPSA. The pattern of change in diabatic heating is dominated by latent heating associated precipitation. The enhanced latent heating over western North Pacific is possibly a manifestation of the “richest get richer” mechanism with regard to the future rainfall change, while the relatively weak SST warming over subtropical North Atlantic and South Pacific are responsible for the reduced local latent heating and enhanced NASA and SPSA.

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REFERENCES

- Bellomo, K., and A. C. Clement, 2015: Evidence for weakening of the Walker circulation from cloud observations. *Geophys. Res. Lett.*, **42**, 7758–7766, doi:10.1002/2015GL065463.
- Bony, S., G. Bellon, D. Klocke, S. Sherwood, S. Fermepein, and S. Denvil, 2013: Robust direct effect of carbon dioxide on tropical circulation and regional precipitation. *Nat. Geosci.*, **6**, 447–451, doi:10.1038/ngeo1799.
- Cai, W., T. Cowan, and M. Thatcher, 2012: Rainfall reductions over Southern Hemisphere semi-arid regions: The role of

- subtropical dry zone expansion. *Sci. Rep.*, **2**, 702, doi:[10.1038/srep00702](https://doi.org/10.1038/srep00702).
- Ceppi, P., and D. L. Hartmann, 2016: Clouds and the atmospheric circulation response to warming. *J. Climate*, **29**, 783–799, doi:[10.1175/JCLI-D-15-0394.1](https://doi.org/10.1175/JCLI-D-15-0394.1).
- Chadwick, R., I. Boutle, and G. Martin, 2013: Spatial patterns of precipitation change in CMIP5: Why the rich do not get richer in the tropics. *J. Climate*, **26**, 3803–3822, doi:[10.1175/JCLI-D-12-00543.1](https://doi.org/10.1175/JCLI-D-12-00543.1).
- Chen, P., M. P. Hoerling, and R. M. Dole, 2001: The origin of the subtropical anticyclones. *J. Atmos. Sci.*, **58**, 1827–1835, doi:[10.1175/1520-0469\(2001\)058<1827:TOOTSA>2.0.CO;2](https://doi.org/10.1175/1520-0469(2001)058<1827:TOOTSA>2.0.CO;2).
- Dima, I. M., and J. M. Wallace, 2003: On the seasonality of the Hadley cell. *J. Atmos. Sci.*, **60**, 1522–1527, doi:[10.1175/1520-0469\(2003\)060<1522:OTSOTH>2.0.CO;2](https://doi.org/10.1175/1520-0469(2003)060<1522:OTSOTH>2.0.CO;2).
- DiNezio, P. N., G. A. Vecchi, and A. C. Clement, 2013: Detectability of changes in the Walker circulation in response to global warming. *J. Climate*, **26**, 4038–4048, doi:[10.1175/JCLI-D-12-00531.1](https://doi.org/10.1175/JCLI-D-12-00531.1).
- Frierson, D. M. W., 2006: Robust increases in midlatitude static stability in simulations of global warming. *Geophys. Res. Lett.*, **33**, L24816, doi:[10.1029/2006GL027504](https://doi.org/10.1029/2006GL027504).
- Garfinkel, C. I., D. W. Waugh, and L. M. Polvani, 2015: Recent Hadley cell expansion: The role of internal atmospheric variability in reconciling modeled and observed trends. *Geophys. Res. Lett.*, **42**, 10 824–10 831, doi:[10.1002/2015GL066942](https://doi.org/10.1002/2015GL066942).
- Gill, A. E., 1980: Some simple solutions for heat-induced tropical circulation. *Quart. J. Roy. Meteor. Soc.*, **106**, 447–462, doi:[10.1002/qj.49710644905](https://doi.org/10.1002/qj.49710644905).
- He, C., and T. Zhou, 2015: Responses of the western North Pacific subtropical high to global warming under RCP4.5 and RCP8.5 scenarios projected by 33 CMIP5 models: The dominance of tropical Indian Ocean–tropical western Pacific SST gradient. *J. Climate*, **28**, 365–380, doi:[10.1175/JCLI-D-13-00494.1](https://doi.org/10.1175/JCLI-D-13-00494.1).
- , —, A. Lin, B. Wu, D. Gu, C. Li, and B. Zheng, 2015: Enhanced or weakened western North Pacific subtropical high under global warming? *Sci. Rep.*, **5**, 16771, doi:[10.1038/srep16771](https://doi.org/10.1038/srep16771).
- Held, I. M., and B. J. Soden, 2006: Robust responses of the hydrological cycle to global warming. *J. Climate*, **19**, 5686–5699, doi:[10.1175/JCLI3990.1](https://doi.org/10.1175/JCLI3990.1).
- Hu, Y., and Q. Fu, 2007: Observed poleward expansion of the Hadley circulation since 1979. *Atmos. Chem. Phys.*, **7**, 5229–5236, doi:[10.5194/acp-7-5229-2007](https://doi.org/10.5194/acp-7-5229-2007).
- Kalnay, E., and Coauthors, 1996: The NCEP/NCAR 40-Year Reanalysis Project. *Bull. Amer. Meteor. Soc.*, **77**, 437–471, doi:[10.1175/1520-0477\(1996\)077<0437:TNYRP>2.0.CO;2](https://doi.org/10.1175/1520-0477(1996)077<0437:TNYRP>2.0.CO;2).
- Knutson, T. R., and S. Manabe, 1995: Time-mean response over the tropical Pacific to increased CO₂ in a coupled ocean–atmosphere model. *J. Climate*, **8**, 2181–2199, doi:[10.1175/1520-0442\(1995\)008<2181:TMROTT>2.0.CO;2](https://doi.org/10.1175/1520-0442(1995)008<2181:TMROTT>2.0.CO;2).
- Lau, W. K. M., and K.-M. Kim, 2015: Robust Hadley circulation changes and increasing global dryness due to CO₂ warming from CMIP5 model projections. *Proc. Natl. Acad. Sci. USA*, **112**, 3630–3635, doi:[10.1073/pnas.1418682112](https://doi.org/10.1073/pnas.1418682112).
- Lee, S.-K., D. B. Enfield, and C. Wang, 2011: Future impact of differential interbasin ocean warming on Atlantic hurricanes. *J. Climate*, **24**, 1264–1275, doi:[10.1175/2010JCLI3883.1](https://doi.org/10.1175/2010JCLI3883.1).
- Leloup, J., and A. Clement, 2009: Why is there a minimum in projected warming in the tropical North Atlantic Ocean? *Geophys. Res. Lett.*, **36**, L14802, doi:[10.1029/2009GL038609](https://doi.org/10.1029/2009GL038609).
- L’Heureux, M. L., S. Lee, and B. Lyon, 2013: Recent multidecadal strengthening of the Walker circulation across the tropical Pacific. *Nat. Climate Change*, **3**, 571–576, doi:[10.1038/nclimate1840](https://doi.org/10.1038/nclimate1840).
- Li, W., L. Li, M. Ting, and Y. Liu, 2012: Intensification of Northern Hemisphere subtropical highs in a warming climate. *Nat. Geosci.*, **5**, 830–834, doi:[10.1038/ngeo1590](https://doi.org/10.1038/ngeo1590).
- , and Coauthors, 2013: Intensification of the Southern Hemisphere summertime subtropical anticyclones in a warming climate. *Geophys. Res. Lett.*, **40**, 5959–5964, doi:[10.1002/2013GL058124](https://doi.org/10.1002/2013GL058124).
- Liu, Y., and G. Wu, 2000: Reviews on the study of the subtropical anticyclone and new insights on some fundamental problems (in Chinese). *Acta Meteor. Sin.*, **58**, 500–512.
- , —, and R. Ren, 2004: Relationship between the subtropical anticyclone and diabatic heating. *J. Climate*, **17**, 682–698, doi:[10.1175/1520-0442\(2004\)017<0682:RBTSAA>2.0.CO;2](https://doi.org/10.1175/1520-0442(2004)017<0682:RBTSAA>2.0.CO;2).
- Lu, J., G. A. Vecchi, and T. Reichler, 2007: Expansion of the Hadley cell under global warming. *Geophys. Res. Lett.*, **34**, L06805, doi:[10.1029/2006GL028443](https://doi.org/10.1029/2006GL028443).
- Ma, J., and S.-P. Xie, 2013: Regional patterns of sea surface temperature change: A source of uncertainty in future projections of precipitation and atmospheric circulation. *J. Climate*, **26**, 2482–2501, doi:[10.1175/JCLI-D-12-00283.1](https://doi.org/10.1175/JCLI-D-12-00283.1).
- , and J.-Y. Yu, 2014: Paradox in South Asian summer monsoon circulation change: Lower tropospheric strengthening and upper tropospheric weakening. *Geophys. Res. Lett.*, **41**, 2934–2940, doi:[10.1002/2014GL059891](https://doi.org/10.1002/2014GL059891).
- , S.-P. Xie, and Y. Kosaka, 2012: Mechanisms for tropical tropospheric circulation change in response to global warming. *J. Climate*, **25**, 2979–2994, doi:[10.1175/JCLI-D-11-00048.1](https://doi.org/10.1175/JCLI-D-11-00048.1).
- Ma, S., and T. Zhou, 2016: Robust strengthening and westward shift of the tropical Pacific Walker circulation during 1979–2012: A comparison of 7 sets of reanalysis data and 26 CMIP5 models. *J. Climate*, **29**, 3097–3118, doi:[10.1175/JCLI-D-15-0398.1](https://doi.org/10.1175/JCLI-D-15-0398.1).
- McGregor, S., A. Timmermann, M. F. Stuecker, M. H. England, M. Merrifield, F.-F. Jin, and Y. Chikamoto, 2014: Recent Walker circulation strengthening and Pacific cooling amplified by Atlantic warming. *Nat. Climate Change*, **4**, 888–892, doi:[10.1038/nclimate2330](https://doi.org/10.1038/nclimate2330).
- Miyasaka, T., and H. Nakamura, 2010: Structure and mechanisms of the Southern Hemisphere summertime subtropical anticyclones. *J. Climate*, **23**, 2115–2130, doi:[10.1175/2009JCLI3008.1](https://doi.org/10.1175/2009JCLI3008.1).
- Namias, J., 1972: Influence of Northern Hemisphere general circulation on drought in northeast Brazil. *Tellus*, **24**, 336–343, doi:[10.3402/tellusa.v24i4.10648](https://doi.org/10.3402/tellusa.v24i4.10648).
- Nguyen, H., A. Evans, C. Lucas, I. Smith, and B. Timbal, 2013: The Hadley circulation in reanalyses: Climatology, variability, and change. *J. Climate*, **26**, 3357–3376, doi:[10.1175/JCLI-D-12-00224.1](https://doi.org/10.1175/JCLI-D-12-00224.1).
- Power, S. B., F. Delage, R. Colman, and A. Moise, 2012: Consensus on twenty-first-century rainfall projections in climate models more widespread than previously thought. *J. Climate*, **25**, 3792–3809, doi:[10.1175/JCLI-D-11-00354.1](https://doi.org/10.1175/JCLI-D-11-00354.1).
- Qu, X., and G. Huang, 2016: The global warming-induced South Asian high change and its uncertainty. *J. Climate*, **29**, 2259–2273, doi:[10.1175/JCLI-D-15-0638.1](https://doi.org/10.1175/JCLI-D-15-0638.1).
- Rodwell, M. J., and B. J. Hoskins, 2001: Subtropical anticyclones and summer monsoons. *J. Climate*, **14**, 3192–3211, doi:[10.1175/1520-0442\(2001\)014<3192:SAASM>2.0.CO;2](https://doi.org/10.1175/1520-0442(2001)014<3192:SAASM>2.0.CO;2).
- Seager, R., R. Murtugudde, N. Naik, A. Clement, N. Gordon, and J. Miller, 2003: Air–sea interaction and the seasonal cycle of

- the subtropical anticyclones. *J. Climate*, **16**, 1948–1966, doi:[10.1175/1520-0442\(2003\)016<1948:AIATSC>2.0.CO;2](https://doi.org/10.1175/1520-0442(2003)016<1948:AIATSC>2.0.CO;2).
- , N. Naik, and G. A. Vecchi, 2010: Thermodynamic and dynamic mechanisms for large-scale changes in the hydrological cycle in response to global warming. *J. Climate*, **23**, 4651–4668, doi:[10.1175/2010JCLI3655.1](https://doi.org/10.1175/2010JCLI3655.1).
- Seo, K.-H., D. M. W. Frierson, and J.-H. Son, 2014: A mechanism for future changes in Hadley circulation strength in CMIP5 climate change simulations. *Geophys. Res. Lett.*, **41**, 5251–5258, doi:[10.1002/2014GL060868](https://doi.org/10.1002/2014GL060868).
- Shaw, T. A., and A. Voigt, 2015: Tug of war on summertime circulation between radiative forcing and sea surface warming. *Nat. Geosci.*, **8**, 560–566, doi:[10.1038/ngeo2449](https://doi.org/10.1038/ngeo2449).
- Su, H., J. H. Jiang, C. Zhai, T. J. Shen, J. D. Neelin, G. L. Stephens, and Y. L. Yung, 2014: Weakening and strengthening structures in the Hadley circulation change under global warming and implications for cloud response and climate sensitivity. *J. Geophys. Res. Atmos.*, **119**, 5787–5805, doi:[10.1002/2014JD021642](https://doi.org/10.1002/2014JD021642).
- Tao, L., Y. Hu, and J. Liu, 2016: Anthropogenic forcing on the Hadley circulation in CMIP5 simulations. *Climate Dyn.*, **46**, 3337–3350, doi:[10.1007/s00382-015-2772-1](https://doi.org/10.1007/s00382-015-2772-1).
- Taylor, K. E., R. J. Stouffer, and G. A. Meehl, 2012: An overview of CMIP5 and the experiment design. *Bull. Amer. Meteor. Soc.*, **93**, 485–498, doi:[10.1175/BAMS-D-11-00094.1](https://doi.org/10.1175/BAMS-D-11-00094.1).
- Ting, M., 1994: Maintenance of northern summer stationary waves in a GCM. *J. Atmos. Sci.*, **51**, 3286–3308, doi:[10.1175/1520-0469\(1994\)051<3286:MONSSW>2.0.CO;2](https://doi.org/10.1175/1520-0469(1994)051<3286:MONSSW>2.0.CO;2).
- Vecchi, G. A., and B. J. Soden, 2007: Global warming and the weakening of the tropical circulation. *J. Climate*, **20**, 4316–4340, doi:[10.1175/JCLI4258.1](https://doi.org/10.1175/JCLI4258.1).
- van Vuuren, D., and Coauthors, 2011: The representative concentration pathways: An overview. *Climatic Change*, **109**, 5–31, doi:[10.1007/s10584-011-0148-z](https://doi.org/10.1007/s10584-011-0148-z).
- Watanabe, M., and M. Kimoto, 2000: Atmosphere–ocean thermal coupling in the North Atlantic: A positive feedback. *Quart. J. Roy. Meteor. Soc.*, **126**, 3343–3369, doi:[10.1002/qj.49712657017](https://doi.org/10.1002/qj.49712657017).
- Xie, S. P., C. Deser, G. A. Vecchi, J. Ma, H. Y. Teng, and A. T. Wittenberg, 2010: Global warming pattern formation: Sea surface temperature and rainfall. *J. Climate*, **23**, 966–986, doi:[10.1175/2009JCLI3329.1](https://doi.org/10.1175/2009JCLI3329.1).
- Yanai, M., S. Esbensen, and J. Chu, 1973: Determination of bulk properties of tropical cloud clusters from large-scale heat and moisture budgets. *J. Atmos. Sci.*, **30**, 611–627, doi:[10.1175/1520-0469\(1973\)030<0611:DOBPOT>2.0.CO;2](https://doi.org/10.1175/1520-0469(1973)030<0611:DOBPOT>2.0.CO;2).
- Zhang, L., and T. Li, 2017: Relative roles of differential SST warming, uniform SST warming and land surface warming in determining the Walker circulation changes under global warming. *Climate Dyn.*, **48**, 987–997, doi:[10.1007/s00382-016-3123-6](https://doi.org/10.1007/s00382-016-3123-6).