Responses of the Hadley Circulation to Regional Sea Surface Temperature Changes

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ABSTRACT

Idealized experiments performed with the Community Atmospheric Model 5.3 indicate that the width and strength of the Hadley circulation (HC) are sensitive to the location of sea surface temperature (SST) increases. The HC edge shifts poleward in response to SST increases over the subtropical regions near and on the equatorward flank of the HC edge, and shifts equatorward in response to warming over the tropical area except for the western Pacific Ocean and Indian Ocean. The HC is strengthened in response to SST increases over the intertropical convergence zone (ITCZ) and is weakened in response to SST increases over the subsidence branch of the HC in the subtropics. Tropical SST increases off the ITCZ tend to weaken the HC in the corresponding hemisphere and strengthen the HC in the opposite hemisphere. These results could be used to explain the simulated HC changes induced by recent SST variations, and it is estimated that more than half of the SST-induced HC widening in 1980–2014 is caused by changes in the spatial pattern of SST.

1. Introduction

Atmospheric circulations are closely coupled with clouds, radiation, and precipitation, and play a key role in both global energy and hydrologic cycles (Lu et al. 2007; Ceppi and Hartmann 2015; Bindoff et al. 2013; Su et al. 2017; Grise and Polvani 2016). However, our understanding of atmospheric circulation change remain limited, and together with clouds and climate sensitivity it is listed as one of the World Climate Research Programme (WCRP) grand challenges (Bony et al. 2015).

One of the most prominent circulation changes under global warming is the expansion of the latitudinal extent

of the Hadley circulation (HC) (Bindoff et al. 2013). Multiple observational studies suggest that the HC has expanded over the last several decades (Hu and Fu 2007; Hu et al. 2011; Grise et al. 2019), and climate models simulate the poleward expansion of HC in both historical and future climate simulations (Lu et al. 2008; Hu et al. 2013; Tao et al. 2016). These changes are accompanied by poleward shifts of subtropical dry zones (Lu et al. 2007), subtropical jets, and storm tracks (Mbengue and Schneider 2013), and thus are important to both global and regional climate changes.

Observed multidecadal trends in HC expansion are much larger than those simulated by climate models forced by the prescribed changes in atmospheric composition (Johanson and Fu 2009; Hu et al. 2013; Allen and

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Kovilakam 2017; Mantsis et al. 2017; Staten et al. 2018; Grise et al. 2019). The magnitude of HC expansion in AMIP simulations is more consistent with observations than that in coupled CMIP5 historical simulations, implying that the observed HC expansion is partially attributable to the low-frequency evolution of the global sea surface temperature (SST) (Allen et al. 2014; Allen and Kovilakam 2017; Staten et al. 2018), in addition to anthropogenic atmospheric forcings from increasing greenhouse gases (Lu et al. 2007; Staten et al. 2012), stratospheric ozone depletion (Son et al. 2010), and aerosol concentration changes (Allen and Sherwood 2011; Rotstayn et al. 2014; Allen and Ajoku 2016).

While the HC edges shift poleward in response to the greenhouse gas-induced global warming, the magnitude of HC expansion is related to global surface temperature anomalies in a nonlinear fashion. In the experiments where abrupt CO₂ concentration is abruptly quadrupled, most of the poleward shift of the midlatitude jets and HC edge occurs within 5–10 years, during which only less than half of the expected equilibrium warming is realized, and the HC width does not expand significantly when the surface temperature increases steadily in the next 140 years (Ceppi et al. 2018).

The warm phase of El Niño–Southern Oscillation (ENSO) can also induce a warming in a globally averaged sense (Seager et al. 2003); however, the HC shrinks as a circulation response (e.g., Lu et al. 2008; Tandon et al. 2013), epitomizing the importance of the SST pattern in shaping the HC configuration. Therefore, it is imperative to understand how the HC width responds to the spatial pattern of the SST forcing.

Brayshaw et al. (2008) analyzed the storm-track response to idealized SST perturbations in an aquaplanet GCM, and their results indicate that the HC edge shifts equatorward in response to an increase of the SST gradient in the subtropics, and shifts poleward in response to an increase of the SST gradient in the midlatitudes. Chen et al. (2010) analyzed the sensitivity of the zonal mean atmospheric circulation to various SST anomaly patterns in an aquaplanet model, and found distinct HC width responses to meridionally narrow and broad tropical warmings. They also suggest that more work is needed to explore the sensitivity to zonally asymmetry in the SST forcing. Baker et al. (2017) analyzed the jet latitude sensitivity, which is closely related to the HC width, to diabatic heating under a series of Gaussian patch thermal forcing simulations, and found that the jet latitude response to forcing scales approximately linearly with the strength of the forcing and when forcings are applied in combination. Feng et al. (2019) analyzed the response of HC to SST variations in models from phase 5 of the Coupled Model Intercomparison Project (CMIP5) by decomposing them into equatorially symmetric and asymmetric components, and concluded that climate models are

capable of reproducing the observed climatological features of the equatorially symmetric components for both SST and the HC.

Similar to HC width, HC strength is also sensitive to the spatial structure of the SST forcing. Modeling studies indicate that the HC would be strengthened under El Niño conditions, but would be weakened in response to the more uniform SST warming induced by greenhouse gases increases (e.g., Lu et al. 2008). Reanalyses indicate that HC strength has intensified over the last three decades, but the positive trend may be partly due to systematic errors in these reanalyses (Mitas and Clement 2006; Hu et al. 2018). It is worth noting that these issues might also contribute to the difference of HC expansion rate between earlier reanalyses and AMIP simulations (Grise et al. 2019). Therefore, it is also important to understand the effect of SST anomaly patterns on HC strength.

In this study, we investigate systematically how the HC responds to SST forcing from different regions of the global ocean using idealized patch warming experiments. The sensitivity derived from the SST patch experiments (or SST Green's function experiments) can provide an important guide for interpreting the evolution of the HC width and strength in both historical period and their response under global warming.

2. HC width and strength changes in individual patch experiments

Following Zhou et al. (2017), patch experiments were carried out with the Community Earth System Model 1.2.1 with Community Atmospheric Model 5.3 (CESM1.2.1-CAM5.3) at 1.9° latitude × 2.5° longitude resolution (Neale et al. 2012). CAM5.3 is a version of CAM5, and has been widely used in climate research. The changes of HC width over the last three decades simulated by CAM5 are consistent with that simulated by an ensemble of CMIP5 simulations (Allen and Kovilakam 2017), indicating that CAM5 is a representative model in simulating the changes of HC width. The experiments include a control experiment, two sets of warm patch experiments, and two sets of cold patch experiments. The control experiments are 41 years long, with SST, sea ice, and climate forcings fixed at year 2000. Then the global ocean is divided into 80 overlapping rectangular patches that cover the entire ice-free ocean surface (Fig. 1). In each warm patch experiment, a positive SST anomaly is added to the ocean surface in a specific patch, with SST elsewhere being the same as the control experiment. In each cool patch experiment, a negative SST anomaly is added to the ocean surface in the corresponding patch. The SST anomaly inside

FIG. 1. Geographic illustration of the 80 patches used for the regional SST Green's function experiments. The dots denote geographic centers of corresponding patches, and the surrounding ellipses illustrate the size of the patches ($20^\circ \times 80^\circ$).

each patch follows the formulation of Barsugli and Sardeshmukh (2002) to avoid nonlinearity caused by unrealistic SST gradients:

$$\Delta \text{SST}_p(\text{lat}, \text{lon}) = A \cos^2 \left(\frac{\pi}{2} \frac{\text{lat} - \text{lat}_p}{\text{lat}_w} \right) \times \cos^2 \left(\frac{\pi}{2} \frac{\text{lon} - \text{lon}_p}{\text{lon}_w} \right), \quad (1)$$

where $|\text{lat} - \text{lat}_p| < \text{lat}_w$, $|\text{lon} - \text{lon}_p| < \text{lon}_w$. The terms lat_p and lon_p are the latitude and longitude of the center point for a specific patch, respectively; lat_w and lon_w are the meridional and zonal half-width of the patch, respectively, with their values set to $\text{lat}_w = 10^\circ$ and $\text{lon}_w = 40^\circ$ in this study; and A is the amplitude of the SST anomaly. To test

for nonlinearity, two suites of experiments with $A = 2$ and 4 K are conducted. The area-averaged SST change inside a land-free patch is 0.5 and 1 K when $A = 2$ and 4 K, respectively; we therefore refer to them as 0.5 and 1 K patch experiments. Each 1 K patch experiment ($A = 4$ K) is run for 41 years, and each 0.5 K patch experiment ($A = 2$ K) is run for 11 years. More details about the patch experiments are summarized in Table 1.

The responses of HC width and strength to patch warmings are calculated as the difference between climatological mean value in conjugate ± 1 K patch experiments. The latitude of HC edge in each hemisphere is calculated as the latitude where the meridional mass streamfunction (MMS) at 500 hPa crosses zero in the subtropics, and the total width of HC can be defined as the latitudinal range between the NH HC edge and SH HC edge. The annual HC width is calculated using annual mean MMS, and seasonal HC width is calculated using the corresponding seasonal MMS. The HC strength is defined as the absolute value of the extremum of MMS between the HC edge and the intertropical convergence zone (ITCZ; defined as the location where the 500 hPa MMS is zero near the equator in this study) in each hemisphere.

Figures 2a and 2b show the degrees of shift of HC edge in the Northern Hemisphere (NH) and Southern Hemisphere (SH), respectively, in response to each of the local SST patches, with the amount of the shift indicated at the correspond patch location. In general, a subtropical SST warming near and on the equatorward flank of HC edge causes the HC edge to move poleward, and a deep tropical SST warming causes the HC edge to move equatorward [consistent with Allen et al. (2012) and Tandon et al. (2013)]. An exception is the response to tropical Indian Ocean warming, which leads to a

TABLE 1. List of idealized experiments used in this study.

Experiment	No. of runs	No. of years used	Description
Control	1	40	SST, sea ice, and climate forcings fixed at year 2000
Patch experiments			
+1 K patch	80	40	As in control, except SST in a specific patch is revised using Eq. (1); A is set to $+4$ or -4 K for 1 K patch experiments, and $+2$ or -2 K for 0.5 K patch experiments
-1 K patch	80	40	
+0.5 K patch	80	10	
-0.5 K patch	80	10	
Historical simulations			
AMIPFF simulations	6	35	SST and sea ice are same as observations, climate forcings are fixed. Each run starts with slightly different initial conditions.
AMIP simulations	5	26	SST, sea ice, and all climate forcings are the same as observations.

FIG. 2. Sensitivity map of HC edge to the locations of the SST perturbations. (a) Differences between the latitude of NH HC edge in conjugate ± 1 K patch experiments. Red (blue) dots denote a northward (southward) shift of the NH HC edge in response to warming in the designated patch. Patches with statistically significant response (at 95% level) are marked with plus symbols. (b) As in (a), but for SH HC edge. The units in (a) and (b) are degrees latitude. (c),(d) As in (a) and (b), but for the zonally averaged sensitivity and for both annual mean and seasonal mean HC response. The estimation of the zonal average of the sensitivity is based on the sensitivity at each grid box, which itself is a weighted average among the patches overlapping at that grid point. Colored lines denote the sensitivities in different seasons (see the legend). The vertical dashed lines mark the climatological mean locations of the HC edge in the corresponding seasons.

significant expansion of the NH HC. The associated zonal asymmetry in the HC width sensitivity has important bearing on the interpretation of the response to ENSO-like forcing: an increase (a decrease) in the equatorial SST gradient in Pacific during La Niña (El Niño) tends to expand (shrink) the HC, consistent with observations (Lu et al. 2008). Thus, the narrowing of the HC under El Niño is not entirely the result of the equatorward confinement of the SST feature; the zonally asymmetric aspect of the SST anomalies is also important.

Figures 2c and 2d show the sensitivity of HC edge to regional SST increases as a function of latitude, where the values at each latitude are calculated as the mean value of the sensitivity matrix (to be explained in section 3). For both hemispheres, the latitude of annual HC edge is most sensitive to sea surface warmings around 30° latitude in the corresponding hemisphere.

Interestingly, the latitude of maximum sensitivity varies with season, progressing poleward as the season warms. The poleward expansion of the NH HC is most sensitive to warming around 20°N in March, April, and May (MAM) and December, January, and February (DJF), but is most sensitive to warmings around 30°N in September, October, and November (SON) and June, July, and August (JJA). Likewise, the poleward expansion of the SH HC is most sensitive to warmings around 30°S in MAM and DJF, and is most sensitive to warmings around 20°S in SON and JJA. This is consistent with the fact that HC edge is farther poleward in autumn and summer (SON and JJA for NH; MAM and DJF for SH) than in spring and winter (MAM and DJF for NH; SON and JJA for SH) in both hemispheres, so the latitude of maximum sensitivity is also higher in autumn and summer than in spring and winter. The sensitivity map of the westerly jet shift bears some

FIG. 3. As in Fig. 2, but for the latitude of the eddy-driven jet.

resemblance to that of the HC (cf. Figs. 2a,b and 3a,b). In particular, the subtropical SST warming appears to be most effective in driving the poleward shift of the jet and HC expansion in the same hemisphere. Some distinctions do exist. The sensitivity map of the westerly jet is somewhat broader meridionally compared to its HC sensitivity counterpart. Inspecting the seasonal sensitivity maps, we notice the sensitivity of the westerly jet to SST perturbations near 40° latitude appears to arise mainly from the winter season, suggestive of different governing mechanisms for HC width and jet location in winter, when the subtropical jet is strong and separated in latitude from the eddy-driven jet. For the NH circulation indices, while the HC edge is most sensitive to equatorial SST changes in summer, the jet position is more sensitive to it in winter and spring (comparing Figs. 2c and 3c). For the sensitivity of the SH circulation indices, the most significant difference is the opposite sensitivities between the HC edge and the jet position over the tropical northwest Pacific (Figs. 2b and 3b). The positive sensitivity of the former seems to arise from the JJA (austral winter) HC sensitivity to the western Pacific monsoon heating in the NH (red line in Fig. 2d). Although the exact mechanism for this

teleconnection remains elusive, it is somewhat in keeping with the classic theory of Lindzen and Hou (1988): as the off-equatorial monsoon heating in the summer hemisphere is enhanced, a greater area of descent must occur to balance the heating through the radiative cooling induced by the descent, as consequentially the winter cell expands in the opposite hemisphere.

It is difficult to explain all the nuanced features revealed by the HC and westerly jet sensitivity maps in one study. Notwithstanding, the greatest sensitivities to the subtropical SST forcing of the both circulation indices are generally consistent with the insights gained from studies utilizing more idealized contexts (e.g., Chen et al. 2010; Tandon et al. 2013; Sun et al. 2013). To the extent that the terminus/edge of the HC is set by location of the sign switch of the eddy momentum forcing (Held et al. 2000; Walker and Schneider 2006; Körtj and Schneider 2008; Lu et al. 2008), the shift of the HC edge is strongly influenced by the shift of the baroclinicity. For a case of the subtropical SST warming in DJF (as, for example, in Fig. S1a in the online supplemental material), the lower-tropospheric baroclinicity is enhanced (reduced) on the poleward (equatorward)

FIG. 4. Response of EP flux and its divergence to surface warming in the (a),(c) eastern Pacific Ocean and (b),(d) western Indian Ocean. Arrows denote the response of the EP flux vector, and shading denotes the EP flux divergence. The contours denote the responses of the zonal mean zonal wind. Contribution from (a),(b) the stationary waves and (c),(d) the transient waves.

side of the mean baroclinicity maximum (Fig. S1e; the 850 hPa temperature gradient is used to indicate the baroclinicity). The responses of the eddy-driven jet and HC boundary then simply follow the shift of the mid-latitude baroclinicity. This seems to be also true for the equatorial SST warming, as exemplified by Figs. S2 and S3.

Another feature in the sensitivity map that deserves an explanation is the notable zonal asymmetry in the tropical oceans, which is most significant in DJF: SST forcing in the tropical eastern Pacific acts to shift the westerly jet equatorward and to shrink the HC in the NH, whereas the SST forcing in the tropical Indian Ocean has the opposite effects (see Figs. 2a and 3a). Similar contrasting effects between the Indian Ocean and eastern Pacific forcing were noticed previously in the response of the winter Aleutian low (Deser and Phillips 2006) and northern annular mode (Fletcher and Kushner 2011). Here, we perform Eliassen–Palm flux (EP flux) analysis for the cases of the eastern Pacific forcing and western Indian Ocean forcing, respectively, and the results are displayed in Fig. 4. In particular, we

decompose the EP flux response into the part due to the stationary wave (Figs. 4a,b) and that due to the transient waves (Figs. 4c,d). First, the stationary wave response is much greater in the NH than in the SH, reflecting the greater zonal asymmetry in the background climatological circulation in the former. More importantly, the midlatitude EP flux convergence (divergence) in the Indian Ocean forcing (east Pacific forcing) case is predominantly caused by the stationary wave, providing the maintaining mechanism for the anomalous zonal mean easterly (westerly) response there. On the other hand, the transient waves appear to counter the stationary wave in the midlatitudes. Further inspection of the spatial patterns of the excited stationary waves (Figs. S2h and S3h) indicates that the wave sources due to Indian Ocean and eastern Pacific SST warming excite teleconnection waves of opposite phase, with the former converging momentum at the poleward side of the mean jet and the latter diverging momentum there. It is worth noting that our results in the case of eastern Pacific SST warming offer a complementary perspective to the wave refractivity argument (e.g., Seager et al. 2003) and the

FIG. 5. As in Fig. 2, but for HC intensity. Vertical dashed lines denote the climatological locations of the ITCZ.

critical-latitude argument (Chen et al. 2008) used to explain the equatorward shift in response to the El Niño SST forcing.

Figure 5 shows the response of HC strength to SST forcing from different locations. Compared to the sensitivity of the HC width, there is considerable zonal symmetry in the HC strength sensitivity in the tropics and subtropics. Surface warming in southern tropics and cooling in northern tropics act to intensify the NH HC (Fig. 5a), while an opposite SST forcing pattern tends to intensify the SH HC (Fig. 5b). A salient anti-correlation exists between the sensitivity map of the NH HC strength and that of the SH HC strength, indicating that the strengthening of HC in one hemisphere occurs at the expense of the strength in the other. The only area where SST warming can drive an intensification of the HC in both hemispheres is that over the tropical Atlantic. Examined season by season, the large zonal mean sensitivity is generally characterized by a dipole for both NH HC (Fig. 5c) and SH HC (Fig. 5d), whereas the nodal point for the NH HC sensitivity tends to shift northward as seasons goes from cold to warm. More specifically, HC can be most effectively intensified if surface warming occurs at its rising branch or at the seasonal ITCZ and/or if surface

cooling occurs at its descending branch. The sensitivity is generally greater for the winter HC, especially the JJA cell in the SH. This seems to be readily explained by the regime behavior of HC manifested in response to moving the tropical heating progressively away from the equator (Lindzen and Hou 1988): the farther away from the ITCZ, the greater the HC strength sensitivity to the ITCZ heating. This seems to be borne out well in the large sensitivity of the HC during winter season, when the zonal mean ITCZ in the summer hemisphere is the most distant from the equator. The dipolar feature across the equator in the sensitivity of the annual mean HC intensity in Figs. 5a and 5b is the result of the annual integration of the dipolar sensitivity for each seasonal HC with the largest contribution from the winter ones (see the blue line in Fig. 5c and the red line in Fig. 5d).

The sensitivity to the equatorial eastern Pacific SST is intriguing: a warming there works to intensify the SH HC but weakens the NH HC (Figs. 5a,b). This is conceivably related to the fact that ITCZ in the eastern Pacific is northward displaced with respect to the equator throughout the year and a sea surface warming on the equator tends to shift the ITCZ heating toward the equator. However, the exact mechanism remains

FIG. 6. HC changes in conjugate ± 1 K experiments ($A = \pm 4$ K) and ± 0.5 K experiments ($A = \pm 2$ K). Each dot denotes the difference of (a),(b) HC width and (c),(d) HC strength in conjugate experiments for a specific patch. The error bars denote the 95% uncertainty intervals, and the black line denotes the $y = 2x$ line, where all dots would lie if the HC responses were perfectly linear.

unclear and cannot be fully explained by the nonlinear regime argument above.

3. Reconstruction of the historical-SST-induced HC changes with the Green's function approach

The HC sensitivities inferred from the patch experiments might be used to explain the HC changes in response to various SST changes if the HC responds approximately linearly with regional SST changes. We check the linearity by comparing the responses of HC width and strength in conjugate ± 0.5 and ± 1 K patch experiments. According to Fig. 6, the difference of HC width and strength in conjugate ± 1 K patch experiments is generally twice that in ± 0.5 K experiments, indicating that HC width and strength responses are approximately linear to regional temperature changes.

The sensitivity of HC width and strength to SST perturbations in a specific grid box (denoted by index i) can be estimated as follows (Zhou et al. 2017):

$$\frac{\partial R}{\partial \text{SST}_i} = \frac{\sum_p \Delta \text{SST}_p \left(\frac{\partial R}{\partial \text{SST}_i} \right)_p}{\sum_p \Delta \text{SST}_p} = \frac{\sum_p \Delta \text{SST}_p \frac{\partial R}{\partial \text{SST}_p} \frac{S_i}{S_p}}{\sum_p \Delta \text{SST}_p}, \quad (2)$$

where R is the width or strength of HC, S_i and S_p are the ocean surface area of the corresponding grid point and the patch, ΔSST_p is the SST anomaly of the grid in the p th patch [Eq. (1), the value of ΔSST_p is zero when a grid point is not included in a patch], and SST_i denotes the SST in the i th grid box. Also, $(\partial R / \partial \text{SST}_i)_p$ is the average response of R in response to 1 K warming in a specific grid box inside the patch, and $\partial R / \partial \text{SST}_p$ is the difference of HC width or strength divided by the patch-averaged SST difference in conjugate ± 1 K patch warming experiments. In Eq. (2), the sensitivity for grid boxes covered by a single patch equals to the average change of R in response to 1 K of warming in a specific grid box inside the corresponding patch, and the sensitivity for grid boxes covered by multiple patches (due to their

FIG. 7. The sensitivity of (a) latitude of NH HC edge, (b) latitude of SH HC edge, (c) NH HC strength, and (d) SH HC strength to surface warming in each grid box, calculated using Eq. (2). The units for (a) and (b) are degrees latitude K^{-1} and for (c) and (d) are $kg s^{-1} K^{-1}$.

overlapping) equals to the weighted mean value $(\partial R / \partial SST_i)_p$.

The sensitivity of annual HC width and strength to SST perturbations in specific grid boxes are shown in Fig. 7, and its seasonal counterparts are shown in Figs. S4–S7. The maps of the HC sensitivities put the response of the HC width in a global perspective. It is evident that the subtropical SST warming is a robust forcing pushing the edge of the HC in the local hemisphere poleward. Also clear is the break of the zonal symmetry in the tropical oceans by the opposite sensitivity over the Indian Ocean.

With the sensitivities of the HC width and strength, we can then predict the HC width and strength response to an arbitrary SST change using the Green's function approach:

$$\Delta R = \sum_i \frac{\partial R}{\partial SST_i} \Delta SST_i + \varepsilon, \quad (3)$$

where ε denotes an error due to the nonlinear terms. If ε is small compared to the linear terms, then Eq. (3) provides an interpretation and prediction framework through which the responses of HC width and strength

to a specific SST warming pattern could be constructed as the superposition of the HC responses to SST changes in each grid box.

We examined the performance of the Green's function in explaining the HC variations driven by historical SST, which is the primary driver of the interannual variability of HC width and strength (Fig. S8). In Fig. 8, the black lines denote the time series of annual HC width and strength in AMIPFF simulations (see Table 1), and the red lines denote the HC responses calculated with the Green's function approach. The Green's function-reconstructed HC width and strength are positively correlated with the actual AMIPFF simulations in both hemispheres, but the correlation coefficient is generally lower in the Northern Hemisphere. Several reasons are behind the lack of skill in capturing the variability of the NH HC. First, the NH circulation is more zonally asymmetric and inherently has more degrees of freedom for variations than the SH. Therefore, it is less constrained by the SST forcing than its southern counterpart, as reflected by the low correlation of HC width/strength among the different individual simulations (<0.4). In addition, the effect of nonlinear relationships between the HC and

FIG. 8. Responses of annual HC (a),(b) width and (c),(d) strength to observed SST changes during July 1980–June 2014 in CAM5.3. The poleward shift of HC edge at each time step is calculated as the difference between the absolute value of HC edge latitude at the corresponding time step and the climatological value. The annual mean value of each time step is averaged from July of the previous year to June of the corresponding year, so that each major El Niño event would be included in a specific annual cycle. Red lines are responses reconstructed with the Green's function approach, gray lines are for the six individual members of AMIPFF simulations (observed SST and sea ice cover, fixed climate forcings at year 2000), and the black lines are for the ensemble mean values of these simulations. The light blue shading denotes years with strong El Niño events. The correlation coefficient between the thick black and red lines is shown in each panel.

the SST forcing, which are likely greater in the NH than SH, is not accounted for in the Green's function approach. Notwithstanding, the performance of the Green's function approach is reasonably skillful in explaining the variance of annual NH/SH HC width and SH HC strength (Figs. 8a,b,d). The Green's function approach is also skillful in explaining the variations of DJF HC (Fig. 9), the season during which the El Niño signal is the strongest. The signal-to-noise ratio in other seasons is smaller than in DJF, so the corresponding performance of the Green's function approach is also relatively poorer compared to DJF (Figs. S9–S11).

Both the AMIPFF simulations and Green's function reconstructions suggest that the SST change during 1980–2014 would result in a trend of HC expansion in both hemispheres, with comparable magnitudes (Fig. 10). We decompose the change of SST into two components, a uniform SST change component where the SST anomaly

in each grid box is the same as global mean SST anomaly and a pattern component with zero global mean. The Green's function approach indicates that the SST pattern component contributes about 60% to the SST-induced HC expansion over 1980–2014 in both hemispheres (Fig. 10). These results are consistent with that of Allen and Kovilakam (2017), which suggests that decadal variability of the ocean might play an important role in recent HC expansion. On the other hand, changes in the SST pattern tend to reduce the total intensity of HC strength, consistent with the discussion of Hu et al. (2018).

4. Conclusions and discussion

The sensitivities of the Hadley cell (HC) edge and intensity, as well as the location of the midlatitude eddy-driven westerly jet, are investigated in a suite of SST Green's function perturbation experiments with NCAR's

FIG. 9. As in Fig. 8, but for HC changes in DJF.

CAM5.3 AGCM. The results show that the HC edge expands poleward in response to SST increases over subtropics, and moves equatorward in response to SST increases over tropical regions except for the western Pacific and Indian Oceans. The HC width is much less sensitive to SST increases over high latitudes in comparison. The HC is strengthened in response to SST increases over the ascending branch of the HC near the ITCZ, and is weakened in response to SST increases over the HC subsidence branch in the subtropics. In addition, the sensitivity to the ITCZ heating appears to be the greatest for the winter circulation regime.

These results are consistent with the change of HC width in response to various SST pattern changes. During strong El Niño events, there is an enhanced warming over tropical eastern Pacific Ocean and a cooling over subtropical Pacific Ocean, and both our AMIPFF simulations and Green's function approach suggest that the HC is generally narrower and stronger during DJF (the season with strongest El Niño signals) of major El Niño years (Fig. 9), consistent with results from previous studies (Lu et al. 2008). These results could also explain why the magnitude of HC expansion in AMIP simulations is larger than that in CMIP5 historical simulations.

Compared with CMIP5 historical simulations, observed warming (and thus warming in AMIP simulations) since 1980 is greater around 30°S and 30°N but weaker over the tropical eastern Pacific Ocean (Zhou et al. 2016; Allen and Kovilakam 2017). The enhanced subtropical warming in both hemispheres leads to a widening HC. Note that climate forcings such as the increase of greenhouse gases and black carbon aerosols might also contribute to the observed HC expansion.

What have been investigated so far are the sensitivities of HC width and strength to SST forcing only in a single AGCM; we cannot rule out the possibility that some of the conclusions may be model dependent. Further investigations with different AGCMs might be necessary to build confidence in the conclusions reached here, so as to translate the understanding of the behavior of HC in the model to the real world. Nevertheless, most of the results here are consistent with those reported in the literature and predicted from basic HC dynamics, and thus should hold at least qualitatively in nature. Some of the sensitivities revealed herein are not well understood, such as the distinct sensitivities between HC and the eddy-driven jet and their seasonality. Further in-depth dynamical analyses are warranted.

FIG. 10. (a),(b) Poleward trend of HC edges during 1980–2014. The four bars from left to right in each subplot denote values in AMIPFF simulations, values reconstructed by the Green's function approach, changes in response to the pattern component of SST trend, and changes in response to the uniform component of SST trend calculated with the Green's function approach, respectively. The error bars denote the 95% uncertainty interval of the trends. (c),(d) The pattern component and uniform component, respectively, of SST trend. The global mean value (averaged over an ocean area of 70°S–70°N) of the pattern component of SST trend is zero, and the actual SST trend equals to the sum of pattern component and the uniform component.

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