Response of the zonal mean atmospheric circulation to El Nino versus global warming.

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Abstract

The change in the zonal mean atmospheric circulation under global warming is studied in comparison with the response to El Nino forcing, by examining the model simulations conducted for the Intergovernmental Panel on Climate Change (IPCC) fourth assessment report (AR4). In contrast to the strengthening and contraction of the Hadley cell and the equatorward shift of the tropospheric zonal jets in response to El Nino, the Hadley cell weakens and expands poleward, and the jets move poleward under a warmed climate, despite the “El Nino-like” enhanced warming over equatorial central and eastern Pacific.

Two feasible mechanisms are proposed for the mean circulation response to global warming: (1) The increase in static stability of the subtropical and mid-latitude troposphere, a result of the quasi-moist adiabatic adjustment to the surface warming, tends to stabilize the eddy growth on the equatorward side of the storm track and push the eddy activity and the associated eddy-driven wind and subsidence poleward, leading to the poleward expansion of the Hadley cell and shift of jet; (2) the strengthening of the mid-latitude wind at the upper-troposphere and lower-stratosphere, arguably a consequence of the rise in the height of the tropopause and the associated increase in the meridional temperature gradient, can increase the phase speed of the eddies emanating from the mid-latitudes, and thus the critical latitudes (where the eddy phase speed matches the background zonal wind, and where the eddies break and extract angular momentum from the thermally driven wind) displace poleward and carry the eddy-driven circulation with it. Both mechanisms are somewhat, if not completely, distinct from those in response to the El Nino condition.

Since both mechanisms do not essentially depend on the details of the SST warming in the equatorial oceans and the variety of model physical parameterizations, these extratropical responses to global warming may be of increased credence due to the fact that they are dynamically determined.
1. Introduction

Given that El Nino and Southern Oscillation (ENSO) is the most prominent mode of interannual variability in the Earth climate system and has profound impacts on the global climate, there is great interest as to how the tropical Pacific sea surface temperature (SST) would respond to global warming. The Fourth Assessment Report (AR4) of the Intergovernmental Panel of Climate Change (IPCC) summarizes the changes in tropical Pacific SST in response to global warming as the following: “models have projected that the background tropical Pacific SST change from global warming (upon which individual ENSO events occur) will be an El Niño-like pattern. ……the El Niño-like change may be attributable to the general reduction of tropical circulations due to the increased static stability in the tropics in a warmer climate.” Meanwhile, there is less agreement in the IPCC models in the feature of the zonal SST gradient in the so-called El Nino-like response (e.g., Collins and CMIP modeling Groups (2005)) than in the feature of the longitudinal SST gradient between the equator and the subtropics (Liu et al., 2006). The latter, to zeroth order, can be explained by the fact that the upward latent heat flux increases more efficiently over the off-equatorial region (owing to the stronger easterly wind there) than at the equator upon a presumed same change in the sea-air humidity difference.

With no intension to join the debate of El Nino-like or La Nina-like response to global warming, we want to point out that, despite the El Nino-like features in the Pacific SST response in most of the models, the extratropical atmospheric responses to global warming may bear no resemblance at all to the El Nino teleconnection pattern; instead, they manifest themselves in somewhat opposite fashion.
Observational and diagnostic analyses (Seager et al., 2003) have shown that during an El Nino event, the tropical atmosphere warms at all longitudes, and the subtropical jets in both hemisphere strengthen on their equatorial flanks and shift towards the equator. Poleward of the tropical warming there are latitudinal belts of marked cooling, extending from the surface to the tropopause in both hemispheres, at all longitudes. The Hadley circulation intensifies and contracts equatorward. Similar responses occur in model simulations as well. Model experiments with a diabatic heating specified in the deep tropics of an idealized dry atmospheric model (Chang, 1995; Robinson 2002 (GRL); Son and Lee, 200?), or a SST warming confined within 15° to the equator over the Indo-Pacific oceans in the NCEP GFS model (Xiaowei Quan, personal communication), or a narrow equatorial SST warming in the GFDL AM2 aquaplanet model (Frierson and Lu, 2007), all produce consistent intensification and contraction of the Hadley cell and the associated changes in jets and atmospheric temperatures, as observed during El Nino conditions.

On the other hand, the atmospheric response to global warming is characterized by a poleward expansion of the Hadley circulation (Lu et al., 2007), a poleward shift in eddy-driven jets (Lorenz and DeWeaver, 2007; Kushner et al., 2001) and mid-latitude storm tracks (Yin, 2005), and a weakening of the tropical atmospheric circulation (Vecchi and Soden, 2007; Lu et al., 2007). Apparently, these features of change cannot be accounted for by the El Nino-like SST response or the enhanced equatorial SST warming. Here, we try to rationalize these changes in the atmospheric circulation and discuss the possible mechanisms under global warming in contrast to El Nino.
First, we describe the data, analysis metrics and techniques used in this study in Section 2. Sections 3 and 4 discuss the characteristics of the SST and atmospheric circulation changes under global warming in comparison with the El Nino condition. Section 5 discusses the differences in energetic and hydrological cycles between global warming and the El Nino. The mechanisms for the variation of the Hadley cell extent and the displacement of the eddy-driven jet are the subject of focus of this study, and will be elaborated upon in Section 6. Lastly, we conclude with summary and discussion in Section 7.

2. Data and analysis metrics

Most of the analyses in this study are based on simulations with the GFDL coupled climate model, CM2.1. The detailed formulation of CM2.1 has been documented by Delworth et al. (2006). The atmospheric component of CM2.1 uses a finite volume advection scheme and has approximately the same resolution (2 latitude × 2.5 longitude; 24 levels in the vertical) and atmospheric and land physics as an earlier version of the GFDL atmospheric model (AM2.0) and land model (LM2.0) (GFDL_GAMDT, 2004), except some necessary tuning of the cloud scheme and the frozen soil. The ocean component of the CM2.1 is referred to as OM3.1, which is based on the Modular Ocean Model code (MOM4; Griffies et al., 2003). OM3.1 has 50 vertical levels and a 1°×1° horizontal resolution poleward of 30°, with the meridional grids becoming progressively finer as one moves toward the equator. See Gnanadesiken et al. (2006) for the detailed configuration of OM3.1. No flux adjustment is applied during the coupling. The climate
sensitivity for a doubling of the CO₂ concentration is estimated to be 3.4 K by coupling the atmospheric component to a slab ocean model (Stouffer et al., 2006).

CM2.1 can reasonably simulate the periodicity and the spatial pattern of ENSO variability, except that the amplitude of the SST variations over the Nino regions are about twice as large as observed. Like other models in the community, CM2.1 also suffers some common biases in the tropical Pacific climate, such as a cold SST bias along the equator, and a westward extension of the trade winds relative to observations (Wittenberg et al., 2006). Nevertheless, the zonal mean atmospheric signature of ENSO is very comparable to observations.

The simulation of the A2 scenario, a high-emission scenario (the CO₂ concentration reaches 800 ppm at the end of the 21st century), will be the focus of the analysis in this study. To facilitate a comparison between ENSO and global warming, first, the simulated time series from 2001-2100 are detrended and sorted into two groups, one for years when the cold tongue index (SST averaged between 6°S-6°N, 180°-90°W, CTI henceforth) is larger than 1K and the other less than -1K. A composite is constructed by normalizing the data of each group member according to the corresponding CTI and then taking the difference between the two groups. As such, the composite anomalies correspond to a 2K departure of the CTI from its basic state. The trend is simply calculated as the difference between the beginning (2001-2020) and the end (2081-2100) of the 21st century; the difference is then rescaled to correspond to a 2K increase in the CTI.

To examine model dependence and sensitivity to model configurations, gridded global monthly data for the various models available from the WCRP CMIP3 multi-
model dataset (http://www-pcmdi.llnl.gov) are also employed. Occasionally, to increase
the sample size for the scaling/sensitivity analysis, scenarios with less aggressive GHG
forcing (A1B, with CO₂ stabilized at 720ppm, and B1, with CO₂ stabilized at 550ppm) are
also used.

To describe and understand the circulation changes, the following metrics are
defined and calculated:

a. *Meridional mass streamfunction* $\psi$

We compute the zonal mean mass streamfunction by vertically integrating the
zonal mean density-weighted meridional wind from the top model level downward. The
poleward edge of the Hadley cell can then be determined by the first turning latitude as
one moves poleward from the extrema of the streamfunction in each hemisphere.

b. *Surface westerly latitude (SWL)* $Y_{\text{max}}$

To relate the expansion of the Hadley cell to the poleward shift of the eddy driven
jet, we also compute the surface westerly latitude (SWL, $Y_{\text{max}}$), which is defined (using
linear interpolation) as the latitude where the meridional derivative of the zonal mean
surface wind with respect to latitude ($dU_s/dy$) becomes zero near the maximum of the
surface westerly winds.

c. *Tropopause height*

Following the World Meteorological Organization (1957) definition, the
tropopause height estimated from temperature data as the lowest pressure level at which
the lapse rate decreases to 2K/km, using the algorithm of Reichler at al. (2003).

d. *Eddy momentum flux spectrum*
The latitude-phase speed spectrum of eddy momentum flux is used to characterize the meridional propagation of baroclinic eddies (Randel and Held 1991, Kim and Lee 2004, Son and Lee 2006, Chen et al. 2007). The eddy momentum fluxes are first decomposed as a function of zonal wavenumber and frequency, using mixed space-time cross-spectral analysis (Hayashi, 1971). Following Randel and Held (1991), the cospectrum is transformed as a function of zonal wavenumber and angular phase speed. Finally, the momentum flux spectrum at each latitude is summed over wavenumber, resulting in a plot as a function of latitude and angular phase speed. The detailed method is summarized in Appendix A.

Daily time series of length 120 days in December-March are used to calculate the space-time spectrum for each year. The resolution in the phase speed space is limited by daily sampling, as indicated by the blank area near zero phase speed in Figures 13 and 14.

3. Characteristics of SST warming

Figure 1 shows the SST difference between 2001-2020 and 2081-2100 in A2 scenario simulations by 16 IPCC models. The multi-model mean exhibits a clear El Nino-like response in the tropical Pacific, with enhanced warming in the equatorial central and eastern Pacific. 14 out of the 16 models show intensified warming in the equatorial Pacific, and an El Nino-like SST pattern appears in all the models (MPI_ECHAM5, GFDL_CM2.0, GFDL_CM2.1, MIROC3.2 and HadCM3) that have been identified by previous studies (van Oldenborgh et al., 2005, Ocean Science; Yamaguchi and Noda, 2006, JMSJ) to be realistic in simulating the temporal and spatial characteristics and feedback mechanisms of the ENSO variability. The mechanisms for the El Nino-like SST
response to global warming have been widely studied. The most probable one may be rooted in the weakening of the tropical atmospheric circulation in a warmed climate (Knutson and Manabe, 1995; Vecchi et al., 2006, 2007 Science; Zhang and Song, 2006; Held and Soden, 2006). The slowdown of the zonal overturning of air across the Pacific (i.e., the Walker circulation) leads to a decrease of the surface easterlies which relaxes the zonal thermocline slope towards a flatter condition and initiates a chain of air-sea feedbacks towards a more El Nino-like condition. The so-called “ocean-thermostat” mechanism (Cane, 1997; Seager and Murtugudde, 1997; Clement et al., 1996), favors a warming at warm pool over the strong upwelling eastern Pacific, tends to counteract against the El Nino-like feedback (Bjerknes feedback). The net outcome from these two competing mechanisms turns out to be something in-between with somewhat dominance by the latter: warmer equatorial SST relative to the off-equatorial region, intensified convection over the central equatorial Pacific (more westward located than that associated with canonical El Nino) and an associated anomalous zonal overturning circulation that projects only weakly, if at all, on the canonical Southern Oscillation (van Oldenborgh et al., 2006 Ocean Science).
**Figure 1** Multi-model ensemble mean least-squared linear trend in SST (°K) during 2001-2100 based on 19 simulations of the A2 scenario.

As mentioned above, the enhanced zonally symmetric equatorial warming in the Pacific is much more robust than the zonally asymmetric warming. This is because the wind speed, to which the evaporation is very sensitive, is stronger off-equator than on the equator. [The evaporation is proportional to $U(q_s^* - q_a)$, where $U$ is the wind speed, $q_s^*$ is the sea surface saturation specific humidity, $q_a$ is the specific humidity at the lowest model level.] The large off-equator surface wind poses a great restriction on the variation of the local SST, preventing it from varying as freely as that over the equatorial Pacific. Weakened equatorial winds and cloud radiative forcing may also contribute to the enhanced equatorial SST response (Liu et al., 2006).

The most prominent feature in the North Pacific SST change is an intensified warming near the annual mean SST front associated with the Kuroshio and its extension. Similar warming is also found in the Atlantic counterpart near the Gulf Stream. The SST
anomalies in the MRI\_CGCM2\_3\_2a model have been attributed to the northward shift of the northern boundary of the oceanic subtropical wind-driven gyre, which in turn is driven by the changes in the atmospheric circulation (Sato et al., 2006, JMSJ). Similar northward expansion of the oceanic subtropical gyre is also noticed in the CCCma coupled model under doubling and quadrupling CO2 forcing (Saenko et al., 2005). This is probably also the case for the ensemble mean changes, as can be inferred from the multi-model ensemble mean sea level pressure (SLP) changes in Figure 2. Comparing to the climatology of the SLP, the pattern of SLP anomalies tend to shift the subtropical high poleward in both hemisphere and both Pacific and Atlantic basins, consistent with the robust poleward expansion of the Hadley cell found by Lu et al. (2007). Over the North Atlantic basin, the tripole-like SST pattern is also consistent with the NAO-like ensemble mean response in the atmospheric circulation. The enhanced cooling centered at 55°N may also be related to the slow-down of the Atlantic meridional overturning circulation in response to global warming (IPCC AR4 report, 2007).
Figure 2 (a) multi-model ensemble mean least-squared linear trend in SLP during 2001-2100 based on 17 model simulations of the A2 scenario; (b) count of models out of the 17 models that simulate an increase in SLP. (to add mean contours for reference)

Many features in the Southern Hemispheric mid-latitude ocean can be understood in terms of response of the ocean to the changes in the atmospheric circulation.

According to the changes of the SLP, the mid-latitude westerlies intensify and shift poleward, and the latitudes of zero wind stress curl shift poleward as well (not shown). Therefore, the wind driven Antarctic Circumpolar Current (ACC) should move poleward.
However the warming near the Antarctic continental shelves inhibits the poleward part of the ACC moving southward, resulting a narrowing of the ACC. The most prominent feature in the Southern Ocean SSTs is the circumpolar belt of suppressed warming centered at about 55S. At face value, this should be related to the intensification of the wind at the same latitudes, which facilitates loss of latent heat flux up into the atmosphere. Of course, other processes, such as the shift of Ekman pumping pattern, oceanic mixing and heat advection, and other energy exchange across the air-sea interface, should play parts in these SST anomalies. Only a full heat and momentum budget can holistically account for the SH SST changes, a task out of the scope of this study.

4. Zonal mean circulation

The analysis in this section is based primarily on simulations with the GFDL CM2.1 model. Nevertheless, the features discussed are not unique to this model and have been found in the majority of the IPCC models or in observations. Additionally, for brevity, in some plots only the DJF season is presented. But many of the features shown about the GHG-induced warming are not limited to NH winter, as will be stated specifically in the relevant parts of the text.

There is a conventional wisdom in the community that tropical SST anomalies are much more efficient than mid-latitude SSTs in driving mid-latitude teleconnections, and possibly zonal mean circulations as well. If the enhanced tropical SST dominates, one may expect that the zonal mean atmospheric response to the global warming should be somewhat reminiscent of the response to the El Nino.
Shown in Figure 3 are the El Nino minus La Nina composites for the air temperature, zonal mean zonal wind, and the tropopause pressure level, those being derived from the difference between 14 warm years and 12 cold years. The inter-hemispheric symmetry of both temperature and zonal wind bears great resemblance to regression patterns associated with the observed El Nino (Seager et al., 2003; L’Heureux and Thompson). As discussed by Seager et al., the warming in the deep tropics can be plainly understood as a consequence of the increase in tropical Pacific SST and the anomalous heat flux from ocean to the atmosphere; the maximum warming at the upper tropical troposphere, a structure also found in the observations, can be related to the quasi-moist adiabatic behavior of the tropical atmosphere in responding to the increase in moist static energy in the boundary layer. The tropospheric cooling in mid-latitudes between 25 and 40 degrees in NH and 35 and 50 in SH is produced by eddy-driven upward motion, further evidence for which will be provided later in the paper. Associated with the temperature anomalies, the tropopause level increases in the deep tropics, but decreases near the boundary of the tropics, if the tropics is defined as the latitudinal extent within which the tropopause exceeds a certain threshold height. According to this definition, the tropics narrow under El Nino conditions.
Figure 3  Temperature in °K (top), zonal wind in m/s (middle), and tropopause height in hPa (bottom) for the El Nino-La Nina composite (left); and the trend (right). Shadings represent the anomalous fields, contours the climatological mean fields. The blue (red) lines in the bottom panels are the pressure levels of the tropopause for the La Nina (El Nino) composite and 2001-2020 (2081-2100) mean; the dashed lines are the difference of El Nino minus La Nina (left) and 2081-2100 minus 2001-2020 (right), respectively, offset by 300hPa for the display purpose.

As in observations, the tropospheric zonal wind intensifies (weakens) near the equatorward (poleward) flank of the jet, resulting in a strengthening and equatorward
shift of the jet. The zonal wind changes poleward of 20N or 30S have a quasi-barotropic structure. These features have been interpreted (by Robinson 2002 (GRL) and Seager et al., 2003) as a chain of feedbacks involving the subtropical critical surface (where the phase speed of the eddy approaches the zonal wind speed and wave breaking occurs) for the eddy propagation. The strengthening of the subtropical jet on its equatorward flank can be thought of being driven by the tropical heating. If eddy phase speeds from midlatitudes remain constant, an increase in the subtropical winds implies that the location of the critical latitude, where EP fluxes are absorbed and the mean flow is decelerated, is pushed equatorward. Both the Hadley cell boundary and the midlatitude jet stream are then pushed equatorward as well, after eddy-mean flow feedbacks occur. While Seager et al (2003) justify this mechanism by diagnoses of baroclinic wave propagation in a quasi-geostrophic model, we will show that the picture is consistent with the response in the full spectrum of eddy momentum fluxes.

The temperature response to global warming (Figure 3b) is in stark contrast to the response to El Nino, although both show warming in the tropics and accentuated warming in the upper tropical troposphere. The midtropospheric temperature warms most at the latitudes where the El Nino tends to cool most. Thus, the associated meridional temperature gradient in the subtropical-to-mid-latitude midtroposphere is exactly opposite to that associated with the El Nino. The temperature structure near the tropopause projects upon a ubiquitous increase in the tropopause height with enhancements near the boundaries of the tropics (Figure 3f). Therefore, in a warmed climate, the tropics as defined by the tropopause broadens, in stark contrast to El Nino conditions. Consistent with the temperature changes, the upper-level subtropical wind
intensifies, which could potentially bring the critical latitude and the eddy-driven wind equatorward. However, the mid-latitude jet stream shifts poleward (Figure 3d). The mechanism for the jet shift and the expansion of the Hadley cell is the focus of this study and will be elaborated upon later. The maximum subtropical to midlatitude warming in the troposphere, the poleward shift of the jet, and the broadening of the tropics appear to be ubiquitous to both hemispheres in all seasons.

DJF 500hPa geopotential height also shows some interesting distinctions between the El Nino composite and GHG-induced trend (Figure 4). The GFDL CM2.1 can realistically capture the canonical Pacific-North America teleconnection response to El Nino, with low anomalies extending downstream to the America and mid-latitude Atlantic (Figure 4a). In the Southern Hemisphere, the wave train from the northwest of Australia to the Ross Sea forms a pattern symmetric about the equator with its Northern Hemisphere counterpart; it resembles markedly the ENSO teleconnection observed in the Southern Hemisphere (Fig.2 in Garreaud and Battisti, 1999). The mid-to-high latitude response is more zonally symmetric outside of the South Pacific. In the zonal mean 500hPa height field, El Nino produces a mid-latitude trough and a high to the poleward side in each hemisphere. This is exactly opposite to the global warming response (Figure 4b), which is characteristic of a mid-latitude ridge and a high-latitude low (especially in the Southern Hemisphere), a pattern that projects positively on to the Annular Mode in each hemisphere.
Figure 4  DJF 500hPa geopotential height field (in meters) for (a) El Nino-La Nina composite; (b) trend. The right side panels are the zonal-mean of the corresponding field.

Contrasting the zonal mean mass streamfunction (Figure 5) during the peak phase of El Nino versus La Nina (DJF months), the Hadley cells in both hemispheres tend to intensify and contract equatorward in the tropics, the ascending branch of the Hadley cell south of the equator slightly displaces northward, and the Ferrell cells move equatorward as well, especially in the Southern Hemisphere. The vertical motions as indicated by the streamfunction match well with the diabatic heating in both tropics and extratropics (Figure A1). While the vertical motions within the tropical cells can be well explained by the convective heating, the balance between the dynamical motion and the diabatic heating in the extratropics does not necessarily mean that the former is driven by the
latter. Adding a tropical wind anomaly onto a linearized quasigeostrophic model, Seager et al. (2003) showed that eddy dynamics, without explicit diabatic processes, can produce a mid-latitude cooling through Rossby wave propagation and refraction. Thus, the hemispherically symmetric mid-latitude cooling (near 30°N in the NH, 45°S in the SH) and the corresponding ascending motion should likely be interpreted as being eddy-driven. Similar argument can be applied to the high latitude warming and descent.

![Figure 5](image_url)

**Figure 5** DJF Zonal mean mass streamfunction for (a) El Nino-La Nina composite; (b) trend. Contours are the climatological mean streamfunction. Unit=1×10⁹ Kg/s

The meridional overturning response to GHG forcing is characterized by a weakening and poleward expansion of the Hadley cell and a poleward displacement of
Ferrell cell. Another important feature, which is consistent with the tropopause rise, is that the Hadley circulation tends to grow deeper under global warming so that its upper level divergent branch carries more potential energy. Both the weakening and the poleward expansion of the Hadley cell under global warming appear to be robust among the climate models those participate the IPCC AR4 (L07). The weakening of the tropical atmospheric circulation has been argued to be thermodynamically rooted (Knutson and Manabe, 1995; Held and Soden, 2006). In the tropics, the free atmospheric temperature is close to the moist adiabat. On a moist adiabat, dry static stability measured by $\frac{\partial \theta}{\partial p}$ averaged over the troposphere is proportional to the boundary layer mixing ratio $q$. Thus, the static stability in the tropics increases with the surface temperature following Clausius-Clapeyron relation as the column integrated moisture does (i.e., $\delta (\frac{\partial \theta}{\partial p})/(\frac{\partial \theta}{\partial p}) = \frac{\delta q}{q}$). The leading order thermodynamical balance in the upward branch of the Hadley cell is between the diabatic heating and the dynamical cooling associated with the updraft, i.e., $Q = \omega \frac{\partial \theta}{\partial p}$, or in the form of fractional change, $\frac{\delta Q}{Q} = \frac{\delta \omega}{\omega} + \delta \left( \frac{\partial \theta}{\partial p} \right) / \left( \frac{\partial \theta}{\partial p} \right)$. Neither of the primary components of the diabatic heating (radiative cooling or condensational heating) increases as rapidly as the increase in stability (i.e., $\frac{\delta Q}{Q} < \frac{\delta q}{q}$). In fact, over the warm pool region where the mean ascending motion takes place, a significant amount of the GHG-induced increase of heating by condensation is compensated by enhanced radiative cooling (Knutson and Manabe, 1995). Therefore, the fractional change in diabatic heating cannot keep up with the fractional change of static stability, and as a consequence, the tropical ascent must weaken at the rate $\frac{\delta \omega}{\omega} = \frac{\delta Q}{Q} - \frac{\delta q}{q}$. The mechanism for the expansion of the Hadley circulation is the subject of focus of the next section.
5. Energy budget and hydrological response

5.1 Energy budget at the surface and TOA

It has been well known that, during El Nino events, the surface flux from the ocean to the atmosphere fuels intensified convection and divergent atmospheric energy transport. Indeed, the net heating for the tropical column is predominantly caused by the flux coming from the surface. Over a typical El Nino winter (DJFMA) with a CTI of 1K, the ocean by average, heats the atmosphere globally about 1W/m² more efficiently than a typical La Nina winter with a CTI of -1K in this model. But 75% of the extra heating gained from the ocean is emitted to the space by the atmosphere; only 25% is retained by the atmosphere, an amount corresponding to 10W/m² of net heating at the equator since it is not uniformly distributed. A same amount of warming at the central and eastern equatorial Pacific under GHG forcing only gives rise to much less net heating (~6W/m²) to the equatorial atmosphere. Note that the zonal mean equatorial SST anomaly under the GHG forcing is even larger than that of the El Nino minus La Nina composite. Moreover, the anomalous globally averaged fluxes induced by the GHG forcing are downward at both the lower and upper boundaries of the atmosphere—in exactly opposite direction to those induced by El Nino (see the legend in Figure 6). The difference sources of the heat/radiative fluxes between El Nino and GHG warming do play out differently in the heating profile in the interior of the atmosphere. For the former case, the heating from below can be efficiently translated into condensational heating in the troposphere (Appendix Figures B1), which is associated with stronger vertical motion and tropical overturning circulations. For the latter, the heating of the atmosphere originates from the
trapping of long wave by the GHGs, which gives rise to a top-heavy heating profile in the deep tropics (Appendix Figures B2). This upper-level heating warms the upper tropospheric air and forces it to penetrate deeper is associated with deeper penetration of convection, which causes the Hadley cell outflow to gain larger dry static energy before traveling poleward, thereby increasing the efficiency of the poleward transport of moist static energy in the Hadley cell. The contrast in the heating rate of the tropical atmosphere between the El Nino and global warming may have some bearings as to why the Hadley cell intensifies in the former and weakens in the latter condition. Additionally, the same tropical SST warm anomalies under the GHG forcing, compared to those occur during the El Nino, is associated with a much smaller amount of the net energy surplus in the tropical atmosphere that needs to be compensated by the circulation.
Figure 6 Radiative and heat budget during DJFMA for (a) El Nino-La Nina composite; (b) trend (to add legend for the globally averaged flux at toa and surface). The red line indicates the net surface flux; the blue net toa flux; and the black the sum of toa and surface fluxes.

4.2 P-E response

In accordance with their distinct circulations, the hydrological response also demonstrates some interesting contrast between El Nino and global warming conditions (Figure 7). For both cases, the precipitation minus evaporation (P-E, DJFMA) increases at the equator and decreases over its neighboring latitudes. In this sense, the zonal mean hydrological
response to global warming is somewhat like El Nino, but the magnitude in the former case is at most half of that in the later, despite that both curves have been scaled to correspond to a 2K CTI discussion from climatology. The increased equatorial precipitation and decreased subtropical precipitation with global warming is due to increased humidity content (Allen and Ingram 2002; Held and Soden 2006), and happens despite a decrease in the Hadley circulation. The hydrological cycle response to El Nino, on the other hand, is amplified by the Hadley circulation change. In the latitudinal band between 25°-45°N in the NH and 35°-55°S in the SH, El Nino generates an above-normal P-E while global warming produces a deficit in P and P-E; as such, the former displaces the outer boundary of the subtropical dry zone (defined as area where P-E<0) equatorward, while the latter displaces the boundary poleward.

Figure 7 P-E for El Nino-La Nina composite (blue); (b) trend (red). For reference purposes, the climatologically mean P-E is plotted (black).

The poleward expansion of the boundary of the subtropical dry zone (defined as the latitude where P=E) is tightly tied to the expansion of the subsidence area of the
Hadley cell. Making use of 38 simulations for three future climate change scenarios (A2, A1B, and B1), Lu07 showed that predominant amount (85% for the southern hemisphere and 72% for the northern hemisphere) of variation in the displacement of the dry zone across the simulations can be explained by a linear relation to the displacement of the outer boundaries of the Hadley cell (see their Figure 2). For the complex climate systems portrayed by the coupled climate models, while it may seem difficult to argue for a cause-effect relationship between the expansion of the dry zone and the Hadley cell, we attempt to probe possible mechanisms that do not directly depend on the diabatic heating and therefore, arguably, hold dynamical responsibilities for the changes in both the Hadley cell and the subtropical dry zone.

6. Possible causes for Hadley cell expansion

6.1 Criticality for baroclinic waves growth

Our understanding of the extent of the Hadley cell has been guided by two alternative views on dynamical and thermodynamical control of the HC. The classic inviscid theory for axisymmetric circulation by Held and Hou (1980) predicts that the meridional extent of the HC scales as

\[ \phi_H \sim \left( \frac{g H_t}{\Omega^2 a^2} \frac{\Delta h}{\theta_0} \right)^{\frac{1}{2}}, \]  

(6a)

where \( H_t \) is the height of the tropical tropopause, \( \theta_0 \) is global mean temperature, \( \Delta h \) is the equator-to-pole surface potential temperature difference in radiative equilibrium, and other parameters have their conventional meanings. This scaling relation, which shows no explicit dependence on static stability, is derived based on assumptions that the upper
tropospheric wind is angular momentum conserving and that the heating and cooling for the atmosphere are balanced within the Hadley cell. The second view sees the width of the HC as being determined by the poleward extent to which angular-momentum conservation continues until the resulting vertical shear becomes baroclinically unstable [Held, 2000]. Thus, an alternative scaling can be derived from solving for the latitude where the angular momentum conserving zonal wind first satisfies the Phillips’ criterion for baroclinic instability (Phillips, 1954):

$$\phi_H \propto \left( \frac{NH_e}{\Omega a} \right)^{\frac{1}{2}}$$  \hspace{1cm} (6b)

Here, $H_e$ is the local tropopause height, $N$ the vertically averaged Brunt-Väisälä frequency, indicative of the tropospheric gross static stability. If the scaling relation (6a) applies, variations of the HC width should be proportional to the temperature gradients and the tropical tropopause height. On the other hand, if the scaling (6b) applies, one may expect the extent of the Hadley circulation to be insensitive to temperature gradients, and sensitive to the gross stability and the tropopause height near the poleward boundary of the circulation.

In the situations we consider here, ENSO variability provides a convincing disproof of scaling relation (6a). When El Nino occurs, both the meridional temperature gradient ($\Delta_h$) and the tropical tropopause height ($H_t$) increase, whereas the extent of the HC decreases. Based on simulations of the A2 scenarios by multiple models, Lu07 investigated the applicability of the two scaling theories and found that scaling relation (6b) is better model for the extent of the HC in the present-day climate. Thus, the extent of the HC can be interpreted as being limited by the latitude at which the thermally driven wind first becomes baroclinically unstable.
To further verify the physical mechanism implicated by the second scaling theory, we first relax the assumption of angular momentum conservation and take into account of the effect of wind shear; and then evaluate the baroclinicity only in the lower half of the troposphere. Now the metric for evaluating the baroclinic criticality becomes

$$C \equiv \frac{f^2(u_{500} - u_{850})}{\beta g H_e (\theta_{500} - \theta_{850}) / \Theta_0},$$  \hspace{1cm} (7)$$

where the vertical wind shear and potential temperature difference are taken between 500 and 850 hPa, and $H_e$ is simply set to be a constant (5000 meter). In deriving (7), Phillips’ 1954 instability criterion is used and only the lower troposphere is focused on because eddy growth is more sensitive to the lower- than to upper-level baroclinicity (Held and O’Brien, 199??). Defined in this way, $C$ is equivalent to a non-dimensional, integrated lower-tropospheric Eady growth rate (Hoskins and Valdes, 1990), except slightly more sensitive to the static stability than the latter, but the qualitative behavior is the same. Our preference to (7) over the Eady growth rate per se is motivated by the success of the scaling (6b) in predicting the extent of the Hadley cell in responding to widely varied SST boundary conditions (Frierson et al., 2007). Again, taking advantage of the simulations of global warming scenarios by various IPCC models (9 from A1B, 14 from A2), we examine to what extent the scatter in the expansion of the HC across models can be accounted for by the difference in the reduction of the subtropical criticality $C$.

In Figure 8a, the displacements of the southern hemisphere (SH) summer (DJF) boundary of the Hadley cell during the 21th century (i.e., 2081-2100 minus 2001-2020) are plotted against the changes in the baroclinic criticality $C$ averaged over a 25°-wide zonal band to the immediate equatorward side of the mid-latitude jet, area where eddies are believed to play crucial role in defining the boundary of the HC. The SH
displacements of the Hadley cell boundary in a warmer climate are strongly correlated with changes in the baroclinic criticality, with over 65% of the spread in displacements being explained by a linear relation to the variations of the subtropical baroclinicity in the 23 simulations. The models with larger reduction in \( C \) tend to produce larger HC expansion, with all models showing reduction in \( C \) in SH. The regression line for the HC boundaries intercepts the \( C=0 \) line at about 0 (-0.01 actually), implicative of that the SH HC boundary does not move without changing baroclinicity during the DJF months. The reduction in \( C \) can be decomposed into contribution from the decrease of vertical wind shear, i.e.,

\[
\delta C_{sb} = \frac{f^2 \delta (u_{500} - u_{850})}{\beta g H_{e1} (\theta_{500} - \theta_{850})_1 / \Theta_0}, \tag{8a}
\]

and increase of static stability, i.e.,

\[
\delta C_{st} = -\frac{f^2 (u_{500} - u_{850})_1 \delta (\theta_{500} - \theta_{850})}{\beta g H_{e1} (\theta_{500} - \theta_{850})_1^2 / \Theta_0}. \tag{8b}
\]

The subscript 1 in (8ab) indicates the values calculated from the period 2001-2020 and the \( \delta \) term the difference of period 2081-2100 minus 2001-2020. In SH, changes in both static stability and the wind shear contribute to the reduction of the baroclinicity with the former being dominant (Figure 8bc). The increased static stability is a well established fact in the climate models as a consequence of the quasi-moist adiabatic adjustment to the surface warming over the better part of the globe (Frierson, 2007, GRL).

For the NH DJF months, the poleward expansion of the HC is much less correlated with the changes in the subtropical baroclinicity (Figure 9a), likely owing to factors other than baroclinic eddies, such as stationary waves and large land-sea contrast, which can exert significant influence on the overturning circulation. Still, the correlation
between the displacements of the NH HC boundaries and the variations in the baroclinicity is significant beyond 95% confidence level (Figure 9a); the dominance of the static stability is also discernable (Figure 9b,c).

Figure 8  Relationship between the change in Phillips’ criticality $\delta C$ and the expansion of the Hadley cell in SH DJF. (a) total change of $\delta C$ due to both stability and wind shear; (b) $\delta C_{st}$ due the change of stability alone; (c) $\delta C_{sh}$ due to change of wind shear alone.
The intimate connection of the HC extent to the eddy dynamics in SH summer is also manifested in its relationship to the eddy driven jet. Figure 10 presents a scatterplot of the displacements of the HC boundaries during the 21st century versus those of the eddy-driven jet position, measured by the latitude of the maximum surface zonal wind ($Y_{max}$). The usage of $Y_{max}$ may be justified by angular momentum balance, that the vertically integrated eddy momentum flux convergence is largely balanced by the surface drag in mid-latitudes (except for the relatively small contributions from mountain torque and gravity wave drag), and thus the eddy-driven jet plausibly collocates with the maximum surface zonal wind. The HC boundary tend to move in concert with the eddy-
driven jet in the SH, more so than in the NH. This clearly corroborates the previous notion that mid-latitude eddies play key role in setting the boundary of the Hadley circulation, particularly in the SH summer.

![Figure 10](image)

**Figure 10** Relationship between the expansion of the Hadley cell and the shift of the surface wind in SH during DJF.

Interestingly, the significant correlation between the Hadley cell boundary and the jet position during SH summer also holds on interannual time scales within each individual model (not shown). From both observations and the model simulations of the GFDL CM2.1 model, we notice that El Nino forcing usually draws the subsidence of the Hadley cell equatorward. Based on the correlation relationship above, the eddy-driven part of the jet should also shift equatorward. Meanwhile, the meridional migration of the SH eddy-driven jet has been delineated by the so-called Southern Annular Mode (SAM), with a positive phase of the SAM representing a strengthened and poleward shifted eddy-driven jet. Put together, this chain of logics seems to suggest that ENSO forcing should project upon the SAM, with the warm phase of the ENSO forcing a negative phase of the SAM in the extratropical SH. Indeed, a significant linear relation has been detected
between the ENSO and the SAM in observations (L’Heureux and Thompson, 2006) and model simulations (Chen and Lu, 2007) for the events since 1979. The mechanisms of the interaction between the tropical heating and the extratropical zonal-mean circulation are yet to be understood, and will be addressed next in the context of eddy momentum flux spectrum, in comparison with the mechanisms in the response to global warming.

6.2 Eddy momentum flux spectrum

a. Streamfunction diagnosis

To further elucidate the roles of eddies in forcing the meridional circulation, we first carry out a simple budget diagnosis for the meridional mass streamfunction based on the zonal mean zonal momentum equation under small Rossby number approximation

\[ \frac{\partial \bar{u}}{\partial t} = \bar{f} \bar{v} + \bar{F}_\lambda - \frac{1}{a \cos \phi} \frac{\partial (u'v' \cos^2 \phi)}{\partial \phi}, \]

(9)

and the zonal mean continuity equation

\[ \frac{1}{a \cos \phi} \frac{\partial (\bar{v} \cos \phi)}{\partial \phi} + \frac{\partial \bar{\omega}}{\partial p} = 0 \]

(10)

For the upper troposphere in an equilibrium state, one may neglect the friction term \( F_\lambda \) and the time tendency of the zonal wind in equation (9). Based on (10), a meridional mass streamfunction \( \psi \) can be defined as

\[ \bar{v} = -\frac{1}{\cos \phi} \frac{\partial \psi}{\partial p} \]

(11a)

\[ \bar{\omega} = \frac{1}{a \cos \phi} \frac{\partial \psi}{\partial \phi} \]

(11b)

Substituting (11a) into (9) through \( \bar{v} \) and integrating (9) from \( p=0 \) to \( p=500\text{hPa} \), we derive an expression for the 500hPa streamfunction
\[ f\psi_{500} = \frac{\cos\phi}{a\cos^2\phi} \frac{\partial(<u'v'> \cos^2\phi)}{\partial\phi}, \]  

(12a)

or

\[ -2\Omega \tan\phi \psi_{500} = -\frac{1}{a\cos^2\phi} \frac{\partial(<u'v'> \cos^2\phi)}{\partial\phi}, \]  

(12b)

where \(<\cdot>\equiv \int_{500hPa}^{1000hPa} dp\). Thus, the meridional overturning circulation across 500hPa level is chiefly maintained by the convergence of the momentum flux by the quasigeostrophic eddies. Next, we examine how close the meridional overturning circulation is to the balance of equation (12).

Figure 11a shows the meridional distribution of the DJFM averaged transient eddy momentum convergence (solid) and the streamfunction \(-2\Omega \tan\phi \psi_{500}\) (dashed, in all panels) for the El Nino minus La Nina composite. Also plotted is the mean DJFM eddy momentum flux convergence (thick gray) and streamfunction (thick dashed gray, in all panels). The positive values of the momentum convergence indicate acceleration of the zonal wind. In line with the conventional view of the mid-latitude overturning circulation, there is a close correspondence between the streamfunction and the eddy momentum flux convergence poleward of 30°. The streamfunction crosses zero at 30°N and 37°S, defining the boundaries of the HC in the northern and southern hemisphere, respectively. 37°S is within the region dominated by eddy dynamics, while 30°N is at the transition between the dominance of the eddy dynamics and thermal forcing. This may have some bearing on the result that the Hadley cell extent scales well with the baroclinicity and location of the eddy-driven jet in the Southern Hemisphere, but not as well in the Northern Hemisphere. Comparing to the climatology, the equatorward shift of
the eddy momentum flux is evident in both hemispheres, consistent with the equatorward shift of the eddy-driven jet under the warm phase of ENSO. The structure of the anomalous streamfunction also conforms to that of the eddy momentum convergence in mid-to-high latitudes, implicative of a dynamical role of the latter in maintaining the former. In the northern hemisphere, the stationary waves play a significant role in transporting zonal momentum from the tropics to the mid-latitudes, by decelerating the subtropical winds and accelerating the mid-latitude winds (Figure 11b). While the anomalous stationary waves act to mitigate the background stationary wave-induced deceleration in the subtropics, their contribution to the shift of the mid-latitude momentum convergence seems to be secondary. Therefore, the shift of the boundary of the Hadley cell during DJF months is, in large degree maintained by the shift of the eddy momentum flux, with stationary waves and other factors (such as mean advection) playing a secondary role in the Northern Hemisphere.
Figure 11 The El Nino minus La Nina composite for the momentum budget for 500hPa streamfunction in DJFM. Gray lines show the climatological values; black the trend values. Dashed lines represent streamfunction \((-2\Omega \tan \phi \psi_{500}\)). Solid lines in (a), (b), and (c) represent the momentum convergence by transient eddies, stationary waves, and the sum of the two, respectively.
Figure 12 The trend of the momentum budget for 500hPa streamfunction in DJFM. The convention of the lines is the same as Figure 11.

A similar argument may be applied to the Hadley cell expansion under global warming. The boundaries of the anomalous overturning cells in the mid-to-high latitudes collocate with the latitudes of transition between the transient eddy-induced acceleration
and deceleration (Figure 12a), and much less so with the combined momentum convergence of both transient eddies and stationary waves (Figure 12c). The good correspondence between the anomalous overturning streamfunction and the transient eddy momentum flux corroborates the previous analysis of instability, that the suppressed baroclinicity in the subtropics pushes eddies and hence the eddy-driven circulation poleward. Thus, the expansion of the Hadley cell during the boreal winter, to some extent, is a manifestation (on the meridional plane) of the shift of the storm track and the eddy-driven jet. In other words, the mechanism responsible for the jet shift may also be accountable for the expansion of the Hadley cell.

The stationary waves also respond differently to the GHG forcing compared to the ENSO. The global stationary wave tends to weaken in amplitude under climate warming (Figure 12b), as discovered by Joseph et al., (2004) from GFDL’s R30 coupled ocean-atmosphere GCM. Not attempting to question the relevance of the more zonally localized tropical convection to this phenomenon (as discussed by Joseph et al.), we suspect that the zonalization of the circulation may share the same root as the weakening of the tropical atmospheric circulation—adjustment of the circulation to compensate the increase of moisture content with surface warming (Held and Soden, 2006; Vecchi and Soden, 2007).

Caution should be used in interpreting the results of momentum budget. The momentum balance does not tell the driving source for the variations of the Hadley cell. For the ENSO case, the driving source is already known to be the heating resulting from the dynamical interaction between the tropical ocean and atmosphere, and is hence of a tropical origin. As argued above, the jet shift and the expansion of the Hadley cell under
global warming may be originated from distinct processes, an example of which is the increased static stability over the subtropical and mid latitudes. Next, eddy momentum cospectra analysis will shed more light on this issue.

b. Eddy momentum flux spectrum

By decomposing the transient eddy momentum flux into components from discrete phase speeds, the different forcing mechanisms for the Hadley cell expansion between La Nina and global warming conditions can be uncovered via the intricate eddy-mean flow interaction near the critical latitudes. The cospectra of eddy momentum flux convergence at 250hPa, as a function of angular phase speed \( c_A = c / \cos \phi \) and latitude, are displayed in Figure 13 together with the zonal mean zonal wind divided by \( \cos \phi \). Near the equatorward flank of the jet, the mean spectra (shadings) are confined poleward of the critical latitudes, where \( c = U \). This implies that the critical latitude acts as a limit for the equatorward propagation of baroclinic waves, which tend to break 10-20 degrees in latitude before reaching their critical latitudes. As a consequence, the waves with smaller phase speed can propagate further into the tropics, while the waves with faster phase speed are restricted to higher latitudes. The confinement of the eddy activity by the critical latitudes evidences the relevance of quasi-linear wave theory to the real atmosphere.
Figure 13 Cospectra of eddy momentum flux convergence during DJFM. Contours: climatology; color: El Nino–La Nina composite. Black solid line: 250hPa U/cos during La Nina; dashed line: U/cos during El Nino. The side panel on the left hand side is composite of the surface wind. The unit of the spectra is m/s/day.

The anomalous cospectra associated with the composite of El Nino minus La Nina are characterized by a systematic equatorward shift of mid-latitude waves, whether
slow or fast. This is especially clear in the Southern Hemisphere. Qualitatively the same ENSO composite pattern is also derived from the ERA40 reanalysis (Chen and Lu, 2007). We attribute the equatorward displacement of the eddy momentum divergence at the equatorward flank of the spectra to the increase of the thermally-driven subtropical wind (dashed line) during the warm phase of the ENSO. The strengthening of $U$ at the subtropics draws equatorward the critical latitude for waves of all phase speed. As a result, the wave activity can penetrate deeper into the tropics than it otherwise would during the cold phase of the ENSO. Assuming the typical scale of eddies would not change, as it is the case here, the equatorward movement of the critical latitudes should result in the equatorward shift of the eddy momentum spectra as a whole, and hence the eddy-driven jet and meridional circulation. At the latitude band where the zonal mean subtropical wind increases in strength during the warm phase of the ENSO, the anomalous eddy momentum flux tend to mitigate the thermally-driven effect of accelerated subtropical winds by exerting a deceleration on the mean wind, causing by the wave breaking. In other words, the eddy momentum flux divergence serves as a damping force for the thermally driven subtropical wind in both the interannual variability and the climatology.

The eddy cospectra response to GHG forcing (Figure 14) shows some opposite features to the El Nino composite, that is, a dipole at mid-latitudes that shifts the mean spectra poleward. However, unlike the ENSO composite, no systematic shift occurs with the critical latitudes; instead, a salient feature accompanying the mid-latitude dipole is the center of the spectra moves towards higher phase speeds. Under the dynamical constraint of the critical latitude, eddies with faster phase speed are trapped in the higher latitudes,
and the associated eddy-induced momentum convergence and divergence also displace poleward accordingly. Similar changes in the spectra are also found in the 40-year trend from 1960-2000 in the same model, which has been attributed to ozone depletion and the associated stratospheric wind anomalies (Chen and Held, 2007 GRL). We are prompted to argue that increase in phase speed of the dominant eddies is a key ingredient of the poleward shift of the jet and eddy-driven overturning circulation, a notion further supported by a study using a shallow-water model, wherein increasing the phase speed of the upper-tropospheric stirring can indeed shift the jet poleward (Chen et al., 2007).
Figure 14 Cospectra of eddy momentum flux convergence during DJFM. Contours: climatology; color: trend. Black solid line: 250hPa U/cos during 2001-2020; dashed line: U/cos during 2081-2100. The side panel on the left hand side is trend in the surface wind. The unit of the spectra is m/s/day.

It remains to explain what causes the increase of the phase speed of the mid-latitude eddies. One probable candidate is the upper tropospheric/lower stratospheric
wind anomalies. Increasing CO$_2$ forcing cools the stratosphere and warms the
troposphere (Manabe and Wetherald, 1967). The water vapor feedback to the surface
warming can also reinforce the upper-level differential heating. Because of the downward
slope of the tropopause as one moves poleward, this differential heating enhances the
poleward temperature gradient and the associate thermal wind near the tropopause slope.
Idealized baroclinic eddy life cycle experiments have shown that the increased lower
stratospheric wind anomalies can accelerate the eastward propagation of tropospheric
eddies (Wittman et al 2007). Thus, the increase in eddy phase speed can offer an
extratropical source for the jet shift and Hadley cell expansion under global warming.

6.3 role of tropopause rise

Several studies (Haigh et al., 2005; Williams 2006; Lorenz and DeWeaver, 2007) have
found that the tropopause rise is the key ingredient for the poleward shift of the jet (and
hence the expansion of the Hadley cell, for cases with a single jet climatology, as during
the SH summer). By manipulating the prescribed radiative equilibrium temperature of the
stratosphere, these authors found that warming (cooling) the stratospheric temperature at
mid-to-high latitudes lowers (raises) the extratropical tropopause and causes the jets to
move equatorward (poleward) and the Hadley cell to shrink (expand), while warming
(cooling) only at the low latitudes result in opposite changes in the jet position and the
Hadley cell extent. Uniform warming (cooling) of the stratosphere can also generate a
significant equatorward (poleward) shift of the jets as well. Recalling that the tropopause
slopes downward from equator to pole, the uniform cooling or heating also introduces a
positive temperature gradient near the tropopause, accompanied by positive stratospheric
wind anomalies. While no physical mechanism is proposed in Lorenz and Dewaver (2007), the results of these idealized experiments are, at least, consistent with the importance of lower stratospheric wind anomalies.

On the other side of the token, a salient feature in the full model (CM2.1)—the reduction of the tropospheric eddy heat flux ($\nu^T$) over the subtropics, which is believed to be related to the suppressed tropospheric baroclinicity under global warming, is not reproduced by raising the tropopause in an idealized dry dynamical model (Lorenz and DeWeaver, 2007). The low level baroclinicity, a factor presumably independent of the tropopause height, may be at least, as important to the position of the jet and the poleward extent of the Hadley cell. Indeed, significant jet shift and expansion of the HC (Figure 15) are simulated in an experiment of imposing a uniform (5K) SST warming to a gray-radiation moist AGCM, where the variation of water vapor is not allowed to feedback to the radiation scheme. In this model, the radiative impacts of the moisture content associated with the SST warming is minimized near the tropopause, and the tropopause rise is significantly less than in a full GCM. [The detailed configuration of the gray-radiation moist AGCM is documented in Frierson et al (2007); the detailed results of this experiment will be reported elsewhere.] Thus, to the extratropical response in this experiment, the change in the tropospheric baroclinicity, i.e., the mechanism discussed in section 4.1, should be more relevant. So far, two prominent dynamical factors—tropospheric baroclinicity and tropopause change—have been identified to be responsible for the poleward shift of the westerly jet and the expansion of the Hadley cell under global warming. But it remains to be seen which one is more substantial to the shift of the jet and the expansion of the Hadley cell in the global warming climate change, a question
we leave for future investigation. We believe that intelligently chosen simulations, either in more simplified domains or with more simplified models similar to the Frierson et al (2007) may be the best approach to distinguishing between these two theories.

**Figure 15** The response of (a) zonal wind (m/s); and (b) zonal mean mass streamfunction (1x10^9 Kg/s) to 5°K uniform SST warming in a gray-radiation model. The color shadings depict the anomalous response; the contours the climatology of the control run.

7. Summary and concluding remarks
Despite the fact that the GFDL CM2.1 model produces an El Nino-like tropical Pacific SST, a weakened Walker circulation, and a reduction of the zonal slope of the equatorial thermocline (Vecchi et al., 2006) in responding to increased GHG forcing, we find no analogy to the ENSO teleconnection in the extratropical atmospheric response. Instead, the mid-to-low level subtropical air temperature gradient decreases, the zonal mean mid-latitude westerlies shift poleward, the Aleutian low weakens and the subtropical highs move poleward, and the Hadley cell weakens and expands poleward—all in an opposite fashion to the responses to El Nino forcing.

To contrast global warming with El Nino, only the boreal winter months are analyzed. We find that, during the DJF months in SH, the displacements of the Hadley cell boundary are closely tied to the variations in the mid-latitude jet, both arguably being integral parts of one three-dimensional circulation, and likely sharing the same driving mechanisms. Two hypothesized mechanisms are proposed based on analyzing the simulations of the IPCC AR4 data and other modeling studies (Lorenz and DeWeaver, 2007, JGR; Williams, 2006; Haigh et al., 2005). A simple scaling based on Phillips’ criterion on baroclinic instability demonstrates that the poleward expansion of the Hadley cell and the shift of the eddy-driven jet are related to the reduction of baroclinic instability at the equatorward flank of the jet, which stabilizes eddy growth rates and pushes the eddy-driven jet and the eddy-driven subsidence towards the pole. The increased static stability, a well-established observational and model phenomenon from the quasi-moist adiabatic adjustment of the atmosphere to surface moistening in both the tropics (Xu and Emanuel 1989) and the midlatitudes (Juckes 2000; Frierson et al 2006; Frierson 2006; Frierson 2007b), seems to be a most fundamental factor in stabilizing the
subtropical eddy activity, while the decreased low-level meridional temperature gradient seems to play only a secondary role.

An alternative mechanism is proposed via varying the tropopause height in dry dynamical models by Lorenz and DeWeaver (2007) among others. By design, the tropospheric static stability in these model experiments remains basically intact, but the jet position and the boundary of the Hadley cell extent can displace significantly poleward in responding to the rise of tropopause height. This hypothesis is also corroborated by a more recent scaling analysis of the IPCC AR4 simulations (Lu et al., 2007). Space-time cospectral analysis of the eddy momentum fluxes reveals that the eddy-mean flow interaction near the critical latitude might be the key ingredient in the responses to tropopause change. A higher than normal tropopause is associated with an upper tropospheric warming and a stratospheric cooling. Because the tropopause height declines poleward at mid latitudes, the differential heating at either sides of the tropopause increases the meridional temperature gradient, giving rise to faster thermal winds, and consequently, faster wave speeds due to Doppler shifting. The fundamental character of the critical layer dynamics, which confines waves of faster phase speed more poleward, implies that wave spectra being shifted towards faster phase speeds should produce a poleward shift in the eddy momentum flux convergence and an associated shift in the eddy-driven jet and the eddy-driven Hadley cell subsidence.
Figure 16 The change in zonal mean northward atmospheric energy transport produced by 5°K uniform SST warming in a gray-radiation model. The dash-dot line indicates the change in moist energy transport; the dashed line the change in dry static energy transport; the solid the change in total energy, or moist static energy transport. The thick lines represent the climatological transport of dry static energy (dashed) and moist energy (dash-dot) from the control run, respectively, plotted for the purpose of reference. The climatological values of transport have been divided by 2 to be accommodated with the anomalous transport profiles.

A third hypothesis for the mechanisms of the jet shift and Hadley cell expansion, which is as appealing as the previous two, emerges from consideration of the poleward fluxes of energy. In the same 5K uniform SST experiment with a gray model discussed above, there is a prominent compensation between the changes in the poleward moist energy flux and dry static energy flux, as can be seen from comparing the black dash-dot
line with the dashed line of Figure 16. We notice a very similar compensation in uniform SST warming experiments with the GFDL aquaplanet AGCMs. An even cleaner compensation (with much smaller total energy flux than that indicated by the solid line in Figure 16) has also been found in the slab equilibrium response to the double CO$_2$ forcing (Held and Soden, 2006). The compensation is almost perfect in the gray model, even in the condition that the water vapor content is artificially increased by 10 times by increasing the saturation vapor pressure parameter in the Clausius-Clapeyron relation (Frierson et al., 2007).

Why could compensation of energy fluxes matter for the jet position? The mystery becomes unfolded by examining the fact that the maximum poleward moisture flux (peak in the gray dash-dot line near 35°) occurs on the equatorward side of the maximum eddy dry static energy flux (peak in the gray dashed line near 45°). As a result, the maximum compensation to the enhanced flux of moisture also occurs at the equatorward side of the eddy dry static energy flux (compare the negative peak of anomalous dry static energy flux versus the positive peak of the climatological flux). Since baroclinic eddy activity is usually highly associated with the dry static energy flux maximum (this is, for instance, where the vertical component of the EP flux vector is maximum), the suppression of eddy dry static energy flux on its equatorward flank is usually associated with a poleward shift of the storm track and eddy-driven circulations. However, this is not likely the dominant mechanism for the changes in the jet position and Hadley extent in the transient response of the fully coupled models to the increasing GHG forcing, since little compensation of dry static energy flux occurs in these simulations (Held and Soden, 2006). On the other hand, with the oceanic adjustment
becoming saturated, this effect may eventually come into play in later stages of the response to the GHG forcing.

So far, we have only focused on the responses during the boreal winter time. The boreal summer Hadley cells and mid-latitude jets show distinct features in their response to global warming in both hemispheres. During JJA months, the SH Hadley cell also undergoes robust poleward expansion, the extent of which is correlated to the extent of the subtropical stabilization as well. However, the SH mid-latitude jet during these months seem to be decoupled from the Hadley cell, no relationship can be identified between the shift of the jet and the expansion of the Hadley cell. In addition, there is more of a flavor of strengthening than shift in the eddy-driven winds (see also Lorenz and DeWeaver, 2007). We hitherto have found no clues (in static stability, tropopause change) that could lead to the understanding of the changes in the SH winter mid-latitude jet. In the NH summer, there is no consensus across the models as to whether the weak subsidence of the summer cell will move poleward or equatorward, an issue we must therefore leave out of discussion.

Despite the uncertainties regarding the NH summer, the robustness of the expansion in the annual mean Hadley cell is remarkable, and it must exert profound impacts on both the subtropical ocean and land in the warming climate. The impact on the zonal-mean subtropical hydrological cycle has been discussed by Lu et al. (2007), though only at a superficial level. It has also been argued (Seager et al., 2007) that the Hadley cell expansion is partially responsible for the projected aridity and water shortage in the NH subtropical land areas (the American Southwest and the Mediterranean Europe). It is also compelling to examine the impacts of the expansion on the subtropical
ocean circulation, and oceanic biological and ecological system. With regard to the
dynamics of the Hadley cell expansion and the poleward shift of the jet stream, questions
still remain as to the role of the feedback from the structures of ocean warming (such as
the accentuated warming near the extension of the Kuroshio, the temperature gradient in
the extratropical Southern Oceans), and more generally, the role of dynamical interaction
between the ocean and atmosphere near the subtropics and mid-latitudes. All these are
fruit-bearing areas of climate change research, and thus warrant further investigations.

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Reference

Appendix A  Eddy momentum flux cospectra

The space-time spectral analysis is a two dimensional harmonic analysis, and is most convenient if the spatial dimension is periodic such as a latitude circle. This approach is useful to detect modes of variability in which the spatial scales are correlated with the temporal scales, and therefore it is ideal to identify wavelike motions. The method
was first introduced in Hayashi (1971), and has been widely used for studies of the characters of waves in the atmosphere.

For the time series $v(\lambda, t)$ as a function of longitude $\lambda$ and time $t$, we first perform a Fourier transform in longitude at each time.

$$ v(\lambda, t) = \sum_m \text{Re}\{ (C_m(t) - iS_m(t)) \exp(i m \lambda) \} $$  \hspace{1cm} (A1)

where $m$ is the zonal wavenumber, and $C_m(t)$ and $S_m(t)$ are the cosine and sine coefficients. Then, we perform a Fourier transform to these coefficients in time.

$$ C_m(t) = \sum_{\omega} \text{Re}\{ (A_{m,\omega} - iB_{m,\omega}) \exp(i \omega t) \} $$  \hspace{1cm} (A2)

$$ S_m(t) = \sum_{\omega} \text{Re}\{ (a_{m,\omega} - ib_{m,\omega}) \exp(i \omega t) \} $$  \hspace{1cm} (A3)

The substitution of (A2) and (A3) into (A1) yields the

$$ v(\lambda, t) = \sum_m \sum_{\omega = 0} \text{Re}\{ \left( (A_{m,\omega} - b_{m,\omega}) - i(\mp B_{m,\omega} + a_{m,\omega}) \right) \times \exp(i(m \lambda \pm \omega t)) \} $$  \hspace{1cm} (A4)

Thus, we obtain the spectral power density and the phase for the time series $v(\lambda, t)$ in the wavenumber and frequency space.

$$ V_{m,\omega}^2 = (A \mp b)^2 + (\mp B - a)^2 $$  \hspace{1cm} (A5)

$$ \phi_{m,\omega} = \tan^{-1} \left( \frac{\mp B - a}{A \mp b} \right) $$  \hspace{1cm} (A6)

Where + and – correspond to westward and eastward propagation waves, respectively.

For each pair of longitude-time series $u(\lambda, t)$ and $v(\lambda, t)$ at certain latitude $\phi$, the cospectra power density is defined as
\[ K_{m,z,u}(u,v) = 2[(A^v \mp b^v)(A^u \mp b^u) + (\mp B^v - a^v)(\mp B^u - a^u)] \]  \hspace{1cm} (A7a)

or

\[ K_{m,z,u}(u,v) = 2\sqrt{V_{m,z,u}} U_{m,z,u} \cos(\phi^v_{m,z,u} - \phi^u_{m,z,u}) \]  \hspace{1cm} (A7b)

and the quadrature spectrum is

\[ Q_{m,z,u}(u,v) = 2\sqrt{V_{m,z,u}} U_{m,z,u} \sin(\phi^v_{m,z,u} - \phi^u_{m,z,u}) \]  \hspace{1cm} (A8)

Following Randel and Held (1991), the spectral power density in wavenumber-frequency space is then transformed to wavenumber-phase speed space (or angular phase speed \( c_A = a_e \cos \phi \omega / m \) space (or angular phase speed \( c_A = a_e \omega / m \)). The power spectra in phase speed space are defined such that the total covariance is conserved.

\[ K_{m,z,c} \cdot \Delta c = K_{m,z,u} \cdot \Delta \omega \quad \text{or} \quad K_{m,z,c_A} \cdot \Delta c_A = K_{m,z,u} \cdot \Delta \omega \]  \hspace{1cm} (A9)

Thus, we obtain

\[ K_{m,z,c} = K_{m,z,u} \cdot \frac{m}{a_e \cos \phi} \quad \text{or} \quad K_{m,z,c_A} = K_{m,z,u} \cdot \frac{m}{a_e} \]  \hspace{1cm} (A10)

In practice, the \( K_{m,z,c} \) (or \( K_{m,z,c_A} \)) is calculated by constructing a phase speed grid and interpolating the spectral density at each phase speed from the two closest frequency estimates of \( K_{m,z,u} \), then multiplying by \( \frac{m}{a_e} \) (or \( \frac{m}{a_e \cos \phi} \)).

In practice, the time series is first tapered by a Hanning window

\[ w(t) = 0.5 + 0.5 \cos(\pi t / T) \quad \text{(where} \ 0 \leq t \leq T \text{).} \] The power spectrum is additionally smoothed by a normalized Gaussian spectral window of the following form at a given frequency,
Appendix B  Comparison of atmospheric heating rate between El Nino and global warming

\[ W(\omega - \omega_0) = e^{-[(\omega - \omega_0)/\Delta \omega]^2}, \quad (A1) \]

with \( \Delta \omega = 3 \).

**Figure B1** The anomalous atmospheric heating rate for El Nino minus La Nino composite. (a) condensation; (b) radiation; (c) diffusion; (d) sum of all above (balanced by the dynamics).
**Figure B2** Same as Figure B1 except for the difference between 2081-2100 and 2001-2020.