Atmospheric Circulation Response to An Instantaneous Doubling of Carbon Dioxide Part II Atmospheric Transient Adjustment and its Dynamics

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ABSTRACT

The dynamical mechanisms underlying the transient circulation adjustment in the extratropical atmosphere after the instantaneous doubling of carbon dioxide is investigated in the National Center for Atmospheric Research Community Atmospheric Model Version 3 coupled to a slab ocean model. In Part I it was shown that the transient process is important in setting up the extratropical circulation response in equilibrium. Three phases are found in the transient process in the Northern Hemisphere: 1) a radiatively-driven easterly anomaly in the stratosphere, 2) a stratospheric westerly acceleration as a result of anomalous planetary-scale eddy momentum flux convergence, and 3) a ‘downward migration’ of the westerly acceleration from the lower stratosphere to the troposphere. Several other possible mechanisms for inducing the poleward movement of the tropospheric jet streams are examined. No significant increase in eddy phase speed is found. The rise in tropopause height appears to lead the tropospheric jet shift but no close relation is observed. The length scale of transient eddies does increase but after the tropospheric jet has shifted. The poleward displacement of the tropospheric jet is found to be largely driven by changes in the index of refraction due to the westerly acceleration in the extratropical lower stratosphere.
1. Introduction

Comprehensive climate models for the Coupled Model Intercomparison Project phase 3 (CMIP3)/Intergovernmental Panel on Climate Change the Fourth Assessment Report (IPCC AR4) have projected many changes in the general circulation of the atmosphere in response to increased carbon dioxide (CO$_2$) concentration. Yin (2005) found a consistent poleward and upward shift of the midlatitude storm tracks along with the poleward shifts of the surface wind stress and the precipitation zone. The changes in the location and intensity of the storm tracks are also closely related to the poleward displacement of the tropospheric zonal jets (Kushner et al. 2001) and the poleward expansion of the Hadley Cell (Lu et al. 2007); however, what causes these circulation changes is not entirely clear.

In this study the atmospheric circulation response to increased greenhouse gases is investigated using the National Center for Atmospheric Research (NCAR) Community Atmospheric Model Version 3 (CAM3) coupled to a slab ocean model (SOM) when the carbon dioxide (CO$_2$) in the atmosphere is instantaneously and uniformly doubled. While the CMIP3/IPCC AR4 climate models more realistically gradually increase the CO$_2$ concentration, these simulations are always in quasi-equilibrium and don’t provide evidence on how and why the general circulation of the atmosphere adjusts to the external CO$_2$ forcing. The methodology used in this study allows for a step-by-step assessment of the cause and effect of the changes in the circulation that occur in response to increased greenhouse warming.

The model description and experiment design were presented in Wu et al. 2011 which is Part I of this two-part study. In Part I it was demonstrated that the simulations approximately reach quasi-equilibrium after about 20 years of model integration after the instantaneous CO$_2$ doubling and that the quasi-equilibrium responses resemble that from the CMIP3/IPCC AR4 coupled climate models under the A1B global warming scenario. In fact, most of the features, such as the enhanced tropical and subtropical upper tropospheric warming and the poleward shift of the tropospheric jets and the midlatitude storm tracks, are well established after a few months of model integration. It has been widely recognized
that the enhanced upper tropospheric warming is closely related to the circulation responses such as the poleward displacement of the tropospheric jet streams and the midlatitude transient eddies (e.g., Wu et al. 2010; O’Gorman 2010; Butler et al. 2010; Rivire 2011), and some studies, for example, Butler et al. (2010) and Rivire (2011) assumed the extensive upper tropospheric warming as the forcing of the circulation response. Whether this is true or not was investigated by looking into the temperature tendency (diabatic vs. adiabatic) in our instantaneous CO$_2$ doubling experiments. In Part I it was shown that the broad upper tropospheric warming expansion in the subtropics is a consequence of the circulation change (rather than the cause) and is primarily dynamically-driven by the intensification of transient eddy momentum flux convergence and resulting anomalous descending motion in this region. Part I also analyzed the day-by-day response of the jet streams and it was shown that the poleward displacement of the tropospheric jets occurs after the intensification of the subpolar westerlies in the stratosphere and the enhancement of the tropospheric transient eddy momentum flux convergence. This ‘downward migration’ process is similar to Baldwin and Dunkerton (2001) who demonstrated using reanalysis data that extreme events in the stratosphere are followed by anomalous weather regimes in the troposphere. This similarity suggests the importance of the stratosphere and its coupling with the troposphere in the circulation adjustment in our model experiments.

As for global warming, a number of studies have suggested that the tropospheric circulation response to increased greenhouse gases critically depends on the stratosphere and its dynamical interaction with the troposphere. Sigmond et al. (2004) studied the separate climatic impacts of middle-atmospheric and tropospheric CO$_2$ doubling using the European Centre Hamburg Model (ECHAM) middle-atmosphere climate model with prescribed sea surface temperatures (SSTs). They found strengthened Northern Hemisphere (NH) midlatitude tropospheric westerlies as a consequence of a uniform CO$_2$ doubling everywhere in the atmosphere and attributed this mainly to the middle-atmosphere CO$_2$ doubling. Sigmond and Scinocca (2010) found that different stratospheric basic states (controlled by different
parameterization settings of orographic gravity wave drag) can result in distinct NH circulation responses to CO$_2$ increase.

In the present model experiments, features such as the poleward displacement of the tropospheric jet streams in the NH are established after a few months of model integration (see Wu et al. 2011). Figure 1 shows the zonal mean zonal wind anomaly in March of year 1 (transient) and in March of year 22 (equilibrium) in the instantaneous CO$_2$ doubling experiment with CAM3-SOM. March is chosen because of the tropospheric jet shift starting from March of year 1 in the experiment (shown in Part I). The zonal wind response in the transient and equilibrium state is similar in the NH extratropical atmosphere, for example, the westerly intensification in the stratosphere and the midlatitude jet shift in the troposphere. The westerly anomalies in the NH subtropical lower stratosphere and in the Southern Hemisphere (SH) later develop roughly after one year of model simulation (for example, in December of year 1 as shown in Figure 5 in Part I). Furthermore, it has been found that, in addition to the NCAR CAM3-SOM, a large proportion of the CMIP3/IPCC AR4 coupled models predict a stratospheric westerly acceleration in response to global warming, and the zonal flow response in March averaged among 22 CMIP3/IPCC AR4 models$^1$ is also shown in Fig. 1.

Figure 2 is the same as Figure 7(b) in Part I and shows the day-by-day evolution of the zonal mean zonal wind anomalies averaged in the extratropics between 30°N and 70°N during January-February-March-April (JFMA) of year 1 in the 100 member ensemble mean of model simulations for the case following the CO$_2$ doubling minus the case with control CO$_2$ concentration. It is shown as a function of time and pressure level and is smoothed with a 5-day temporal running average. Based on Fig. 2, we define three phases during this 120-day transient adjustment in the NH. Phase 1 roughly covers the first month after

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$^1$All 24 available CMIP3/IPCC AR4 coupled climate models are included except for the UKMO-HadGEM1 (no available output at 10mb) and the MIUB ECHO-G (no available output above 100mb). The difference is taken between 2081-2100 A1B scenario and 1961-2000 in the 20th century simulation.
the instantaneous doubling of CO$_2$ on January 1st and shows an easterly anomaly in the subpolar stratosphere. Phase 2 presents a transition into a westerly acceleration in the stratosphere which takes place in February. Phase 3 occurs during March and April and features a 'downward migration' of the westerly anomalies from the lower stratosphere to the troposphere and a poleward displacement of the tropospheric jet. In this paper, we analyze the dynamical mechanisms involved in each of the three phases, in particular, what drives the easterly (westerly) acceleration in the stratosphere in Phase 1(2) and what causes the descent of the anomalous westerly acceleration signal from the lower stratosphere to the troposphere and leads to the poleward displacement of the tropospheric zonal jets in Phase 3. There are several mechanisms that have been raised to understand the tropospheric circulation shift in response to global warming such as the increase in eddy phase speed (Lu et al. 2008; Chen et al. 2008), the rise in tropopause height (Lorenz and DeWeaver 2007) and the increase in eddy length scale (Kidston et al. 2010, 2011). We examine all the above possible mechanisms using our model simulations in order to assess whether or not they can explain the circulation changes seen in the experiments. This analysis provides a unified assessment of the possible mechanisms within the same framework.

In section 2 we introduce the diagnostic methodologies that have been used in this study. Section 3 presents aspects of the climatological simulation results from CAM3-SOM experiments. Section 4 analyzes the dynamical mechanisms underlying each of the three phases during the transient adjustment process. Conclusions and discussions are presented in section 5.
2. Diagnostic Methodologies

a. Eliassen-Palm (EP) Flux and Its Convergence

The quasi-geostrophic (QG) EP flux in spherical and pressure coordinate is defined as:

\[
F(\phi) = -\cos \phi (\langle uv \rangle - \langle u \rangle \langle v \rangle) \\
F(p) = a f \cos \phi (\langle v \theta \rangle - \langle v \rangle \langle \theta \rangle) / \langle \theta \rangle_p
\]

where \( f \) is the Coriolis parameter and \( \theta \) is potential temperature (Edmon et al. 1980). Eddy momentum flux and heat flux are denoted by \( \langle uv \rangle - \langle u \rangle \langle v \rangle \) and \( \langle v \theta \rangle - \langle v \rangle \langle \theta \rangle \), respectively, which include both transient and stationary waves, where the angle brackets in this study follow the same notation in Part I and denote zonal averages. The direction of the flux vectors, \((F(\phi), F(p))\), generally indicates the propagation of waves and the flux divergence, denoted by \( \frac{1}{\cos \phi} \nabla \cdot \vec{F} = \frac{1}{\cos \phi} \left\{ \frac{1}{\cos \phi} \frac{\partial}{\partial \phi} (F(\phi) \cos \phi) + \frac{\partial}{\partial p} F(p) \right\} \), measures the wave forcing on the zonal mean flow. The relationship between the wave forcing and the zonal mean flow is presented in the Transformed Eulerian Mean (TEM) framework and the TEM zonal momentum equation is written as:

\[
\frac{\partial \langle u \rangle}{\partial t} - \left[ f - \frac{1}{\cos \phi} \frac{\partial}{\partial \phi} (\langle u \rangle \cos \phi) \right] \langle v \rangle^* = \frac{1}{\cos \phi} \nabla \cdot \vec{F}
\]

where \( \langle v \rangle^* \) is the meridional component of the residual mean circulation defined as \( \langle v \rangle^* = \langle v \rangle - \frac{\partial}{\partial p} \left( \frac{\langle v \theta \rangle - \langle v \rangle \langle \theta \rangle}{\langle \theta \rangle_p} \right) \) (Andrews et al. 1987). The Brewer-Dobson circulation, characterized by upwelling motion in the tropics and poleward and downwelling motion in the extratropics, is regarded as a Lagrangian mean circulation and is usually approximated by the residual circulation of the TEM equations (Dunkerton 1978). In the NH winter stratosphere, the climatological EP flux divergence is negative and implies a westward forcing on the zonal mean flow due to the breaking and dissipation of vertically propagating planetary waves and is primarily responsible for the deceleration of the polar night jets in the stratosphere. This deceleration tendency is mainly balanced by the Coriolis torque associated with the residual circulation.
b. **Spectral and Cross-spectral Analysis**

In order to identify the dominant waves during the transient adjustment process, the EP flux is decomposed into different zonal wave numbers as the following:

\[ u(\lambda) - \langle u \rangle \xrightarrow{\text{FT}} \sum_{k=0}^{k_{\text{max}}} \hat{u}(k) \]  
\[ (u - \langle u \rangle)(v - \langle v \rangle) \xrightarrow{\text{FT}} 2\Re \sum_{k=0}^{k_{\text{max}}} \hat{u}(k) \cdot \hat{v}^*(k) \]  

where \( \lambda \) is longitude and \( k \) is zonal wave number. \( \hat{u}(k) \) is the Fourier transform (FT) of the zonal (meridional) eddy velocity and \( \hat{v}^*(k) \) denotes the complex conjugate of \( \hat{v}(k) \). The same methodology applies to the meridional heat flux.

In addition, following Randel and Held (1991), a phase speed spectrum for eddy momentum flux convergence is computed. We first compute the zonal wave number (\( k \))-frequency (\( \nu \)) co-spectra of \( (u, v) \) using the daily data of \( u \) and \( v \). The wave number (\( k \))-angular phase speed (\( C_{p,a} \)) co-spectra is then defined and transformed from the \( k-\nu \) co-spectra by conserving the total power of momentum flux convergence, where \( C_{p,a} \) here is defined as \( C_{p,a} = \frac{\nu \cos \phi}{k} \).

Finally the phase speed spectrum of momentum flux convergence is constructed by summing over all the zonal wave numbers and is plotted as a function of latitude and angular phase speed.

c. **Linear Quasi-geostrophic (QG) Refractive Index**

The linear QG refractive index is a useful predictor and diagnostic for the propagation of planetary waves and has been widely used in various climate states to help interpret the behavior of waves and their interaction with the mean flow (e.g., Charney and Drazin 1961; Matsuno 1970; Butchart et al. 1982; Chen and Robinson 1992; Lorenz and Hartmann 2001; Harnik and Lindzen 2001; Seager et al. 2003; Simpson et al. 2009; Harnik et al. 2010; Sigmond and Scinocca 2010; Shaw et al. 2010).
Karoly and Hoskins (1982) demonstrated that under linear Wentzel-Kramers-Brillouin (WKB) theory waves are refracted by the gradients of linear refractive index and thus tend to propagate from regions of low refractive index to regions of high refractive index. Although WKB theory requires the wavelength of the perturbations to be smaller than the scale of the basic state, which may not be well satisfied in the stratosphere, interpretations of wave propagation in the stratosphere using the index of refraction have been successful (e.g., Matsuno 1970; Harnik and Lindzen 2001; Perlwitz and Harnik 2003; Shaw et al. 2010).

Matsuno (1970) provided an analytical formula for the stationary QG linear refractive index by assuming the atmosphere isothermal (thus constant buoyancy frequency, $N$). This assumption is a reasonable approximation for the stratosphere but may not for the troposphere. The zonal mean QG refractive index for both stationary and transient eddies with phase speed $C_p$ is written as:

$$n_{\text{ref}}^2 = \frac{a\langle q_\phi \rangle}{\langle u \rangle - C_p} - \frac{k^2}{\cos^2 \phi} - \frac{f^2 a^2}{4N^2 H_o^2} - \frac{k^2}{\cos^2 \phi} - \frac{f^2 a^2}{4N^2 H_o^2}$$  \hspace{1cm} (6)

$$\langle q_\phi \rangle = 2\Omega \cos \phi - \frac{\partial}{\partial \phi} \left[ \frac{1}{\cos \phi} \frac{\partial \langle u \rangle \cos \phi}{\partial \phi} \right] + \frac{f^2 a}{R} \frac{\partial}{\partial p} \left( \frac{p}{T} \langle \theta \rangle \langle u \rangle \right)$$  \hspace{1cm} (7)

where $R_d$ is the dry air gas constant (287 J/kg/K), $q_\phi$ is the meridional potential vorticity (PV) gradient and $H_o$ is the scale height of pressure ($H_o = 7$ km). The $n_{\text{ref}}^2$ is dimensionless in Equation (6) which is the same as that in Simpson et al. (2009).

Furthermore, we also make use of a linear quasi-geostrophic (QG) model to diagnose the wave propagation characteristics of a two-dimensional zonal mean basic state (Harnik and Lindzen 2001). The model is basically a QGPV conservation equation and takes the zonal mean of $u$ and $T^2$ as inputs with a specified eddy phase speed ($C_p$) and zonal wave number ($k$). The model then numerically calculates the steady-state eddy fluxes purely using the solution to the QGPV model. The model also diagnostically separates the index of refraction ($n_{\text{ref}}^2$) into the vertical ($m_z^2$) and meridional wave numbers ($l_y^2$) which serve as more accurate indicators of wave propagation in the vertical and meridional directions (e.g., Harnik and

\[2\text{The linear QG model doesn’t assume an isothermal atmosphere.}\]
3. Climatological CAM3-SOM Simulations

This section compares the climatological CAM3-SOM simulations in NH winter with the latest Reanalysis data from the European Centre for Medium-Range Weather Forecasts (ECMWF) which is generally regarded to have a good representation of the stratosphere (e.g., Seviour et al. (2011)). The ECMWF Interim Reanalysis (ERA-Interim; Dee and co-authors (2011)) data are available over the period 1979-2010 and have 37 pressure levels up to 1mb. The circulation in the NH winter stratosphere is largely modulated by upward propagating planetary waves generated in the troposphere by the orographic forcing and large-scale zonally asymmetric diabatic heating. Figure 3(a) shows the climatological EP flux and convergence calculated from the ERA-Interim daily variables for February\(^3\). The EP flux vectors in this study are all normalized by the basic state density as in Edmon et al. (1980) to better display the wave activity in the stratosphere. The flux vectors clearly indicate that waves are generated in the lower troposphere and propagate upward into the stratosphere. These upward propagating Rossby waves in the extratropics are always refracted equatorward toward the critical layer where the eddy phase speed\(^4\) equals the zonal mean flow velocity and the waves are absorbed.

The maxima in climatological EP flux convergence, e.g. \(\frac{1}{\cos\phi} \nabla \cdot \vec{F} < 0\), in general occur in regions where waves are absorbed or dissipated, for example, north of the subtropical critical layer, in the high latitude middle troposphere and in high latitudes below the polar

\(^3\)February is chosen here for a better comparison with the anomalies in February from the model simulations to be shown later. The climatological features in February are generally similar to Dec-Jan-Feb averages.

\(^4\)The dominant waves in the stratosphere are planetary stationary waves with zero phase speed while those in the troposphere are transient eddies with typical phase speed of 8 m/s. Contours of \(C_p = 0\) m/s and 8 m/s are highlighted in Fig. 3(e) to be discussed later.
jet (shown in Fig. 3(a)). In particular, the net convergence of EP flux in the high latitude stratosphere is consistent with the Brewer-Dobson circulation with upwelling in the tropics and poleward and downward motion at high latitudes. However, as shown in Fig. 3(b), the CAM3-SOM fails to capture the convergence in the high latitude stratosphere correctly and instead produces a net divergence of EP flux (i.e., $\frac{1}{\cos \phi} \nabla \cdot \vec{F} > 0$). This is balanced by the Coriolis torque and imposes an unrealistic westerly acceleration tendency in the polar stratosphere as a result of the amplification of the momentum flux and the under-estimate of the heat flux near the model lid of CAM3 (not shown). This is consistent with the difference in the stratospheric polar jet which is stronger in CAM3-SOM than in the ERA-Interim Reanalysis (shown in Fig. 3(a)(b)). Upper boundary conditions are commonly applied in general circulation models and the upper model lid in CAM3 is at about 2.9mb where the vertical velocity is assumed zero. The effects of this artificial upper boundary on climate model simulations in both the stratosphere and the troposphere have been long recognized (e.g., Boville 1984; Boville and Cheng 1988; Shaw and Perlwitz 2010; Sassi et al. 2010). The upper model lid leads to reflection of vertically propagating wave activity, changes the meridional/vertical phase structures, causes increased (decreased) poleward eddy momentum (heat) flux, and results in a net westerly forcing on the zonal mean flow (Boville and Cheng 1988; Sassi et al. 2010). Although Rayleigh friction is usually applied at the model top to damp vertically propagating planetary waves, it does not prevent the reflection of wave activity. Sassi et al. (2010) compared the present-day simulations between CAM3 and the Whole Atmosphere Community Climate Model version 3 (WACCM3) (its vertical domain extends to $5.9 \times 10^{-6}$mb) and found substantial differences in the zonal mean state of the stratosphere and the behavior of the stratospheric variability, such as the life cycle of weak stratospheric vortex events, and their influence on the tropospheric circulation. In particular, they demonstrated that the amplification of the momentum flux and the reduction of the heat flux near the model lid of CAM3 coincided with the region of wave reflection in the high latitude stratosphere (see Figure 4 and 5 in Sassi et al. 2010). This suggests caution
in interpreting the circulation behavior in CAM3-SOM as a consequence of CO₂ doubling and its relevance to the real world. However, with this in mind, our goal here is to explain the response in CAM3 because it represents a state-of-the-art IPCC AR4 model that has predicted a poleward shift of the tropospheric jet in response to increased CO₂.

Figure 3(c)(d) shows the contributions from planetary waves including wave-1 and wave-2 from the ERA-Interim Reanalysis dataset and the CAM3-SOM simulations, respectively. The similarities in the stratosphere between Fig. 3(a) and (c) (also (b) and (d)) indicate the dominance of the planetary-scale long waves \((k = 1 \text{ and } 2)\) in the stratosphere, which is consistent with the theoretical work of Charney and Drazin (1961).

Figure 3(e)(f) shows the calculated index of refraction \(n_{\text{ref}}^2\) for planetary-scale stationary waves in the stratosphere with \(C_p = 0\). The \(n_{\text{ref}}^2\) in Equation (6) is dominated by the contribution from \(\frac{a(q\phi)}{(u-C_p)}\), while the term carrying the zonal wave number \(-\frac{k^2}{\cos^2\phi}\) is much smaller than other terms for planetary-scale waves and is thus neglected. The index of refraction is positive almost everywhere in the extratropics except for the region of minimum values in the midlatitude lower stratosphere at about 40°N between 70mb and 100mb (shown in Fig. 3(e)(f)). Matsuno (1970) noted the significance of this minimum in refractive index and argued that it creates a partial wave guide for vertical propagation on its poleward side. In addition, the \(n_{\text{ref}}^2\) increases almost monotonically from high to low latitudes and becomes infinitely large as the waves reach the zero wind line (critical layer) which is highlighted in blue in Fig. 3(e)(f). As can be seen, the propagation of waves, indicated by the EP flux vectors, generally follows the gradients of \(n_{\text{ref}}^2\), and the waves are indeed refracted equatorward toward increasing \(n_{\text{ref}}^2\). The linkage between wave propagation and the index of refraction appears to be qualitatively robust for the climatological basic state in both the stratosphere and the troposphere.
4. Three-Phase Atmospheric Transient Circulation Adjustment Process

The three phases of the circulation response occur in January-February-March-April (JFMA) in year 1. At the end of this transient adjustment process, the extratropical circulation response resembles that in the quasi-equilibrium state and the tropospheric jet streams are shifted poleward. Figure 4 shows the latitude-pressure level plot of the $\langle T \rangle$ and $\langle u \rangle$ anomalies in Phase 1 (January), 2 (February) and 3 (March and April), respectively. The climatological $\langle T \rangle$ and $\langle u \rangle$ on January 1st are also shown in Fig. 4 for reference. In the following we discuss the dynamical mechanisms involved in each of the three phases.

a. Phase One (January): Stratospheric Subpolar Easterly Anomaly

An easterly anomaly in the high latitude stratosphere, together with a westerly anomaly in low latitudes occur in the first few days after the instantaneous doubling of CO$_2$ on January 1st. This is a fast purely radiatively-driven response. The stratosphere cools with increased CO$_2$ and emits increased longwave radiation to space. In general, with a uniform CO$_2$ increase, the stratosphere cools due to black body radiation more (less) where the control basic state temperature is warmer (colder). Thus the cold anomaly in the stratosphere increases with height and also varies with latitude. As shown in Fig. 4(a), the NH basic state temperature in the stratosphere increases with latitude but only to the midlatitudes and then decreases towards the North Pole where there is no incoming solar radiation. The radiative response basically follows the basic state temperature structure in the stratosphere, with more longwave radiation emitted out to space in the northern middle latitudes, causing maximum cooling there and generating a poleward (equatorward) flow and a westerly (easterly) anomaly in the low (high) latitudes due to geostrophic adjustment (shown in Fig. 4(c)(d)). In Phase 1 the circulation response is primarily in the stratosphere.

The radiative response in the SH is different from that in the NH because of the dif-
ference in basic state temperature. Here the stratospheric basic state temperature has a minimum at the Equator and monotonically increases towards the South Pole as a consequence of absorption of incoming solar radiation by the ozone during summer. This reversed temperature gradient in the stratosphere is consistent with the climatological easterlies in the southern summer (shown in Fig. 4(b)). After the CO$_2$ is increased, the SH stratosphere cools most at the pole and least at the Equator, reducing the meridional temperature gradient causing a poleward flow and a westerly zonal wind anomaly. Because of the weak planetary wave forcing in the southern summer, the stratospheric circulation response is primarily controlled by the radiative forcing until the zonal wind anomaly penetrates into the upper troposphere/lower stratosphere where the transient eddies are expected to respond and impact the whole troposphere. In contrast, the planetary wave activity in the NH and its upward propagation greatly modulates the stratospheric circulation as to be discussed below.

b. Phase Two (February): Stratospheric Westerly Acceleration

In February there is a westerly anomaly in the NH stratosphere which is consistent with further cooling in the subpolar stratosphere (shown in Fig. 4(e)(f)). A diagnosis of the EP flux and the TEM zonal momentum equation indicates that the westerly acceleration in the stratosphere is mainly eddy-driven. Figure 5(a) shows the EP flux anomaly in Phase 2 and the combined contribution from planetary wave-1 and wave-2 is shown in Fig. 5(b). The agreement between Fig. 5(a) and Fig. 5(b) indicates that the dominant waves controlling the anomalies in Phase 2 are of planetary scale too. Figure 5(c) shows the EP flux anomalies from wave-1 and its associated horizontal divergence. It is found that the westerly acceleration in the stratosphere in Phase 2 is primarily caused by the increased momentum flux convergence from planetary wave 1 as a result of increased equatorward wave propagation (as shown in Fig. 5(c)).

To understand why more planetary waves are refracted equatorward as a consequence
of the CO$_2$ increase, the index of refraction $n^2_{\text{ref}}$ is computed, the anomaly of which is shown in Fig. 5(d) for $C_p = 0$ m/s superimposed with the corresponding total EP flux anomalies during Phase 2. The largest equatorward wave refraction occurs in the midlatitude stratosphere between about 50°N and 60°N whereas the change in $n^2_{\text{ref}}$ is positive (negative) on the poleward (equatorward) side. This is not in agreement with predictions of the linear refraction theory even though this theory explains the climatological state well. The reason why stationary eddies refract more equatorward remains unclear. However, as mentioned in the previous section, the low model top is likely to affect the wave propagation in the climatological state. It is possible that, in response to CO$_2$ increase, the wave propagation anomalies, in particular, near the model top, are also influenced by the model upper boundary, and if this is the case, the wave propagation may not follow the index of refraction since it does not account for an upper boundary condition associated with a low model lid. Increased equatorward wave propagation in response to climate change is also found in one of the equilibrium responses to doubling of CO$_2$ in Sigmond et al. (2008) and Sigmond and Scinocca (2010) (see Figure 6(a)(d)(g) in Sigmond and Scinocca 2010). They attributed the enhanced equatorward propagation to the disappearance of the negative $n^2_{\text{ref}}$ region in the subtropical lower stratosphere. The region of negative $n^2_{\text{ref}}$, however, does not disappear in our simulations.

c. Phase Three (March and April): Poleward Displacement of Tropospheric Jets

As shown in Figure 8(b) in Part I (Wu et al. 2011), the tropospheric jet stream shifts poleward in early March. This circulation change in the troposphere appears to follow the westerly acceleration in the stratosphere and the intensification of transient eddy momentum flux convergence in the subtropical middle and upper troposphere (shown in Figs. 7(b) and 8(d) in Part I). This is consistent with the conclusion in Kushner and Polvani (2004) and Song and Robinson (2004) which demonstrated that a westerly acceleration in the stratosphere and
a corresponding weakening of the residual circulation can induce the zonal wind anomaly in
the mid-troposphere and a poleward jet shift that the transient eddies are crucial in shifting
the tropospheric jet streams.

Previous studies have proposed hypotheses to explain the tropospheric jet shift in re-
response to global warming. The hypotheses include: (1) an increase in eddy phase speed (Lu
et al. 2008; Chen et al. 2008); (2) a rise in tropopause height (Lorenz and DeWeaver 2007);
(3) an increase in eddy length scale (Kidston et al. 2011). Another possibility comes from
the idea of changing index of refraction which has been used to understand the transient
eddy propagation during El Niños (Seager et al. 2003; Harnik et al. 2010) as well as solar
cycles (Simpson et al. 2009). We analyze each of the above possible mechanisms and see
whether or not they can explain the jet shift that occurs during Phase 3. This study provides
a unified assessment of all the possible mechanisms within the same framework.

1) Eddy Phase Speed

Lu et al. (2008) and Chen et al. (2008) identified an increase in eddy phase speed in
the GFDL CM2.1 model simulations under the 'business-as-usual' A2 scenario in which the
CO₂ concentration reaches 800 ppmv at the end of the 21st century. They argued that this
eddy phase speed increase causes the critical line, subtropical breaking region, transient eddy
momentum flux convergence and tropospheric zonal jets to move poleward. Here we follow
the computational methodology for the eddy phase speed cross-spectra as in Randel and
Held (1991) and Chen and Held (2007) and see whether this hypothesis helps explain the
shift in the jet position in our modeling experiments.

Figure 6 shows the co-spectra of eddy momentum flux convergence at 250mb\(^5\) during
the first 120 days of transient adjustment in January-February-March-April of year 1 as
a function of angular phase speed \(C_{p,a}\) and latitude along with the 250mb zonal wind

\(^5\)250mb is chosen to be consistent with Lu et al. (2008). The conclusion doesn’t change when other
pressure levels are used.
distribution. The difference between the 2CO$_2$ and the 1CO$_2$ runs is shown in contours while the climatology is shown in color shadings. As expected the climatological co-spectrum shows a divergence in eddy momentum flux in the subtropics and a convergence in the midlatitudes. The 250mb waves are primarily eastward propagating transient waves with an angular phase speed of about 10m/s in the NH. In addition, the meridional wave propagation is confined by the subtropical critical layer, consistent with linear wave refraction theory. The transient anomalies in eddy momentum flux convergence co-spectra show a poleward shift in the NH relative to the climatology with an intensification (reduction) on the poleward (equatorward) flank of the climatological maximum position. However, our model experiments with CAM3-SOM don’t show any significant increase in eddy phase speed during the transient adjustment process. The change in zonal mean zonal wind at 250mb is also small (shown in Fig. 6). We find no significant increase in eddy phase speed in the quasi-equilibrium state either (not shown).

Simpson et al. (2009) also found no increase in eddy phase speed in their study investigating the transient response of tropospheric circulation anomalies to stratospheric heating perturbations in a simple general circulation model. Rivire (2011) investigated the effect of changing eddy length scale on wave breaking and resulting changes in jet position, and a slightly decreased phase speed was found during the process. Perhaps increased eddy phase speed in response to an increase in CO$_2$ is model-dependent, but at least in this model it is not an explanation of the tropospheric jet shift.

2) Tropopause Height

As the location of the jet streams and the scale of eddies are closely related to the depth of the troposphere, it is possible that the jet shift could be induced by a change in tropopause height. Since the scale of the eddies is characterized by the Rossby radius, $L_R = \frac{NH}{f}$, where $H$ denotes the thickness of the troposphere, eddies in theory should become larger as the tropopause height is raised. As the baroclinic zone shifts poleward as a consequence of higher
tropopause, the jets are expected to move toward higher latitudes as well. Observations indicate that the height of the tropopause has gone up by several hundred meters since 1979, which is predominantly driven by anthropogenic forcing (Santer et al. 2003). This rising trend in tropopause height has also been found to be closely related to the warming of the troposphere and the poleward expansion of the Hadley Cell (e.g., Santer et al. 2003; Lu et al. 2009). Lorenz and DeWeaver (2007) found similarities in the extratropical circulation response between IPCC AR4 coupled models (A2 scenario) and a simple dry GCM when the tropopause height is raised (by about 400 meters), suggesting that the rise in tropopause height is the dominant driver of the extratropical circulation response to global warming although in their experiments the effect of increasing baroclinic instability in the upper troposphere wasn’t excluded.

Figure 7 shows the day-by-day evolution of the 850mb midlatitude jet maximum location and the rise in tropopause height averaged in the midlatitude region between 30°N and 70°N. The calculation of tropopause height follows the algorithm in Reichler et al. (2003) and finds the lowest pressure level at which the temperature lapse rate decreases to 2 K/km. Because of the coarse resolution in latitude of CAM3, the zonal mean zonal wind is first interpolated to a finer latitude grid before locating the jet maximum. As a result of CO$_2$ doubling and the fast radiative stratospheric cooling, the tropopause starts to rise after about 15 days and keeps on rising by about 1mb before it drops at the end of February. Despite these changes in tropopause height, there is not much change in jet maximum position near the surface. In early March, the tropopause height starts to rise again sharply and this time is followed by a poleward shift in the jet position near the surface with a lag of a few days. This is a robust result for the jets at various vertical levels and for tropopause height at different latitudes. The tropopause, on average, rises by 2 mb while the low-level jet moves by about 1°N in March. Although the rise in tropopause height leads the low-level jet shift in early March, there is overall little correlation between the time history of these two quantities.

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Cubic spline interpolation is used here but the results don’t change much for other interpolation schemes.
argued in Kidston et al. (2011), the dynamics could connect the tropopause height rise with
the poleward displacement of the tropospheric jet streams via an increase in eddy length
scale, the day-by-day evolution of which is further investigated below.

3) Eddy Length Scale

Kidston et al. (2010) found a robust increase in eddy length scale in the A2 scenario
simulation of the future climate among an ensemble of CMIP3/IPCC AR4 models. Wu
et al. (2010) also noticed this increase in eddy length scale in the GFDL CM2.1 model under
the A1B scenario and found that the larger eddies are partially responsible for the increased
poleward energy transport carried by the storm tracks in the future climate. Kidston et al.
(2011) refined the idea that an increase in eddy length scale can cause the jet streams to
move poleward rather than vice versa, which is confirmed in a simple barotropic model
experiment. Rivire (2011) emphasized the role of enhanced upper-tropospheric baroclinic
instability in the poleward shift of the jet streams via changes in eddy length scale and
anticyclonic/cyclonic wave breaking in the global warming scenario.

Following the methodology in Kidston et al. (2010), we calculate the eddy length scale
day by day during the transient adjustment process. The mean eddy length scale is defined
as
\[ \bar{L}_{\text{eddy}} = \frac{2\pi\cos\phi}{{\bar{k}}} \]
with \( {\bar{k}} \) measuring the energy weighted zonal wave number
\[ {\bar{k}} = \frac{\sum_{k} k |\hat{v}(k)|^2}{\sum_{k} |\hat{v}(k)|^2} \]
where \( |\hat{v}(k)|^2 \) denotes the high-pass filtered\(^7\) meridional component of the eddy kinetic energy
in wave number \( (k) \) space. Figure 8 plots the day-by-day evolution of the change in eddy
length scale \( \bar{L}_{\text{eddy}} \) averaged in the midlatitudes between 30°N and 70°N at 500mb along
with the jet shift at 500mb. The evolution in eddy length scale is noisy compared with
that of the jet shift and tropopause height rise but clearly shows a rapid transition from
a negative anomaly to a positive anomaly starting in early March. The increase in eddy
length scale on average reaches about 60km at the end of March and in April. The time
sequence in Fig. 8 indicates that the increase in eddy length scale takes place after the jet

\(^7\)The high-pass filter retains synoptic time scales of 2-8 days.
shift in the middle troposphere, with a time lag of about two days. This time sequence is robust for the jet shift and the eddy length scale increase at different vertical levels, which implies that during the transient adjustment process the eddy length scale increase is a consequence of the tropospheric jet shift. This is supported by Barnes and Hartmann (2011) which demonstrated in a barotropic model that as the eddy-driven jet is located at higher latitudes, the eddy length scale increases as suggested by the linear Rossby theory.

The time sequence between the jet shift and the increase in eddy length scale in the extratropical troposphere supports the idea that the eddy length scale increase occurs as a consequence of the poleward jet shift in our modeling experiments rather than vice versa as suggested by Kidston et al. (2011). In addition, as the increase in eddy length scale follows the jet shift, the dynamical mechanisms linking the tropopause height rise and the jet shift remain unclear.

4) Linear Refractive Index

The index of refraction, in particular the meridional wave number calculated from the linear QG model, was used to understand the equatorward displacement of the transient eddies, and the associated dynamical mechanisms, during El Niños (Seager et al. 2003; Harnik et al. 2010). They showed that, during El Niños as a consequence of the equatorward shift of the subtropical jets and resulting changes in meridional wave number, the transient eddies act to persistently maintain the mean flow anomalies via anomalous convergence (divergence) of momentum flux in the subtropical (midlatitude) region. Simpson et al. (2009) also successfully used the index of refraction to diagnose and interpret how changing eddy propagation and eddy momentum fluxes drive anomalous tropospheric circulation as a result of initial stratospheric heating perturbations applied in a simple GCM.

In this section, we use the linear QG model from Harnik and Lindzen (2001) to isolate the effect of linear wave refraction and to quantify the transient eddy feedback to the tropospheric
A zonal wave number of 6 and a phase speed of about 10 m/s is prescribed\textsuperscript{8}. The daily zonal mean zonal wind and temperature fields from each day of the 120-day adjustment process are used as input for the linear QG model, and the corresponding daily eddy fluxes are calculated. Figure 9 shows the day-by-day evolution of the anomalies in 150mb zonal wind from the CAM3-SOM experiments, 150mb horizontal eddy momentum flux convergence (HEMFC) calculated from the linear QG model, for comparison, and the 150mb high-pass filtered HEMFC and 500mb zonal wind from the CAM3-SOM experiments. A 5-day temporal running average has been applied to these variables\textsuperscript{9}. The intensified convergence of transient eddy momentum flux at 150mb occurs roughly on March 5th after the westerly acceleration in the extratropical lower stratosphere at 150mb (shown in Fig. 9(c) and Fig. 9(a)). Both the HEMFC and zonal wind changes at 150mb take place before the poleward displacement of the tropospheric jet streams, for example, at 500mb, which occurs roughly on March 9th (shown in Fig. 9(d)). The agreement of the response in HEMFC between the linear QG model and the CAM3-SOM experiments, especially the well-organized dipole structure starting from early March as shown in Fig. 9(b)(c), suggests the dominant mechanism of linear refraction as the cause of the poleward shift of the tropospheric midlatitude jet streams.

To further demonstrate how the zonal wind response migrates downward from the lower stratosphere to the troposphere as a result of a changing index of refraction, we focus on two time intervals: (1) before the 500mb jet shift from March 4th to 8th (highlighted in dashed lines in Fig. 9); (2) after the 500mb jet shift from March 9th to 13th (highlighted in solid lines in Fig. 9). Figure 10(a) shows the zonal wind anomaly before the 500mb jet shift when the westerly intensification is primarily located in the stratosphere. As a result of the subpolar lower stratospheric westerly anomaly, the transient eddies in the troposphere

\textsuperscript{8}We have used the same set of model parameters as in Seager et al. (2003) and Harnik et al. (2010). We have also assumed \( k \) and \( C_p \) constant for the 1CO\textsubscript{2} and the 2CO\textsubscript{2} runs. This mechanism of linear wave refraction is different from section 1) and 3) and assumes no changes in eddy properties.

\textsuperscript{9}The results are not sensitive to the choice of running averages.
of the linear QG model respond by refracting equatorward, roughly from poleward of 50°N to equatorward of 50°N, from negative to positive changes in meridional wave number \( (l_y^2) \), as shown in Fig. 10(b). The changes in \( l_y^2 \) in general follow that in \(-\frac{\partial}{\partial \phi} \left[ \frac{1}{\cos \phi} \frac{\partial (u \cos \phi)}{\partial \phi} \right] \) in the troposphere (not shown), the 2nd term in the meridional PV gradient as in Equation (7). This anomalous propagation of transient eddies implies a westerly acceleration (deceleration) tendency poleward (equatorward) of 50°N in the troposphere right below the lower stratospheric wind anomaly which would shift the climatological tropospheric jet streams poleward. Figure 10(c) shows the 'implied' zonal wind anomaly by adding the HEMFC due to changes in refractive index as calculated by the linear QG model. As shown in Fig. 10(c)(d), there is a general agreement between the 'implied' zonal flow anomaly based on the linear QG model and the actual zonal jet anomaly in the GCM, in particular, the poleward shift of the midlatitude jet streams in the troposphere. This suggests that the linear refraction theory is able to capture the poleward jet shift.

However, the linear refraction theory cannot capture the response of the subtropical jets in the upper troposphere. As shown in Fig. 4(f)(h), the CAM3-SOM tends to weaken the subtropical jet from Phase 2 to Phase 3. On the contrary, according to the results of the linear QG model, the subtropical jets should strengthen (Fig. 10(a)(c)). Hence, for the subtropical jets, mechanisms other than linear refraction theory might be important and we leave this part for future studies.

Returning to the midlatitude jet, as shown in Fig. 9(b)(c), after the tropospheric jet has shifted, transient eddies act to feed back positively onto the zonal wind by accelerating the zonal flow on the poleward flank between 40°N and 60°N while decelerating on the equatorward side. Therefore, it is the wave-mean flow interaction in the lower stratosphere that initiates the poleward movement of the tropospheric jet streams and the positive feedback between the zonal flow and the transient eddies acts to maintain the jet position change.

In the diagnosis of Phase 2, it has been found that the stratospheric westerly acceleration is a consequence of increased equatorward refraction of stationary waves, which, however,
can’t be interpreted from the theory of linear refraction. As discussed previously, the reason is probably related to the existence of the low model top that ‘artificially’ alters the wave propagation. In contrast, in this section, we focus on the dynamics of transient wave-6 (far away from being influenced by the model upper boundary) and the propagation and eddy fluxes are largely determined by the index of refraction, consistent with prior work explaining transient eddy response to perturbations in terms of wave refraction (e.g., Seager et al. 2003; Simpson et al. 2009; Harnik et al. 2010).

5. Conclusion and Discussions

The daily evolution of the atmospheric circulation adjustment to an instantaneous and uniform doubling of CO$_2$ has been investigated. As a consequence of instantaneous and uniform doubling of CO$_2$, the daily evolution of the atmospheric circulation reveals the chain of causality that occurs in the adjustment. This was examined by analyzing the day-to-day and week-to-week forced change over an ensemble of 100 runs with slightly different initial conditions at the time of instantaneous CO$_2$ doubling. It is found that after a few months of integration, the circulation and thermal responses in the extratropical troposphere resemble the major features seen in the quasi-equilibrium simulations from the CMIP3/IPCC AR4 coupled models using the A1B emission scenario, which suggests the usefulness and relevance of examining the transient adjustment process. Part I of the paper mainly focused on the transient thermal response in the troposphere. It showed that the extensive warming in the upper and middle subtropical troposphere is caused adiabatically by the anomalous descending motion driven by the transient eddy momentum flux anomalies in the troposphere. Here, Part II explores the dynamical mechanisms underlying the sequential transient adjustment leading up to the establishment of the circulation response in the extratropical troposphere. From the day-by-day evolution of the zonal mean zonal wind in the extratropics, three phases are defined. The initial response takes place in the stratosphere and
involves a westerly flow anomaly in the low latitude stratosphere, together with an easterly anomaly in the northern high latitudes, both driven radiatively by the CO$_2$ increase and associated latitudinal gradients of the temperature response. The easterly anomaly in the northern high latitude stratosphere switches to a westerly flow acceleration throughout the stratosphere in Phase 2 driven by enhanced planetary-scale eddy horizontal momentum flux divergence. The index of refraction could not explain the eddy response in the stratosphere during Phase 2. This may be related to the low model upper boundary and possible wave reflection. Phase 3 involves the downward migration of the westerly acceleration from the lower stratosphere into the troposphere, followed by the poleward shift of the tropospheric jet streams. Previous studies have provided possible mechanisms to interpret this process and they are all examined here to see whether they cause the tropospheric jet shift in these modeling experiments: 1) Different from Lu et al. (2008) and Chen et al. (2008), we found no significant increase in eddy phase speed in either the transient evolution or the quasi-equilibrium state. 2) The day-by-day evolution of the rise in tropopause height (averaged in the extratropics) appears to lead the tropospheric jet shift by a few days but there is not a close relation of the time evolution of these two quantities and the dynamical mechanism linking the two is not clear yet. 3) The transient eddy length scale increases but only after the tropospheric jet has shifted, different from the sequence proposed by Kidston et al. (2011). 4) The transient eddies play an important role in causing and maintaining the poleward displacement of the tropospheric jet streams via an acceleration (deceleration) on the poleward (equatorward) flank and the anomalies in eddy propagation and momentum flux are attributed to the changes in the basic state and resulting changes in the index of refraction.

There are a few caveats in this study and future work is needed to achieve a better and thorough understanding of the dynamics. First is the dependence of the transient response on initial conditions. The radiative response strongly depends on the latitudinal distribution of the basic state temperature which controls the temperature anomaly following the CO$_2$
doubling. Hence experiments starting from January 1st and others starting from July 1st are expected to behave differently. This suggests the necessity of another set of modeling experiments with the initial condition of July 1st. Second, although the initial radiative response in the NH wintertime is an equatorward shift of the polar jets with an easterly (westerly) anomaly at high (low) latitudes in the stratosphere, this feature is later strongly modified by the dynamical forcing imposed by the planetary waves. This response is not well understood and could possibly be caused by the existence of the low model upper lid and resulting downward wave reflection, more planetary-scale long waves are refracted equatorward, leading to a westerly flow acceleration in the stratosphere. Although this consequence of a low model upper boundary might be universal among other CMIP3/IPCC AR4 coupled climate models, it leads to another question of what the stratospheric and tropospheric circulation responses would be in a stratosphere resolving model. For example, Scaife et al. (2011) demonstrated the circulation responses to climate change for standard (CMIP3/IPCC AR4) and stratosphere resolving climate models (models for the Chemistry Climate Model Validation (CCMVal) project). It should be noted that the difference in basic state between the two sets of models probably is not excluded as a cause of differences in the responses (Sigmond and Scinocca 2010). Finally, in our model simulations, the stratospheric circulation responses in the two hemispheres are different. For example, the stratospheric polar jet strengthens in northern winter while that in southern winter shifts equatorward associated with an easterly (westerly) anomaly at high (low) latitudes (as shown in Figure 3(c)(d) in Part I). This feature is true for both the transient and quasi-equilibrium state in our model simulations. It is possible that the differences in both the radiative and dynamical adjustment for the two hemispheres cause the different circulation anomalies.

While the current work represents an advance in our understanding of the poleward shift of the jet stream and storm tracks in response to global warming, clearly more work is needed, especially with models with well resolved stratosphere, to fully understand this important aspect of climate change.
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