The Transient Circulation Response to Radiative Forcings and Sea Surface Warming*

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(Manuscript received 19 November 2013, in final form 27 August 2014)

ABSTRACT

Tropospheric circulation shifts have strong potential to impact surface climate. However, the magnitude of these shifts in a changing climate and the attending regional hydrological changes are difficult to project. Part of this difficulty arises from the lack of understanding of the physical mechanisms behind the circulation shifts themselves. To better delineate circulation shifts and their respective causes the circulation response is decomposed into 1) the "direct" response to radiative forcings themselves and 2) the "indirect" response to changing sea surface temperatures. Using ensembles of 90-day climate model simulations with immediate switch-on forcings, including perturbed greenhouse gas concentrations, stratospheric ozone concentrations, and sea surface temperatures, this paper documents the direct and indirect transient responses of the zonal-mean general circulation, and investigates the roles of previously proposed mechanisms in shifting the midlatitude jet. It is found that both the direct and indirect wind responses often begin in the lower stratosphere. Changes in midlatitude eddies are ubiquitous and synchronous with the midlatitude zonal wind response. Shifts in the critical latitude of wave absorption on either flank of the jet are not indiced as primary factors for the poleward-shifting jet, although some evidence for increasing equatorward wave reflection over the Southern Hemisphere in response to sea surface warming is seen. Mechanisms for the Northern Hemisphere jet shift are less clear.

1. Introduction

The global climate response to a given radiative forcing is often partitioned into 1) the "direct" response of the Earth climate system to the radiative forcing itself before sea ice and sea surface temperatures (SSTs) appreciably respond and 2) the "indirect" response of the Earth climate system to the SST anomalies that result from the applied forcing. This sort of decomposition is not only useful in estimating the strength of climate feedbacks (Hansen et al. 2005; Andrews et al. 2012), but it also has been used to study the response of the general circulation to specific forcings (Deser and Phillips 2009;

DOI: 10.1175/JCLI-D-14-00035.1

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Polvani et al. 2011a,b; Staten et al. 2012). These latter studies describe a poleward shift in the Southern Hemisphere (SH) midlatitude jet and a widening of the tropical Hadley circulation as greenhouse gas concentrations increase, stratospheric ozone depletes, and SSTs warm.

Staten et al. (2012) examine the equilibrium direct and indirect response of the general circulation to radiative forcings. The general circulation is shown to be sensitive to ozone depletion and recovery over the SH during austral spring and summer. However, the indirect response to global climate change—that is, the atmospheric response to sea surface warming—is shown to have a large impact year-round. Changes in the mean flow and eddies in each case resemble those expected from the work of Chen and Held (2007). However, by construction, the methods of Staten et al. (2012) do not distinguish between cause and effect.

To better understand the order of events in the circulation response, several studies investigate the transient response to switch-on forcings or perturbations in

^{*} Supplemental information related to this paper is available at the Journals Online website: http://dx.doi.org/10.1175/JCLI-D-14-00035.s1.

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idealized simulations (Kushner and Polvani 2004; Simpson et al. 2009, 2010, 2012; Wittman et al. 2004). These studies emphasize the importance of eddies in altering the mean state and variability of the zonal mean circulation. However, these are highly idealized studies, and they do not directly address the realistic circulation response to anthropogenic climate forcings.

Recently, studies have begun to address the ensemblemean transient circulation response to switch-on climate forcings in complex general circulation models. For example, Wu et al. (2012) document the transient circulation response to a doubling of CO_2 , in the Community Atmosphere Model, version 3 (CAM3), coupled to a slab ocean. They suggest that the circulation response begins in the stratosphere, and produces (rather than results from) a broad mid- and upper-tropospheric temperature response in the subtropics. Wu et al. (2013) take a closer look at changes in the structure and behavior of eddies and implicate the changing index of refraction in the jet shift, rather than changes in critical latitude, as in Chen and Held (2007). Chen et al. (2013) examine an ensemble of switch-on uniform sea surface warming experiments in an aquaplanet simulation and document a two-stage circulation response, consisting of an initial, transitory widening of the Hadley cell edge during the first few days and a gradual, steady poleward shift of both the Hadley cell boundaries and the eddydriven jets.

In this study, we document the transient zonal-mean response to several individual forcings, each applied instantaneously. We examine the response to twice the observed ozone depletion since $1980 (2 \times O_3)$, 4 times the preindustrial greenhouse gas concentrations (4 × CO₂), and projected twenty-first-century sea surface warming (SST₂₁₀₀). To distinguish between the direct and indirect circulation response, we perform atmosphere-only simulations, in contrast to Wu et al. (2012), who couple their atmosphere to a slab ocean. Our simulations include topography, unlike the aquaplanet simulations of Chen et al. (2013).

In addition to documenting the transient direct and indirect response of the zonal-mean circulation, we investigate changes in midlatitude eddies, comparing the changes we see to those representative of the mechanisms proposed by Chen and Held (2007), Kidston et al. (2011), and Lorenz (2014b). In this way, we hope to gain some insight into the relative importance of these proposed mechanisms in each step of the transient circulation response.

The changing pattern of eddy momentum flux and the intensification of wave reflection on the poleward flanks of the SH jet in our SST warming experiment are in line with the mechanism proposed by Lorenz (2014b).

Changes in eddy phase speeds in our warm ocean scenario are also correlated with poleward jet shifts, and some evidence for the mechanisms of both the Chen and Held (2007) and Kidston et al. (2011) mechanisms can be seen in the SST₂₁₀₀ experiment after the first 12 days. However, these mechanisms to some degree are mutually exclusive (Lorenz 2014b); specifically, the efficacy of the (Lorenz 2014b) mechanism implies a minimal role of the Kidston et al. (2011) mechanism and a reversed role of the Chen and Held (2007) mechanism.

We outline our model framework and our methodology in section 2. We then present the results of our work in a two-part fashion. In section 3, we document the ensemble-mean zonal-mean temperature, zonal wind, geopotential height, and meridional circulation response for our six experiments. In section 4, we examine the changing spectra of midlatitude eddies, and discuss their role in modifying the zonal-mean circulation. We summarize our results in section 5.

2. Data and methods

We examine output from transient simulations performed with the Geophysical Fluid Dynamics Laboratory Atmosphere Model, version 2.1 (AM2.1) (Delworth et al. 2006; Anderson et al. 2004). We first describe our AM2.1 model framework and simulations, before explaining our circulation metrics.

a. The AM2.1 model formulations

Our AM2.1 framework is similar to that in Staten et al. (2012) and Staten and Reichler (2013). AM2.1 has a finite volume dynamical core, with 2° latitude by 2.5° longitude resolution and 24 levels in the vertical. We also perform smaller ensembles using a stratosphere-resolving version of the model. These 48-level ensembles produce similar circulation responses and are not documented here.

We analyze two types of experiments in this study, in addition to a control (see Fig. 1). The experiments are listed in Table 1 and described below. The control simulation consists of a 3500-yr uncoupled preindustrial integration, with prescribed concentrations of greenhouse gases, zonal-mean stratospheric ozone, and prescribed sea ice coverage and sea surface temperatures (together referred to as SSTs) calculated from a coupled preindustrial control simulation, all held at preindustrial levels from year to year.

The first set of experiments we examine in this study is composed of 640-member transient ensembles, with each member initialized from a different 1 January from the control simulation. We prescribe the assigned switch-on forcing and integrate the model for 90 days



FIG. 1. Illustration of the ensemble generation method described in section 2a. Shown are 2-week-smoothed zonal-mean temperature anomalies with respect to the control climatology at 50 hPa, averaged over 35°–55°N for 5 yr of the control and time-slice experiments. Switch-on experiment temperature anomalies (dark blue curves) can be seen transitioning from their respective control values (black curve) toward the climatology of an identically forced time-slice simulation (light blue).

(refer to the thick blue curves in Fig. 1). Our forcings include twice the observed ozone depletion calculated using zonal-mean values from the Randel and Wu (2007) dataset, 4 times the preindustrial CO₂ concentrations, and 10-yr average SSTs from the years 2091-2100 in a coupled A1B scenario from CM2.1. The prescribed SST warming varies by longitude, latitude, and season and amounts to roughly 1.8-K warming in the global mean during our transient experiments. For more details on the structure of the applied SST warming (refer to Fig. S1 in the supplementary material). Differences between control and spinup simulations are calculated using only the respective years of the control simulation. To more closely examine the response in the transient SST_{2100} simulations, we include 1740 additional SST₂₁₀₀ January-only integrations, for a total of 2380 January integrations for the SST_{2100} .

The second set of experiments consists of three 501-yr time-slice simulations with constant forcings identical to those in each of our spinup simulations. These fixed-SST time-slice simulations equilibrate quickly, and we compute the equilibrium response as the average over 500 yr, leaving out the first year. By comparing the

TABLE 1. The simulations examined in this study.

Name Control	Туре	
	Time slice	3500 years
$2 \times O_3$	Spin up	640 members
	Time slice	500 years
$4 \times CO_2$	Spin up	640 members
	Time slice	500 years
SST ₂₁₀₀	Spin up	640 members
	Spin up (January)	2380 members
	Time slice	500 years



FIG. 2. A sample circulation feature calculation, based on the monthly-mean Ψ_{500} during one control January (specifically the year 2970). Note that the weighted FCC over the NH follows the center of the Ferrel cell, rather than the peak, as described in section 2b.

transient response in the spinup ensembles to the equilibrium response from the time-slice experiments, we can better discern the transient response from the seasonal cycle of the equilibrium response.

b. Circulation indices

In section 3b, we document changes in the latitudes of the intertropical convergence zone (ITCZ), the poleward edges of the Hadley cells, and the latitude of the eddy-driven jet, all derived from the meridional streamfunction at 500 hPa (Ψ_{500}). The ITCZ and the Hadley cell edges are based off of streamfunction zero crossings, while the latitude of the eddy-driven jet is estimated from the Ferrel cell center. To focus on latitudinal shifts in the jet, rather than changes in jet structure, we calculate the Ferrel cell center as the Ψ_{500} weighted mean latitude of either 1) positive Ψ_{500} south of the SH Hadley cell or 2) negative Ψ_{500} north of the Northern Hemisphere (NH) Hadley cell (see Fig. 2).

c. Reflecting levels and critical latitudes

In section 4a we evaluate some possible mechanisms by which eddy-mean flow interactions may shift the jet poleward. These mechanisms focus on Rossby wave propagation—specifically on critical latitudes and reflecting levels. These structures can be understood from the index of refraction for Rossby waves, which is derived from the dispersion relation for the wave solution of the vorticity equation on a sphere. While a more complete index of refraction can be calculated for the quasigeostrophic system (Harnik and Lindzen 2001), in the interest of clarity, we focus on the barotropic vorticity equation. The dispersion relation for a linear, unforced, barotropic plane wave perturbation on background zonal-mean zonal wind in Mercator coordinates can be written as

$$c_{\omega} = \frac{\overline{u}}{a\cos\phi} - \frac{\beta^* a\cos\phi}{m^2 + l^2},\tag{1}$$

where c_{ω} is the angular phase speed of the waves about Earth's axis; \overline{u} is the background zonal-mean zonal wind; *a* is the radius of Earth; ϕ is latitude; β^* is the zonalmean absolute vorticity gradient; and *m* and *l* are the zonal and meridional wavenumbers, respectively. Solving for the index of refraction, or l^2 , we have

$$l^{2} = \frac{\beta^{*}a \cos\phi}{\overline{u}(a \cos\phi)^{-1} - c_{\omega}} - m^{2}.$$
 (2)

Linear barotropic waves tend to propagate from small values of l^2 to large values of l^2 and dissipate in the limit of $l^2 \rightarrow \infty$. This occurs when

$$c_{\omega} = \overline{u}(a\cos\phi)^{-1} \tag{3}$$

or when the zonal phase speed of the eddies matches the zonal-mean background flow. The latitude at which this occurs is termed the critical latitude. In the presence of even slight damping, waves are dampened out before reaching their critical latitude.

The latitude at which $l^2 \rightarrow 0$ for a given wavenumber and phase speed is termed the reflecting (or turning) latitude. This latitude is reached when

$$c_{\omega} = \frac{\overline{u}}{a\cos\phi} - \frac{\beta^* a\cos\phi}{m^2}.$$
 (4)

We calculate this reflecting latitude and the critical latitude described above for a range of wavenumbers, although we focus on wavenumber 6 in this work. When the latitude—rather than phase speed—is held constant, we define the reflecting level as the phase speed below which waves are reflected. While Lorenz (2014b) focuses on wavenumber 8, in our experiments most of the anomalous wave activity takes place at or below wavenumber 6. In either case, the reflecting-level profiles for the different wavenumbers are functionally similar, although profiles for lower wavenumbers affect slower waves.

d. Eddy cospectra

We examine changes in eddy spectra during the first 48 days of our spinup ensembles, and the first 48 days of the year in our time-slice climatologies. We first compute the space–time cospectra of daily horizontal winds over 12-day intervals, tapered with a Hanning window, and transform the spectra from frequency–wavenumber space to angular phase speed–wavenumber space as in Randel and Held (1991). The use of this 12-day window implies a coarse spectral resolution, and we are unable to resolve waves traveling slower than about 8 m s⁻¹. On the other hand, we find that performing the calculations over longer time periods obscures the changes occurring

during the first few critical weeks. The 12-day period, then, represents a compromise between temporal and spectral resolution.

Covariance calculations like those examined here can be quite noisy. To confirm changes seen in the eddy flux spectra in the SST_{2100} spinup ensemble, we perform additional simulations for this experiment (see Table 1). These results form the basis for much of our discussion in section 4.

We plot momentum flux spectra for qualitative comparison with (Lorenz 2014b), but to quantify changes in eddy phase speed and the latitude of wave activity associated with the eddy-driven jet we also calculate the mean latitude and phase speed of eddy activity using the kinetic energy spectrum, as it is positive definite. These phase speed and latitude calculations are calculated as the weighted mean latitude and phase speed of eddy kinetic energy poleward of $\pm 10^{\circ}$.

3. Documenting the zonal-mean circulation response

In this section, we document the zonal-mean transient response to switch-on forcings and changing SSTs. We begin by discussing the zonal-mean responses of temperature, zonal wind, and polar cap averaged geopotential height during the first 90 days of the spinup simulations in section 3a. We then document changes in the meridional circulation in section 3b.

a. The temperature and zonal wind response

While radiative forcings initially impact temperatures most strongly in the stratosphere and SST warming initially warms the atmosphere near the surface, the zonalmean zonal wind response usually begins near the tropopause in each of our experiments. Here we analyze this step-by-step change in zonal-mean temperature and wind (Figs. 3 and 4). We discuss the response in each ensemble, in turn.

1) $2 \times O_3$

The SH ozone depletion-induced temperature response (Figs. 3a,b, top, and Fig. 4a, top 2 panels) is weak compared to that seen in the Staten et al. (2012) and in Fig. 5a (top 2 panels), because of our spinup framework. Staten et al. (2012) show that the seasonal SH ozone hole anomaly weakens by January (see their Fig. 1), whereas the ozone hole in our study is only prescribed after 1 January. That is, the ozone hole in the present study winds up while the seasonal ozone depletion forcing winds down. Thus, while the effects of stratospheric cooling can be seen to impact the troposphere after about two weeks (Fig. 3d, top), the peak tropospheric



FIG. 3. Latitude-height plots of spinup ensemble-mean zonal-mean (a),(b) temperature and (c),(d) wind averaged over three time periods: (a) day 1; (b),(c) days 1–12; and (d) days 13–24. Control values are contoured every 10 K or 10 m s^{-1} , and spinup anomalies are shaded as shown. [Red lines in top panel of (c) demark the boundaries for temperature and zonal wind averaging in Figs. 4 and 5.]

response occurs around day 70 (Fig. 4a, bottom 4 panels).

The timing of the onset of our switch-on forcing, combined with the seasonality of stratospheric ozone anomalies, allows for a more convenient interpretation of the temperature and wind response over the NH, however. We interpret the NH wind and temperature anomalies as follows: During boreal winter, the lack of insolation implies a weak forcing over the region. Thus, the anomalous cooling due to stratospheric ozone depletion in the NH stratosphere is very weak compared to the rest of the stratosphere during days 1–12 (Fig. 3b, top). The anomalous poleward temperature gradient imposed by this differential forcing is associated with an easterly wind anomaly in the stratosphere (Fig. 3c, top), likely making the background zonal flow even more favorable to the vertical propagation of planetary waves from the troposphere (Charney and Drazin 1961) and possibly leading to further warming and deceleration by days 13–24 (Fig. 3d, top). The background conditions favorable to this positive feedback appear strongly seasonal, however, and the pressure and wind responses over the NH reverse around day 40, both in the transient response (Fig. 4a, panels 3 and 5 from top) and in the equilibrium simulations (see Fig. 5a, bottom 4 panels).

2) $4 \times CO_2$

The structure of the imposed temperature anomaly because of greenhouse gas increases changes little during our simulation, but the amplitude of the anomaly increases at least through day 60 (Fig. 4b, top 2 panels). The temperature response includes the usual fast stratospheric adjustment (Fig. 3a, middle), as well as a tropical upper-tropospheric warming maximum peaking at about 100 hPa (Fig. 3b, middle), although the latter appears transient, as it is absent from the equilibrium response. Over the NH, the temperature gradient points poleward only above about 50 hPa; over the SH, the gradient points poleward above 300 hPa.

These contrasting temperature gradient anomalies are followed by contrasting zonal wind anomalies in the troposphere (Figs. 3c,d, middle). The NH polar vortex and midlatitude surface westerlies decelerate, perhaps for reasons similar to those described above for the deceleration due to stratospheric ozone depletion. This stratospheric deceleration is defined in Wu et al. (2012)



FIG. 4. (a)–(c) Time–height plots for the spinup experiments. Shown are ensemble-mean zonal-mean temperature (top 2 panels) and zonal wind (panels 3 and 4 from top), averaged poleward of the zonal wind maxima as marked in Fig. 3c (top), specifically 35° – 55° N (panels 1 and 3 from top) and 50° – 70° S (panels 2 and 4 from top). Also shown are corresponding ensemble-mean zonal-mean geopotential heights (multiplied by -20; bottom 2 panels) averaged poleward of 70° over each hemisphere. Control values are contoured every 10 K, 10 m s^{-1} , or 10 km, and spinup anomalies are shaded as in Fig. 3. Note that increasing heights in the bottom 2 panels are shaded blue.



°C (rows 1–2), ms⁻¹ (rows 3–4), -20 m (rows 5–6)

FIG. 5. As in Fig. 4, but for the 500-yr time-slice experiments. Shown are climatological-mean zonal-mean temperature (top 2 panels), zonal wind (panels 3 and 4 from top), and geopotential heights (bottom 2 panels).

as the first phase in the transient response. However, it is also present in the equilibrium CO_2 simulation (Fig. 5b, panels 3 and 5 from top); indeed, most of the features of the NH transient zonal wind response in Fig. 7b in Wu et al. (2012) are present in the equilibrium response, suggesting that the seasonal cycle of the equilibrium response over the NH dominates the transient response.

In contrast to the tropospheric wind anomalies over the NH, SH tropospheric westerlies exhibit—at least during days 13–24—an increase in zonal-mean zonal winds primarily poleward of the midlatitude jet, as seen in previous studies (Figs. 3c,d, middle). As with $2 \times O_3$, wind anomalies grow quickly in the stratosphere, propagating into the troposphere on a time scale of weeks (Fig. 4b, panels 3 and 4 from top).

Also in contrast to the NH wind response (Fig. 4b), the SH transient response and the seasonal cycle of the equilibrium response are very different over the SH. While the seasonal cycle of the equilibrium response suggests a downward component in March (Fig. 5b, bottom), the transient response exhibits a distinct downward propagation of anomalies during the entire 90-day integration period (Fig. 4b, bottom) as in Fig. 7d in Wu et al. (2012). The downward-propagating response, then, appears to be a robust component of the transient response to the imposed radiative forcing.

3) SST₂₁₀₀

In climate sensitivity studies, temperature-dependent feedbacks are often considered the slow response of the climate system because of the ocean's large heat capacity and the resulting slow response of SSTs (Solomon et al. 2007). On the other hand, the atmospheric response to warming SSTs is the most rapid in this study (Fig. 4c, top 4 panels), because of our switch-on forcing experimental framework. That is, while direct radiative forcings warm the atmosphere on radiative time scales (a few weeks), SST warming warms the troposphere primarily on dynamical time scales (a few days).

The quiescent stratosphere above, however, still warms and—over the tropics—cools on radiative time scales. Interestingly, the stratospheric heating in the equilibrium simulations (Fig. 5c, top 2 panels) is smaller than in our spinup simulations. This suggests that the initial SST warming "shock" induces a relatively large dynamical warming in the lower stratosphere, and the stratosphere "overshoots" radiative–dynamical equilibrium, at least for the duration of our switch-on experiments.

Although SST₂₁₀₀ temperature anomalies initially propagate upward from the surface (Fig. 4c, top 2 panels), the zonal-mean zonal wind response (Fig. 4c, panels 3 and 4 from top) peaks in the upper troposphere– lower stratosphere as with the $2 \times O_3$ and $4 \times CO_2$ experiments. Extratropical westerly anomalies then propagate down to the surface after about 10 days, again similar to that in $2 \times O_3$ and $4 \times CO_2$. By the end of our spinup experiments, the vertical structure of our transient SST₂₁₀₀ response matches that of the equilibrium response (Fig. 5c) qualitatively, although the full magnitude of the tropospheric temperature, pressure, and wind response is not realized during our spinup experiment.

b. Meridional circulation indicators

While the changes in zonal-mean temperature, wind, and pressure described above are of dynamical interest, their implications for surface climate are of societal interest. These effects on surface climate can in turn be inferred from threshold indicators of the overturning circulation. In this section, we document changes in latitude of meridional circulation features described in section 2 (Fig. 6). Recalling the positive annular mode– like anomalies in polar cap heights described above (and seen in Fig. 4, bottom), one may expect a general poleward shift in circulation features over the SH.

Although some uncertainty remains in these shifts, even with large ensembles, overall southward shifts are seen in the SH circulation for each ensemble. For example, even with the late onset of the SH stratospheric ozone hole, the southward shift of the SH jet [labeled the Ferrel cell center (FCC)] and the Hadley cell edge (HC|FC) in the $2 \times O_3$ experiment are unambiguous (Fig. 6a).

The gradual widening of the Hadley cell from the $4 \times$ CO₂ ensemble (Fig. 6b) is in agreement with the equilibrium widening seen in Staten et al. (2012). We do note, however, that while the SH jet shifts poleward, it also weakens in this case. The poleward shift of the NH FCC is surprising, given the weakening of the stratospheric polar vortex and the dipole of the zonal winds about the NH subtropical jet (Fig. 3d, middle).

Both the ITCZ and SH HC|FC exhibit a transient northward shift in the SST_{2100} spinup simulation, before trending toward their equilibrium values in February and March. This initial contraction of the tropics, and the broader, gradual overall widening is reminiscent of that in Simpson et al. (2012) and Chen et al. (2013), even though these studies maintain constant relaxation profiles, solstitial SSTs, and aquaplanet lower boundary conditions.

The HC|FC and FCC both shift poleward in the SST_{2100} experiment, with shifts over the NH occurring within the first week and shifts of the SH FCC occurring after about four weeks, on similar time scales to that in the 2 × O₃ and 4 × CO₂ experiments. In each case, the shifts in the HC|FC and FCC are strongly correlated



FIG. 6. Time series of smoothed zonal-mean Ψ_{500} threshold anomalies for each spinup experiment (filled curves) and equilibrium experiment (stroked curves), scaled as shown in each panel. Zero lines for each anomaly are separated for readability. Thresholds include the integrated FCC, calculated after Staten and Reichler (2013), the HC | FC boundary, and the ITCZ. Shifts away from (toward) the equator are shaded dark (light).

(Kang and Polvani 2011; Staten and Reichler 2013). The poleward shifts of the SH Hadley cell edge and eddydriven jet are among the most consistent transient shifts in our spinup experiments.

4. The role of eddies

The preceding sections document the zonal-mean temperature, pressure, and zonal wind responses. However, in order to understand the events leading to these changes, it is necessary to examine changes in the eddies as well. In the following sections, we examine eddy spectra and assess the extent to which several proposed mechanisms may play a role in the general circulation shifts described previously. These mechanisms center on changes in Rossby wave propagation and dissipation, as described in section 2c. We first examine changes in barotropic wave reflection, following Lorenz (2014b), in section 4a. We then discuss changes in the phase speed and the associated change of critical latitude for wave absorption on the equatorward flanks of the jet (Chen and Held 2007) in section 4b. Last, we discuss the potential effects of changes in eddy length scale on the critical latitude for wave absorption on the poleward flanks of the jet as in Kidston et al. (2011) in section 4c.

All three of the mechanisms below rely on the nonacceleration theorem and the pseudomomentum rule (Eliassen and Palm 1961; Charney and Drazin 1961; Dunkerton 1980). The former states that in the absence of wave growth or dissipation, the zonal-mean flow does not change. The latter states that eddies deposit momentum where they form and from which they propagate (i.e., the jet core or region of maximum baroclinic instability) and remove momentum from the latitudes in which they dissipate (e.g., near the subtropics). The mechanisms below all involve some initial change in the mean flow that often occurs in the stratosphere and serves to change eddy propagation. These changes in eddy propagation then alter the statistics of eddy momentum flux convergence and divergence, which in turn alters the mean flow. In the case of the mechanisms described in the following sections, the changing eddy statistics help to shift the jet poleward.

a. Wave reflection

Lorenz (2014b,c) highlights the role of wave reflection in maintaining poleward shifts of the midlatitude jet. Midlatitude Rossby waves develop at the core of the midlatitude jet, in regions of strong baroclinicity. These waves propagate zonally and meridionally away from the jet. Waves that propagate poleward encounter regions of low index of refraction l^2 and are often reflected



FIG. 7. Phase speed–latitude plots of ensemble-mean total poleward eddy momentum flux at 250 hPa for (a)–(c) the experiments and (top),(bottom) the different time periods. Climatological poleward momentum flux (i.e., equatorward wave flux) is contoured in black every $5 \times 10^{-2} \text{ m s}^{-1}$, with anomalies shaded in red. Climatological equatorward momentum flux (i.e., poleward wave flux) is contoured gray, with anomalies shaded in blue. The zonal-mean background zonal wind divided by $\cos\phi$ is also plotted for both the control (black solid) and spinup (gray solid) cases. Profiles of reflecting level [as defined in Eq. (4)] for zonal wavenumber 6 are also plotted for the control (black dashed) and spinup (gray dashed) cases.

equatorward, as depicted in Fig. 3 of Lorenz (2014b). Rather than transporting momentum from the poleward flanks of the jet, these waves transport momentum from the equatorward flanks, converging the momentum in their source region at the jet core. By modifying wave sources to simulate specific aspects of the changing climate as described in Lorenz (2014a), Lorenz (2014b) produce just such an amplification of the jet and an increase in equatorward reflection along its poleward flanks.

Kidston and Vallis (2012) note that an increase in jet speed is accompanied by an increase in wave reflection away from high latitudes and attribute the increase in reflectivity (or decrease in the index of refraction) to the increased meridional shear on the poleward flanks of the jet and the reduced absolute vorticity gradient β^* at these latitudes [see Eq. (2)]. The work of Lorenz (2014b) simplifies the discussion by focusing on which phase speeds will be reflected for a given pressure level, latitude, and zonal wavenumber, as given by Eq. (4).

Here we test whether the signatures from Lorenz (2014b) are present in the eddy momentum flux cospectra anomalies from our spinup simulations. In the radiatively forced experiments (Figs. 7a,b), the reflecting level—the phase speed below which waves are reflected—changes



FIG. 8. Scatterplots of the centroid latitude of eddy kinetic energy vs (a) peak wavenumber-6 reflecting-level phase speed, (b) integrated eddy phase speed, and (c) integrated relative phase speed for the spinup experiments (shown by color), during the first four 12-day time periods in the spinup experiments (labeled 1–4). Numbers are offset from values for readability.

little over the SH, even as the circulation indices (section 3) shift poleward. Thus, the poleward shift of the jet in these experiments occurs in the absence of a strong change in the index of refraction. As the $2 \times O_3$ and $4 \times CO_2$ experiments both yield such weak eddy and zonal wind responses during the first weeks, we focus our discussion on changes seen in the SST₂₁₀₀ experiment.

Warming SSTs amplify the climatological pattern of eddy momentum flux over the SH during the first 12 days of our SST₂₁₀₀ spinup simulations (Fig. 7c, top). Chen et al. (2013) see a similar increase in wave activity in response to warming SSTs and attribute this increase to an increasing flux of wave activity from the surface or (equivalently) an increase in poleward eddy heat flux. This vertical wave activity anomaly is confirmed in our study (Fig. S2 of the supplementary material).

The initial burst of the upward wave activity flux during days 1–12 is followed by a downward flux in the SH subtropics below 200 hPa during days 13–24, and the associated strong divergence of Eliassen–Palm flux in the upper troposphere has been attributed to a reduction of wave activity dissipation through a carefully designed finite-amplitude wave activity budget by Chen et al. (2013). These authors further argue that this reduction of wave dissipation plays an integral part in modifying irreversible potential vorticity (PV) mixing and in shifting the jet. The exact finite-amplitude wave activity analysis is beyond the scope of the present study and is a topic of future investigation.

The initial increase in wave activity should accelerate the eddy-driven jet. Indeed, a small increase in zonalmean zonal winds can be seen during the first 12 days, along with an increase in the reflecting level. By days 13–24 (Fig. 7c, bottom), the increase in both the zonal winds and the reflecting level is quite pronounced.

As expected from Lorenz (2014b), this increase in zonal winds is accompanied by a poleward shift of the SH eddy-driven jet, judging from the stronger increase in westerlies on the poleward flanks of the jet and in the poleward shift of eddy momentum flux (Fig. 7c). The signatures of the Lorenz (2014b) mechanism appear during days 13-24. Specifically, cross-jet equatorward wave flux (i.e., poleward momentum flux in the jet core region) increases in the same neighborhood where the reflecting level increases and at phase speeds where additional waves are likely to be reflected (Fig. 7c, bottom). This reflecting-level increase at 250 hPa is a robust feature of the upper troposphere (Fig. S2a in the supplementary material). During the first 48 days of our SST₂₁₀₀ experiment, the jet shifts poleward approximately linearly with the reflecting-level increase (Fig. 8a). Changes in the eddy flux for wavenumber 6 alone resemble those for the total wave flux.

A poleward shift of the jet and the eddy flux convergence signature of the Lorenz (2014b) mechanism are both present over the NH as well. However, the initial amplification of the existing eddy momentum flux pattern at 250 hPa is missing (Fig. 7c, top), in spite of 1) a strong increase in poleward eddy heat flux (i.e., upward wave activity flux) over the NH during the first 12 days (Fig. S2a) and 2) strengthened westerlies on the poleward flanks of the jet.

Lorenz (2014b) discuss the dependency of the reflectivity profile on the structure of zonal winds, citing Barnes and Hartmann (2011) as an example of a study with a contrasting reflectivity profile. The equatorward extension of our climatological reflecting-level profile into the wave source region of the eddy-driven jet in the SH is a notable departure from the more confined reflecting-level profile in Lorenz (2014b). However, inasmuch as 1) the reflecting levels are primarily poleward of the wave source region and 2) peak reflecting-level phase speeds increase, it seems likely that the mechanism cited in Lorenz (2014b) contributes to poleward maintenance of the jet.

b. Eddy phase speeds

The second mechanism we examine in this section is that proposed by Chen and Held (2007), who argue that anomalous lower-stratospheric westerlies result in an increase in eddy phase speeds. This increase in phase speeds implies that waves will reach their critical latitude where their phase speed matches the zonal-mean background flow: nearer to the jet core. Inasmuch as the majority of eddies propagate equatorward, this would imply a shift in dissipation closer to the jet on the equatorward side, reducing baroclinicity on the equatorward flank of the jet core and pushing the primary region of baroclinic generation of eddies poleward. On the other hand, shifts in the critical latitude on the poleward side of the jet may shift the jet equatorward.

The increase in lower-stratospheric winds over the midlatitude jet cores is clear over both hemispheres for all three of our experiments (Fig. 3), even in the presence of a weakening NH stratospheric polar vortex in the radiative forcing experiments. In addition, most of the eddy momentum flux in our simulations is equatorward, so we may expect an increase in eddy phase speeds alone to shift the jet poleward. Such anomalies are very small for the radiatively forced experiments (Figs. 7a,b, bottom). Furthermore, flux anomalies over the NH for the SST₂₁₀₀ experiment (Fig. 7c) are primarily poleward of the jet, in the region of increased wave reflection, rather than on the equatorward flanks near the critical latitude. Changes in flux over the SH reveal an increase in phase speeds and a poleward shift of the jet as predicted after the first 12 days (Fig. 8b), although zonalmean zonal winds on the equatorward flanks of the SH jet in the SST_{2100} experiment also increase, which may allow waves to penetrate farther into the subtropics and weaken the poleward shift of the jet.

As highlighted in Lorenz (2014b), the Chen and Held (2007) mechanism does not account for the existence of the reflecting level. That is, while increasing phase speeds should indeed shift the critical latitude for wave absorption closer to the jet, the existence of a reflecting level implies that an increase of phase speeds (but not of the reflecting level) should also allow more waves to propagate poleward beyond their previous reflecting

latitude and break on the poleward side of the jet, pushing the jet equatorward. Lorenz (2014b) argues that this equatorward push resulting from increased poleward propagation would outweigh the poleward push due to a shifting critical latitude. Thus, while we cannot eliminate the Chen and Held (2007) mechanism as a possible contributor, the signatures for their mechanism in the early days of our spinup simulations may be better explained by the changing reflecting latitude, and the structure of changes in the zonal flow and the reflecting level seem to favor the Lorenz (2014b) mechanism as well. It is possible, however, that the Chen and Held (2007) mechanism may play a role as a feedback, once initial changes in the zonal flow have taken place.

c. Eddy length scale

The third and final mechanism we discuss in this section is that proposed by Kidston et al. (2011). They focus on the flow-relative phase speed of the jets, noting that the observed and modeled increase in eddy length scale in a changing climate should be accompanied by a decrease in the flow-relative phase speed of midlatitude waves as can be seen from the second term on the righthand side of Eq. (1)]. This ought to allow waves to propagate farther poleward from the jet core before reaching their critical latitude poleward of the jet. In essence, while Chen and Held (2007) and Kidston et al. (2011) both acknowledge the increase in phase speeds in a shifting climate, Kidston et al. (2011) emphasize the role of eddy dissipation and accelerating zonal winds on the poleward flank of the jet. Like Chen and Held (2007), Kidston et al. (2011) ignore changes in wave reflection.

The strong increase we see in zonal winds from the first 12 days gives us good reason to examine the Kidston et al. (2011) mechanism. If zonal-mean zonal winds increase initially and if eddy phase speeds increase accordingly, the critical latitude for wave absorption should remain steady, unless an additional factor comes into play. In the Lorenz (2014b) mechanism, this additional factor is wave reflection on the poleward flank. In the Chen and Held (2007) mechanism, this factor is a weaker increase of zonal wind on the equatorward flanks of the jet. In the Kidston et al. (2011) mechanism, this factor is an increasing length scale and the asymmetric structure of eddy dissipation about the jet. If dissipation is far enough removed on the poleward flank of the jet, Kidston et al. (2011) argue that the net eddy momentum flux convergence on the poleward flank of the jet may increase.

Flow-relative phase speeds decrease strongly during the first 12 days (Fig. 8c). On the other hand, flowrelative phase speeds do not monotonically decrease during the successive 12-day periods, although the jet continues to shift poleward. In addition, while some additional poleward wave flux is seen beyond the critical latitude of wave absorption for the SH jet (blue shading in Fig. 7c, bottom) and eddy-length scales indeed tend to increase in the SST_{2100} experiment (not shown), this change occurs primarily after day 12. Finally, this mechanism fails to explain the increase in equatorward eddy momentum flux in the neighborhood of the reflecting latitudes. Of the three mechanisms examined in this work, the Kidston et al. (2011) mechanism explains the least of the zonal wind and eddy momentum flux changes that we see. In addition, the missing relationship between the relative phase speeds and the jet latitudes after the first 12 days suggests that, if the Kidston et al. (2011) mechanism does play a role, it is likely minimal.

5. Conclusions

We document the transient response of the atmospheric global circulation to instantaneously applied climate forcings and sea surface warming in the context of a seasonally varying background climate. We further investigate the roles of three proposed mechanisms in the poleward shifts seen in the midlatitude jet. We find that 1) the zonal-mean zonal wind response begins near the upper troposphere–lower stratosphere region for both radiatively forced and warm SST experiments and 2) changes in wave reflection may help to explain the jet shifts in response to warming SSTs.

Greenhouse gas increases, stratospheric ozone depletion, and sea surface warming generally produce an anomalous equatorward temperature gradient in the upper troposphere-lower stratosphere region. Westerlies in this region accelerate and, after about a week, wind anomalies spread into all levels of the free troposphere. An exception to this chain of events occurs in the $2 \times O_3$ experiment over the NH polar stratosphere, where the forcing due to ozone depletion is weak. The presence of lower-stratospheric temperature and wind anomalies early on in each response underscores the importance of the wind structure in the region in modulating the circulation response throughout the troposphere (Song and Robinson 2004).

Section 4 constitutes an early validation of the work of Lorenz (2014b) using a full general circulation model. Barotropic wave reflection on the poleward side of midlatitude jets strengthens during the first four consecutive 12-day periods of our SST₂₁₀₀ experiment and is well correlated with a poleward shift in wave activity. Equatorward cross-jet wave flux anomalies also suggest a role of wave reflection. The initial strengthening of the

jet, which Lorenz (2014b) suggests triggers the change in reflecting level, can be seen during the first 12-day periods and corresponds with an increase in upward wave flux from the surface as in Chen et al. (2013). We interpret these facts as evidence for the efficacy of the Lorenz (2014b) mechanism in initiating—or at least maintaining—a poleward shift of the jet in response to SST warming and recommend further study of the initial jet intensification in response to SST warming.

One limitation of this study is the inability of our experimental design to simulate the full extent of the cooling of the SH lower stratosphere because of ozone depletion. The observed ozone depletion peaks in October, but we start our simulations on the first day of January and prescribe forcings for the actual months. However, this work has shown that the principle of downward progression of circulation anomalies through the troposphere is not specific to individual radiative forcings and is likely to apply in the presence of stronger cooling of the lower stratosphere as well. Another potential shortcoming of our study is the focus on the index of refraction calculated from the barotropic vorticity equation, rather than a more complete formulation. On the other hand, the utility of even the barotropic vorticity equation highlights the explanatory power of the Lorenz (2014b) mechanism.

Acknowledgments. The author would like to thank Paulo Ceppi for his valuable input and three anonymous reviewers for the constructive comments and suggestions. We acknowledge the University of Utah CHPC for computing support. AM2.1 model data were archived at the National Energy Research Scientific Computing Center. Support of this research is provided by an NSG-GK12 grant. The work described in this publication was performed at the University of Utah. The writing and publication of this publication was supported by the JPL, Caltech, under a contract with NASA. Jian Lu was supported by the Office of Science and the U.S. Department of Energy, as part of the Regional and Global Climate Modeling program.

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