The Influence of the Basic State on the Northern Hemisphere Circulation Response to Climate Change

MICHAEL SIGMOND

University of Toronto, Toronto, Ontario, Canada

JOHN F. SCINOCCA

Canadian Centre for Climate Modelling and Analysis, Victoria, British Columbia, Canada

(Manuscript received 1 April 2009, in final form 7 October 2009)

ABSTRACT

Employing a comprehensive atmospheric general circulation model, the authors have shown in a previous study that the time-mean Northern Hemisphere (NH) winter circulation response to a CO_2 doubling perturbation depends significantly on parameterized orographic gravity wave drag (OGWD) parameter settings, which are essentially related to the strength of OGWD. A possible implication is that aspects of the greenhouse gas–induced circulation response could depend directly on the formulation and internal parameters settings of the OGWD scheme. Such a result would further heighten the importance of OGWD parameterizations for climate studies and have far-reaching implications for modeled projections of future climate change.

In this study the causal relationship between OGWD and changes in time-mean NH wintertime circulation response to CO_2 doubling is investigated. This is accomplished by introducing a methodology that allows one to hold the OGWD forcing fixed to its $1 \times CO_2$ value when CO_2 is doubled. Employing this methodology for perturbation experiments with different strengths of OGWD, the authors find that the changes in OGWD forcing due to CO_2 doubling have essentially no impact on the time-mean zonal-mean zonal wind response. The primary conclusion is that the OGWD influence is limited to its impact on the $1 \times CO_2$ basic-state climatology, which defines the propagation characteristics of resolved waves. Different strengths of OGWD result in control basic states with different refractive properties for the resolved waves. It is shown that the action of resolved waves, as well as their sensitivity to such differences in the control climatology, explains essentially all of the NH wintertime circulation sensitivity identified here and in a previous study. Implications for climate change projections and climate-model development are discussed.

1. Introduction

Over the last two decades, significant effort has been invested in the generation of credible projections of climate change (e.g., Solomon et al. 2007). Predictions of extratropical regional climate change in Northern Hemisphere (NH) winter has been particularly challenging since the current generation of atmosphere–ocean climate models shows a wide range in projected responses in that region. In particular, the part of the response that projects onto the northern annular mode (NAM) (Thompson and Wallace 2000) varies considerably from model to model (Miller et al. 2006). Because extratropical weather patterns heavily depend on the polarity of the NAM, the uncertainty in the prediction of the future NAM arguably represents one of the largest uncertainties in midlatitude regional climate predictions.

At the same time, evidence has mounted that the extratropical troposphere is influenced by the stratosphere. For example, Baldwin and Dunkerton (2001) have shown that low-frequency variations in the extratropical tropospheric circulation are preceded by similar variations in the stratospheric circulation. A number of studies have also suggested that the tropospheric circulation response to increasing greenhouse gases (GHGs) critically depends on the stratospheric representation in climate models. Shindell et al. (1999), for example, argued that, to reproduce circulation changes that project positively onto the NAM, climate models need a well-resolved stratosphere. Sigmond et al. (2008, hereafter S08), on

Corresponding author address: Michael Sigmond, University of Toronto, 60 St. George St., Toronto, Ontario M5S 1A7, Canada.
E-mail: sigmond@atmosp.physics.utoronto.ca

DOI: 10.1175/2009JCLI3167.1

the other hand, showed with a more controlled experimental setup that the time-mean tropospheric circulation response to increasing greenhouse gases is not modified when the model lid is lowered from the mesopause to the middle stratosphere and all other model settings are kept identical.

The S08 study considered a variety of present-day control simulations subjected to a doubled-CO_2 perturbation. The primary result of that study was that the time-mean tropospheric circulation response to increased GHG forcing was most sensitive to the parameter settings of the orographic gravity wave drag (OGWD) scheme, which were related to the OGWD strength. Simulations with parameter settings leading to relatively weak (strong) OGWD were found to produce a warming response in the form of a positive (neutral) projection onto the NAM. The strength of the OGWD also influenced the control climatology upon which the CO_2 perturbation was applied. The main difference was that the weak OGWD case (referred to as ''LOW-G'' in S08) had relatively strong winds in the NH midlatitude lower stratosphere in winter, while the strong OGWD case (''LOW'' in S08) had relatively weak winds in this location.

There would seem two possible explanations for the response sensitivity to OGWD settings found in S08. One possibility would be that the NH extratropical wintertime circulation response is directly influenced by the OGWD forcing response to CO_2 increases, which in turn could directly depend on the specific formulation and parameter settings of the OGWD scheme used. This is a very daunting proposition. The difficulty of OGWD parameterization for modeling present-day climate is already well appreciated. If the response of OGWD to CO_2 forcing was also sensitive to its internal parameters and if such sensitivity also influenced the warming response, then there would be very far-reaching implications. OGWD would be even more critical than previously thought and the credibility of climate change projections would require a level of OGWD validation that does not presently exist.

A second possible explanation is that differences in the climatology of winds and temperatures, brought about by the different settings of OGWD, are primarily responsible for the sensitivity of the NAM response to CO_2 forcing. The possible mechanism here would involve resolved waves. Different basic states could cause different behavior of the planetary waves under CO_2 doubling and result in the sensitivity found by S08. This alternative explanation implies that the influence of OGWD is limited to its influence on the basic state and that the circulation response to CO_2 forcing is essentially divorced from the OGWD scheme's specific formulation and parameter settings. This would seem the most desirable of the two explanations in that the focus would shift from solving the OGWD problem, which is extremely difficult, to reducing the biases of the control climate relative to observations, which is more tractable.

In practice, the influence of OGWD on the climate change response may arise through some complicated combination of the aforementioned mechanisms: that is, 1) directly by the changes in OGWD forcing due to increasing greenhouse gases and 2) indirectly through the influence of OGWD on the basic state of the control climate. In this paper, we will derive a method to distinguish between these possible mechanisms of direct and indirect influence. From this analysis we are able to conclude that it is the indirect influence of OGWD on the control climate that explains the results of S08 and, therefore, that the specific details of the OGWD parameterization scheme are not as critical as might have originally been thought.

The outline of the paper is as follows. In section 2 we provide a description of the model and experimental design. In section 3 the main results are presented and discussed. In section 4 we conclude with a brief summary and discussion of the implications of these results.

2. Model and simulations

In this study we use a very similar model setup to that used in S08, to which we refer the reader for details. We employ the operational version of the Canadian Centre for Climate Modelling and Analysis third-generation atmospheric general circulation model (AGCM3) (Scinocca et al. 2008) with 32 levels from the surface to 1 hPa at T63 horizontal resolution, which in S08 was referred to as the LOW model. The results comprise 40-yr time-slice simulations for the present-day climate ($1 \times CO_2$; i.e., control run) and for a doubled CO_2 ($2 \times CO_2$) climate. In the $2 \times CO_2$ simulations, the atmospheric CO_2 is doubled, while the sea surface temperature (SST) field is perturbed with a monthly varying anomaly calculated from an ensemble average over the World Climate Research Programme (WCRP) Coupled Model Intercomparison Project phase 3 (CMIP3) simulations at the time of CO_2 doubling (see S08 for details).

As with almost all other comprehensive climate models with a lid located below the mesosphere, this version of the model does not include a parameterization scheme for nonorographic gravity waves. The parameterization scheme for orographic wave drag is that of Scinocca and McFarlane (2000). This scheme not only accounts for OGWD associated with freely propagating waves, but also includes the drag for dynamics associated with low-level ''blocking.'' Two of the scheme's internal parameters relevant to the present study are 1) $G(\nu)$, a multiplicative factor that scales the amount of gravity wave momentum flux produced by the interaction of the circulation with the topography, and 2) Fr_{crit}, the inverse critical Froude number, which sets the maximum nondimensional amplitude that a parameterized wave may attain before it begins to break and deposit its momentum to the background flow. This parameter directly influences the elevation of momentum deposition and indirectly influences the amount of launch momentum when low-level blocking occurs [Eq. (29a) of Scinocca and McFarlane (2000)]. In general, these two types of internal parameters are common to OGWD schemes.

The sensitivity of the lower-stratospheric winds and temperatures to the values of these two parameters, and their adjustment for the purpose of polar-ozone studies with the Canadian Middle Atmosphere Model (CMAM), is discussed at length by Scinocca et al. (2008). Basically, two general settings are employed. The first combination, $[G(\nu), Fr_{crit}] = (1.0, 0.5^{1/2})$, referred to as the "STRONG" setting, is typically used in the tropospheric version of the model (AGCM3) so as to reduce both wind and mean sea level pressure (MSLP) biases. With respect to observations, this setting generally results in anomalously weak winds and warmer temperatures in the NH wintertime polar lower stratosphere. Simulations with these settings in S08 were referred to as the LOW simulations.

The second combination of parameters, $[G(\nu), Fr_{crit}] = (0.65, 0.375)$, referred to as a "WEAK" setting of the OGWD scheme, is typically used in CMAM for the purpose of polar-ozone chemistry studies. This setting tends to produce stronger winds and colder polar temperatures in the wintertime lower stratosphere. These colder temperatures come at the expense of larger MSLP biases when the WEAK setting is used. Simulations with these settings in S08 were referred to as the LOW-G simulations.

In Scinocca et al. (2008), it was shown that the climatology associated with the WEAK setting of the scheme are equally well obtained with adjustments to only one of these parameters: $G(\nu)$. Retaining the more physical value of $Fr_{crit} = 0.5^{1/2}$, nearly identical results for the WEAK setting of the scheme can be obtained with $G(\nu) = 0.25$. In S08, the primary sensitivity of the time-mean climate change response of the NH circulation was found to depend on whether the WEAK or STRONG setting of the OGWD scheme was employed. Since the variation of only one parameter lends to a more straightforward interpretation of results, in this study we focus on the response sensitivity to the multiplicative factor $G(\nu)$ alone, with $G(\nu) = 0.25$ in the WEAK drag case and $G(\nu) = 1.0$ in the STRONG drag case. In section 3 we will show that the same sensitivity identified in S08 is obtained when the WEAK and STRONG settings of the orographic scheme are prescribed by $G(\nu)$.

To determine the direct influence of OGWD on the circulation response to CO_2 doubling we undertake a series of "prescribed OGWD" simulations. The methodology is as follows: for each of the $1 \times CO_2$ simulations employing WEAK and STRONG settings of OGWD, a three-dimensional monthly mean climatology of the momentum deposition is diagnosed. Then, in subsequent simulations, this climatology of OGWD is used in place of the OGWD parameterization. As a first step, it is verified that prescribing OGWD in the $1 \times CO_2$ simulations results in climatologies that are consistent with the original control runs (not shown).[1] Next, $2 \times CO_2$ simulations are performed with prescribed OGWD forcing from the $1 \times CO_2$ simulations. In this way, the influence of changes in OGWD forcing due to CO_2 doubling is eliminated from the perturbative response. If the NH wintertime circulation response to CO_2 doubling for the WEAK and STRONG cases were found to be insensitive to the use of such prescribed OGWD, one could conclude that OGWD does not have a direct influence on the climate change response. The results of these simulations will be described in section 3b.

3. Results

a. *Sensitivity of circulation response to OGWD under CO_2 doubling*

As previously shown by S08, the time-mean NH circulation response to CO_2 doubling is very sensitive to settings of the OGWD. The December–February (DJF) MSLP response to CO_2 doubling (i.e., $2 \times CO_2$ minus $1 \times CO_2$) is significantly different in the WEAK drag case (Fig. 1a) compared to the STRONG drag case (Fig. 1b). The WEAK drag response is characterized by a hemispheric-wide meridional dipole pattern similar to the NAM pattern, whereas the STRONG response is more localized and largest in the North Pacific. The difference in response patterns can be quantified by the area-averaged spatial correlation coefficient north of 45°N between the MSLP response and the surface NAM in the $1 \times CO_2$ simulations (defined as the first EOF of MSLP north of 20°N), which is 0.70 for the WEAK case and only 0.02 for the STRONG case. These results are consistent with the sensitivity to WEAK and STRONG OGWD settings first identified by S08.

[1] It was found that replacing the three-dimensional monthly mean climatology with two-dimensional does not result in significant differences in the zonal-mean climate. This suggests that the zonal-mean climate does not depend on the longitudinal structure of OGWD.

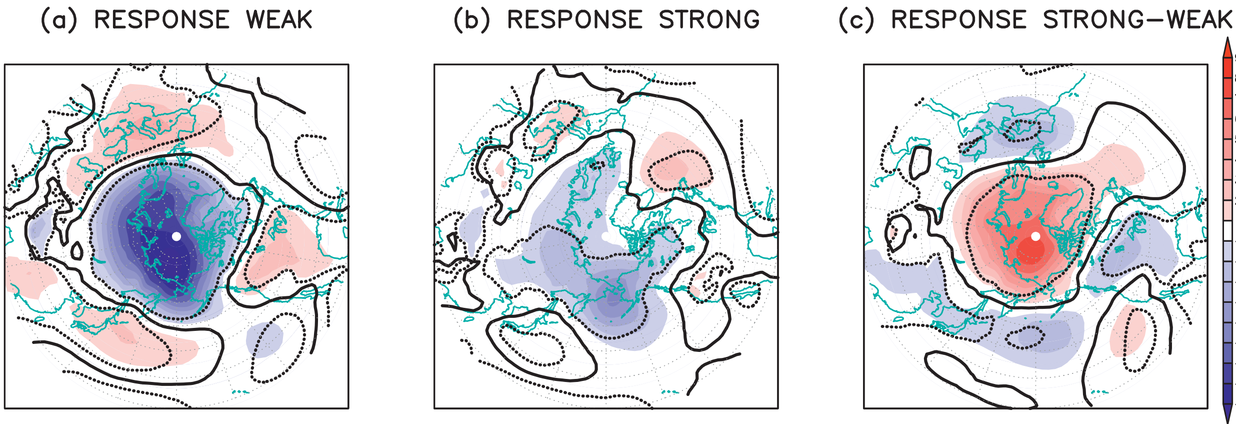

FIG. 1. The $2 \times CO_2$ response of the DJF MSLP in (a) the WEAK drag case, (b) the STRONG drag case, and (c) the difference between them (hPa). The black solid line denotes the zero contour line, and the dotted line denotes statistical significance at 95% level (according to a standard t test).

Because midlatitude weather is largely affected by the polarity of the NAM (e.g., Thompson and Wallace 2001), the difference between the WEAK and STRONG drag responses results in large differences in regional climate projections. In the WEAK drag case (Fig. 2a) the positive NAM response (with an increased meridional gradient of MSLP; see Fig. 1a) is consistent with stronger westerlies and thus more advection of mild air from the ocean, amplifying the warming over NH continental areas by up to 70% compared to the STRONG drag case (Fig. 2b) in which no NAM response was found. It is important to realize that, in both WEAK and STRONG cases, the same SST perturbation is prescribed. This artificially limits the response sensitivity of the near-surface climate, particularly over the ocean. The sensitivity to OGWD of the surface temperature response is therefore potentially larger in coupled atmosphere–ocean models. This will be reported in a future study.

For the case of WEAK and STRONG OGWD, Fig. 3 shows the zonal-mean zonal wind (contours) and OGWD momentum deposition due to the freely propagating gravity waves (shading) for $1 \times CO_2$ (Fig. 3, top), $2 \times CO_2$ (Fig. 3, middle), and their differences (Fig. 3, bottom), which by definition is the response to climate change. The primary feature to be understood is the circulation response difference displayed in Fig. 3i (contours) resulting from the use of WEAK versus STRONG OGWD settings. This response difference is not restricted to the surface (Fig. 1c), but extends all the way up to the stratosphere where the largest differences are found. In particular, for WEAK OGWD the NH polar vortex strengthens in response to CO_2 doubling, whereas for STRONG OGWD it weakens.

It is apparent that the climate change response for the WEAK and STRONG OGWD settings are substantially different. Both WEAK and STRONG settings arguably yield reasonable present-day simulations—the STRONG having the smaller MSLP biases and the WEAK with improved temperatures in the polar lower stratosphere at the expense of increased MSLP biases (Scinocca et al. 2008). It is not obvious, therefore, which of the two OGWD settings yield the more credible response to climate change, leaving a large uncertainty for regional climate projections in the NH extratropical winter. More insight into the mechanisms governing this sensitivity would be helpful in deciding which response is most credible.

To better understand the role of OGWD in this sensitivity we first focus on the impact of the WEAK versus STRONG settings at constant CO_2 forcing (i.e., along the top and middle panels of Fig. 3). In each of Figs. 3a,b,d,e, a region with large OGWD can be found on the upper and poleward flank of the subtropical jet near 100 hPa at 40°N. In this region upward propagating gravity waves have rapidly increasing amplitudes owing to the combined effects of increasing static stability, decreasing background wind, and decreasing density with height. Qualitatively, the same behavior is seen at $1 \times CO_2$ (Fig. 3, top) and $2 \times CO_2$ (Fig. 3, middle): relative to the WEAK settings, the STRONG settings increase the amount of OGWD deposited in this region (shading, Figs. 3c,f). The question then becomes, why does the increased OGWD in this region cause the wind responses displayed in Figs. 3c,f?

The dynamics underlying the wind response to the OGWD perturbation patterns displayed in Figs. 3c,f would seem to be explained by the study of Chen and Zurita-Gotor (2008, hereafter CZ08). There, the zonal wind response to OGWD in a comprehensive AGCM was investigated by a series of idealized simulations in

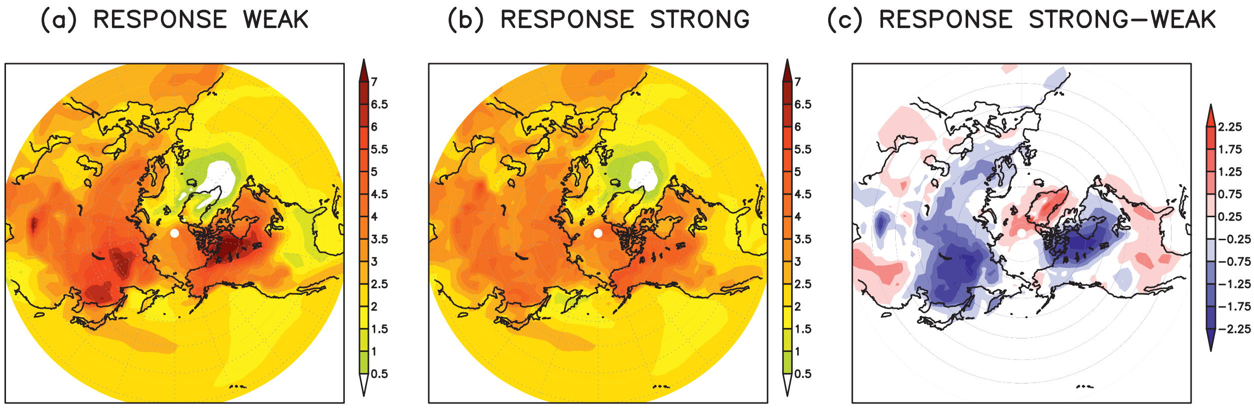

FIG. 2. As in Fig. 1 but for the surface temperature response (K).

which a zonal-mean torque was prescribed at various locations (latitude and height) relative to the jet core. For a westward torque located on the upward and poleward flank of the subtropical jet, as found in Figs. 3c,f, the CZ08 study predicts a stratospheric zonal wind deceleration and equatorward shift of the tropospheric jet. Detailed analysis in that study revealed that the full response required both the zonal-mean torque and the action of tropospheric and stratospheric eddies. They argue that the dynamical mechanism involves alteration of the eastward propagation speed of tropospheric eddies by the stratospheric zonal wind changes. The analysis of CZ08, therefore, would seem to provide some explanation for the dynamics responsible for the response sensitivity to WEAK versus STRONG OGWD settings at constant CO_2.

The focus of the present study involves the dynamics responsible for the response sensitivity to WEAK versus STRONG OGWD settings for changing CO_2 (i.e., the difference between Figs. 3c and 3f indicated in Fig. 3i). The central question is encapsulated by Fig. 3i: what is the role of differences in the OGWD forcing response between the WEAK and STRONG settings (shading) in determining the difference in the circulation response (contours)? In addressing this question, it would seem that the study of CZ08 may also provide some guidance. Closer inspection of Figs. 3c and 3f reveals that, in going from 1 to 2 $\times$ CO_2, the OGWD perturbation on the upper flank of the subtropical jet shifts to higher elevation. This is indicated by the dipole pattern of OGWD in Fig. 3i. Appealing to the analysis of CZ08, they also found that an upward shift of upper-level drag resulted in a stronger stratospheric wind response due to the decreased density, which in turn induced a stronger tropospheric response. This is consistent with what is found here for the 2 $\times$ CO_2 response. It is possible then that the same dynamics responsible for the response sensitivity to OGWD for constant GHG forcing is responsible for the response sensitivity to OGWD for changing GHG forcing. In other words, the results of CZ08 suggest that differences in the OGWD response between WEAK and STRONG cases (shading, Fig. 3i) could directly cause the differences in the circulation response (contours, Fig. 3i). The circulation response would then directly depend on the OGWD response, which in turn could directly depend on details of the OGWD parameterization scheme used. As stated in the introduction, this would have far-reaching implications for the importance of OGWD in climate change studies.

An alternative possibility, however, is that the dipole pattern of OGWD forcing in Fig. 3i is irrelevant to the circulation response sensitivity and that it simply results from changes in the background wind due to changes in GHG forcing. Owing to the increased meridional thermal gradient associated with CO_2-induced radiative cooling of the stratosphere and radiative heating of the troposphere, the subtropical jet strengthens and shifts upward (contours, Figs. 3g and 3h). The increased background stratospheric winds results in reduced gravity wave amplitudes, allowing these waves to reach higher elevation before breaking, resulting in a dipole pattern. [This dipole pattern was discussed by McLandress and Shepherd (2009) in the context of its influence on the Brewer–Dobson circulation (BDC).] From this perspective, the dipole pattern of OGWD would thus be a consequence, rather than the cause, of the zonal-mean circulation response to CO_2 doubling. Consequently, the difference in the OGWD response (Fig. 3i, shading) would be irrelevant to the difference in the zonal-mean circulation response (Fig. 3i, contours). The influence of the OGWD on the circulation response would thus be indirect. Although this and the preceding explanation following CZ08 are both reasonable, Fig. 3 does not provide the causal information required to distinguish

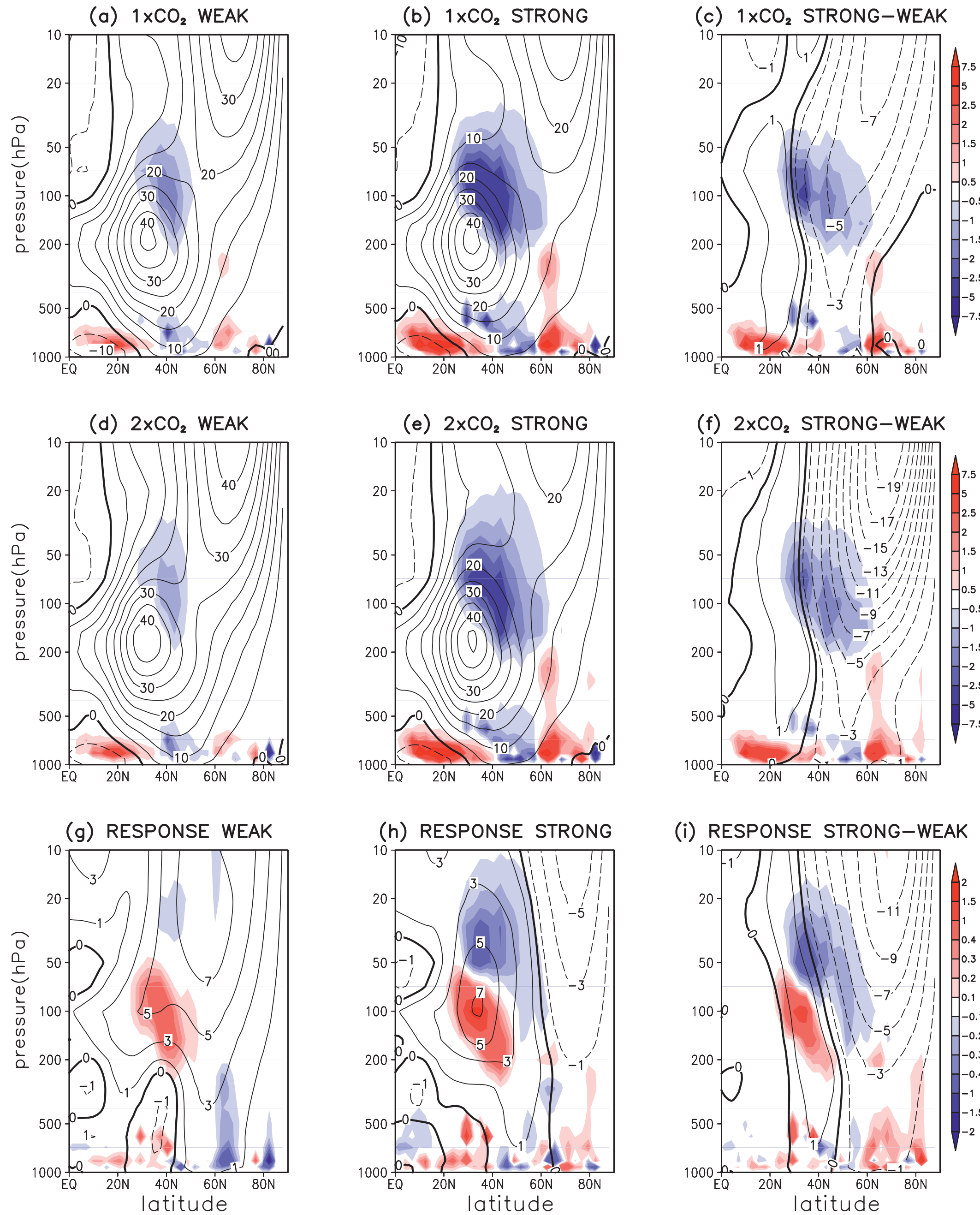

FIG. 3. Zonal-mean zonal wind (contours) and OGWD momentum deposition due to freely propagating gravity waves (shading, 10^6 Pa m^{-1}) in NH winter (DJF) for (top) the $1 \times CO_2$, (middle) the $2 \times CO_2$ runs, and (bottom) the response to climate change ($2 \times CO_2$ minus $1 \times CO_2$) for (left) the WEAK drag case, (center) the STRONG drag case, and (right) its difference (STRONG minus WEAK). Contour interval of the zonal wind is 5 m s^{-1} in (a),(b),(d),(e). The contour lines are at ±1, ±3, ±5 m s^{-1}, etc. in (c),(f)–(i). The solid line denotes the zero zonal wind contour line.

between the two. In the next section, we attempt to definitively answer whether the OGWD differences displayed in Fig. 3i directly affect the circulation response by employing prescribed OGWD forcing in these experiments.

b. *Prescribed OGWD experiments*

The analysis of the previous subsection established the dependence of the NH extratropical wintertime circulation response to WEAK and STRONG settings of

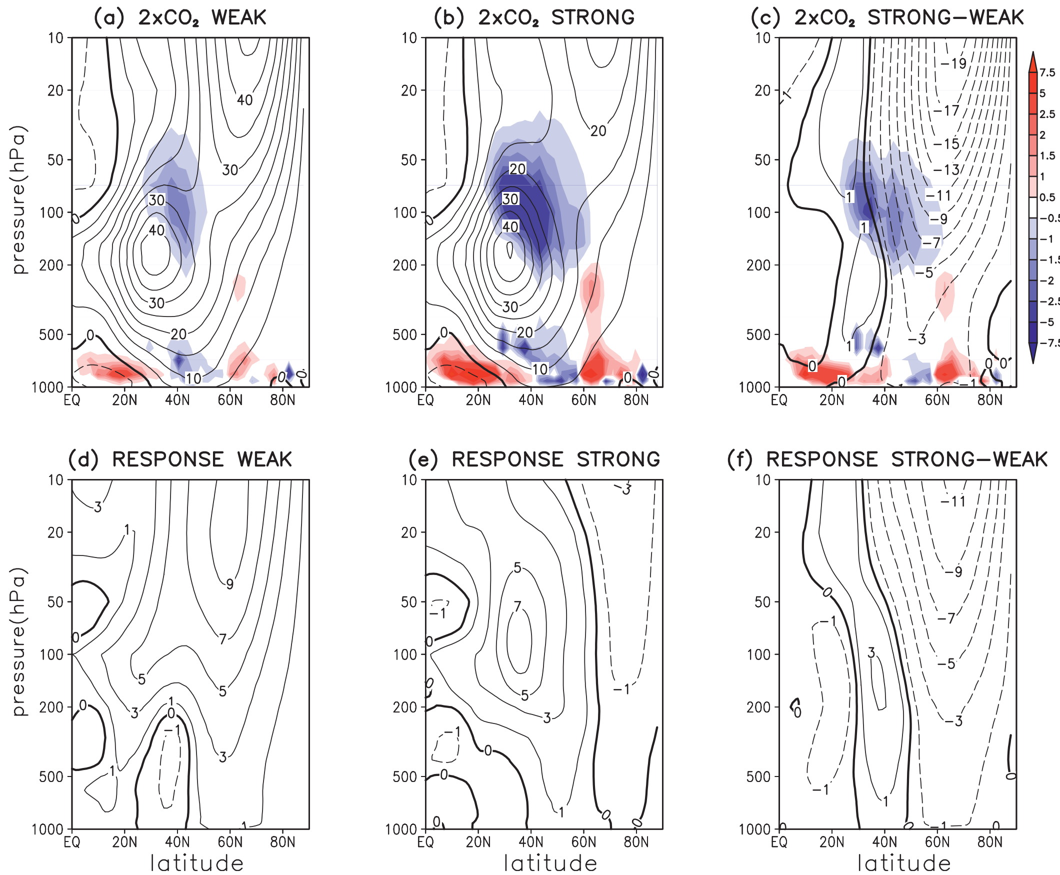

FIG. 4. Zonal-mean zonal wind (contours, m s^{-1}) and OGWD momentum deposition due to freely propagating gravity waves (shading, 10^6 Pa m^{-1}) in NH winter (DJF) in $2 \times CO_2$ runs with OGWD fixed to $1 \times CO_2$ values. (top) The zonal wind and (prescribed) OGWD deposition in the $2 \times CO_2$ simulations and (bottom) the response to climate change [which is the difference between Fig. 3 (top) and Fig. 4 (top)].

the OGWD under CO_2 doubling. As discussed, it is not yet clear whether the specific mechanism by which OGWD induces this sensitivity is due to a direct or an indirect influence of the OGWD. In this section, we specifically address this issue by employing prescribed OGWD (from the $1 \times CO_2$ simulations) in the $2 \times CO_2$ runs. In this way, the OGWD is held fixed and its response to CO_2 doubling in the WEAK and STRONG cases is forced to be identical (zero) by construction.

The top row of Fig. 4 presents the new $2 \times CO_2$ simulations with prescribed $1 \times CO_2$ OGWD forcing for the WEAK and STRONG cases. The top row of Fig. 4 should be compared to the second row of Fig. 3: both have the same $1 \times CO_2$ simulations as controls (Fig. 3, top). The bottom row of Fig. 4 shows the response to CO_2 doubling for the WEAK and STRONG cases (Figs. 4d and 4e) and their difference (Fig. 4f). The absence of shading in these panels indicates that, by construction, no changes occur in gravity wave drag in the production of the wind response. The main result of these new experiments derives from the comparison of the bottom row of Fig. 4 with the bottom row of Fig. 3. The remarkable result is that the $2 \times CO_2$ zonal-mean zonal wind responses in the WEAK and STRONG cases remain essentially unchanged when the OGWD is held fixed during CO_2 doubling. In other words, changes in the OGWD forcing under CO_2 doubling (i.e., the shaded fields in Figs. 3g,h) do not play an active role in shaping the circulation response. Consequently, the sensitivity of the $2 \times CO_2$ zonal wind response (Fig. 4f) is also unchanged when OGWD is held fixed during CO_2 doubling.

The conclusion that OGWD does not play a direct role in forcing the NH wintertime circulation response is further supported by Fig. 5. Even though OGWD was imposed in these new $2 \times CO_2$ simulations, we continued to run the OGWD scheme in an offline mode. We will refer to this output of the scheme run in this mode as the "offline OGWD." In general, the offline OGWD in the $2 \times CO_2$ runs will not be equal to the $1 \times CO_2$

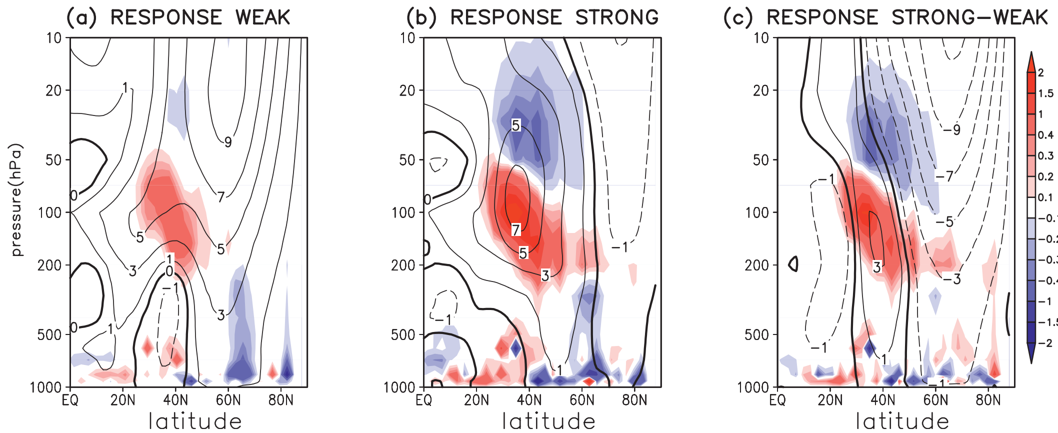

FIG. 5. Zonal-mean zonal wind response (contours, m s^{-1}) and the offline OGWD momentum deposition response due to freely propagating gravity waves (shading, 10^6 Pa m^{-1}) in NH winter (DJF) in 2 × CO_2 runs with OGWD fixed to 1 × CO_2 values (details in text).

OGWD imposed in these simulations. Shading in Fig. 5 represents the difference between the offline OGWD of the new 2 × CO_2 runs and the OGWD in the control simulations (Fig. 3, top). This response can be regarded as the passive OGWD response to the zonal wind changes under CO_2 doubling (shown by the contours, copied from the bottom row of Fig. 4). The passive OGWD response in Fig. 5 can be compared directly with the interactive OGWD response displayed in the bottom row of Fig. 3. Their structures are nearly identical. This confirms that the change in OGWD forcing under CO_2 doubling in the original interactive runs (Fig. 3) is not the cause but, rather, the result of the circulation response to CO_2 forcing. This result validates the interpretation of the previous subsection that the OGWD dipole response pattern in Fig. 3i is a consequence of basic-state changes due to CO_2 doubling. Consequently, although the CZ08 mechanism seems to explain the dynamics responsible for the response to WEAK versus STRONG OGWD settings at constant CO_2, it does not seem to be relevant for the sensitivity of the response to OGWD for changing CO_2.

The conclusion then is that differences in the response of OGWD to CO_2 forcing (shading in Fig. 3i) play no role in causing the sensitivity of the circulation response to the WEAK and STRONG settings (contours in Figs. 3i and 4f); that is, OGWD does not directly influence the NH wintertime zonal-mean zonal wind response to CO_2 forcing. The primary mechanism that explains the sensitivity would appear to involve the action of the resolved waves, which presumably differs between the WEAK and STRONG cases owing to basic-state differences of the 1 × CO_2 control climates to which the CO_2 forcing is applied. This will be investigated in the next subsection.

c. *Resolved waves and the impact of the basic state*

We are still left with the question of how exactly the strength of OGWD influences the NAM response to CO_2 forcing. In eliminating the possibility of a direct influence of OGWD on the circulation response, the previous subsection points to the key role of resolved waves. In this section we will investigate the role of resolved wave driving to understand how the difference in the resolved wave responses might be explained by differences in the refractive properties of the WEAK and STRONG OGWD basic states.

As discussed in section 3a, WEAK (STRONG) OGWD settings lead to strong (weak) zonal-mean zonal winds in the lower stratosphere. The planetary wave refractive properties of these different background flows can be characterized by the square of the quasigeostrophic refractive index R^2 (Matsuno 1970). Where R^2 is negative, wave propagation is generally inhibited. The impact of such basic-state refractive properties on planetary wave propagation may be visualized and quantified by the Eliassen–Palm (EP) flux. Here we present the EP flux derived from the primitive equations as given by Eq. (3.5.3) and R^2 as given by Eq. (5.3.7) of Andrews et al. (1987). In Fig. 6, both R^2 for wavenumber zero and the EP flux are presented for the 1 × CO_2 and 2 × CO_2 basic states of the WEAK and STRONG cases in the same nine-panel layout presented earlier for Fig. 3. In Fig. 6, regions of $R^2 \leq 0$ have been shaded black to indicate the location of potential barriers to wave propagation, and EP flux vectors (green arrows) have been divided by density. In addition, an EP flux budget [following Kushner and Polvani (2004), Eq. (7)] over a control region, or box, bounded by 100 and 10 hPa and 45°–90°N is provided for the difference/response plots. The red

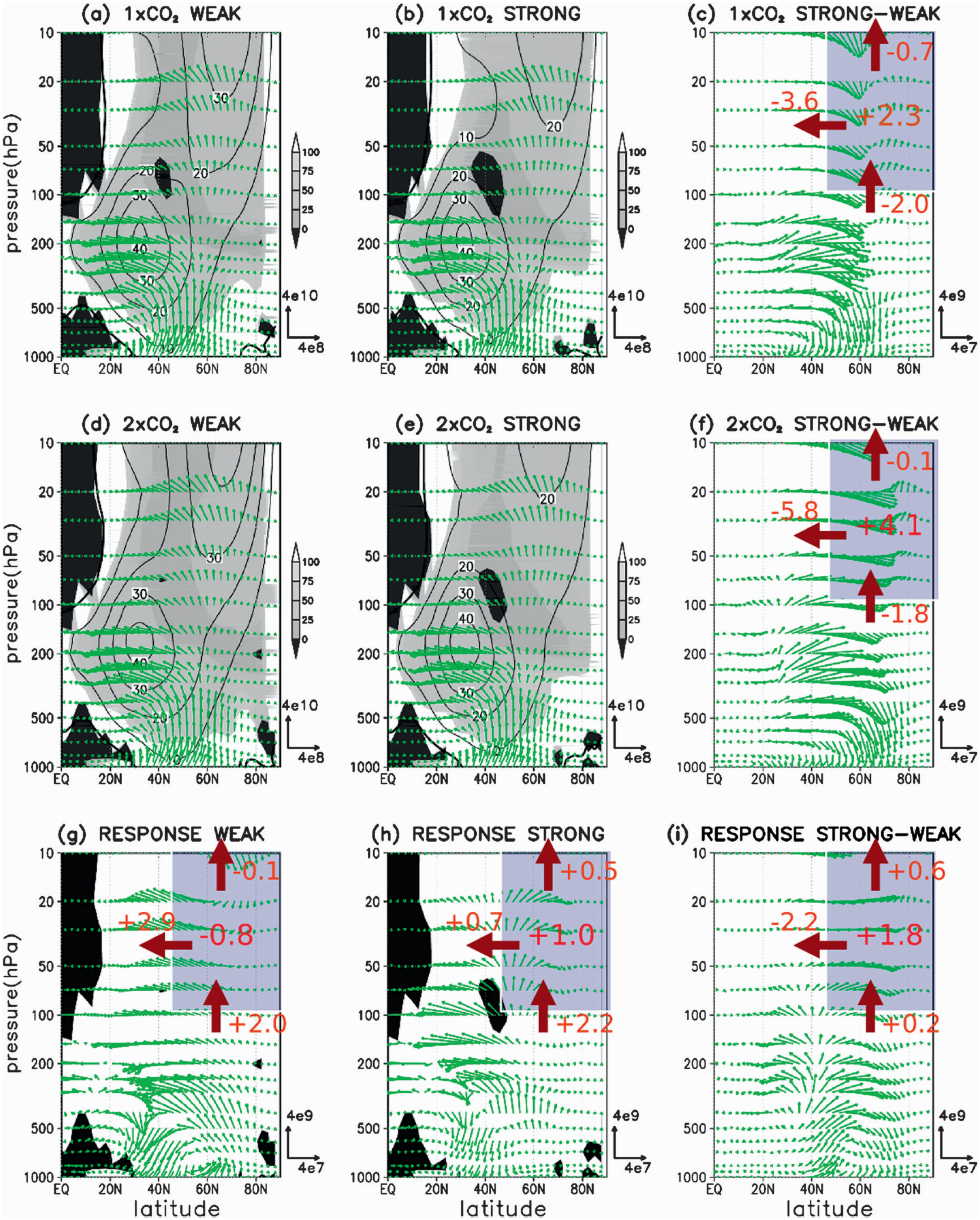

FIG. 6. As in Fig. 3 but for resolved wave quantities. (a),(b),(d),(e) Contours represent zonal-mean zonal wind (m s^{-1}, copied from Fig. 3), and shading represents the quasigeostrophic refractive index squared R^2 (for zonal wavenumber zero and multiplied by the square of the earth's radius). (g),(h) Black shading indicates regions of $R^2 \leq 0$ for the $2 \times CO_2$ simulations. EP flux vectors are represented by the green arrows (scale at bottom right of each panel, kg s^{-2}) and have been divided by density. (c),(f)–(i) A budget for resolved wave driving is presented for the stratospheric box between 100 and 10 hPa and between 45° and 90°N. Red numbers across the box represent differences/responses of EP fluxes integrated over the box boundaries, and the red numbers in the box represent the differences/responses of the resolved wave driving integrated over the box (10^4 kg m s^{-4}).

arrows (and associated numbers) across the sides of the box represent the integrated EP flux difference through the boundaries of the box, whereas the red numbers in the box represent the time- and area-mean momentum deposition, or EP flux divergence (EPFD), associated with the resolved waves (all wavenumbers). We focus on the high-latitude stratospheric EP flux budget since this is the region where the difference between the WEAK and STRONG circulation response maximizes (Fig. 3i).

Focusing first on the $1 \times CO_2$ basic states (Fig. 6, top), it is seen that both WEAK and STRONG cases have a region of negative R^2 centered near 40°N at 70 hPa. This region is significantly larger in the STRONG case than in the WEAK case. An analysis of the relative magnitude of terms that make up the index of refraction in the WEAK and STRONG cases indicates that the larger region of negative R^2 in the STRONG case is related to a smaller meridional gradient of potential vorticity resulting from flow curvature changes associated with its weaker lower-stratospheric winds. The larger region of $R^2 \leq 0$ in the STRONG case suggests that more of a barrier to the equatorward propagation of planetary waves exists in the STRONG case compared to the WEAK case. This is supported by the EP flux budget in Fig. 6c: the STRONG case has a 3.6×10^4 kg m s^{-4} smaller equatorward flux across the 45°N control surface, which is 33% less equatorward flux than occurs for the WEAK case. Note that, in the STRONG case, the incoming planetary wave flux through the 100-hPa boundary is 2.0×10^4 kg m s^{-4} smaller than in the WEAK case. This does not result in smaller planetary wave driving averaged over the control box. In fact, the planetary wave driving is 2.3×10^4 kg m s^{-4} more than in the WEAK case, which can be ascribed to the smaller outflux of momentum across the 45°N control boundary and, to a lesser extent, to the smaller outflux across the 10-hPa upper boundary.

When CO_2 is doubled (Fig. 6, middle), the subtropical jet strengthens, increasing the meridional gradient of potential vorticity, which results in larger values of R^2. In the STRONG OGWD case these changes do not result in a significant decrease of the size of the region with $R^2 \leq 0$. A significant barrier to equatorward propagation still remains in the STRONG OGWD $2 \times CO_2$ climate. For the WEAK OGWD, however, the region of negative R^2 almost disappears (Fig. 6d), and an increased equatorward propagation of planetary wave activity is expected relative to its $1 \times CO_2$ control and to the $2 \times CO_2$ STRONG case. This is verified by the EP flux budgets: The equatorward flux across 45°N in the WEAK $2 \times CO_2$ climate is 2.9×10^4 kg m s^{-4} (27%) larger than in the WEAK $1 \times CO_2$ climate (Fig. 6g) and 5.8×10^4 kg m s^{-4} (73%) larger than in the STRONG $2 \times CO_2$ climate (Fig. 6f). Recalling from the previous section that changes in the OGWD have no impact on the $2 \times CO_2$ response, we may conclude that these relative differences in planetary wave propagation characteristics arise solely because of the different $1 \times CO_2$ control climates to which the CO_2 perturbation was applied.

The sensitivity of the circulation response to the strength of OGWD can be understood by considering the EP flux budgets in the last row of Fig. 6 (for reference, the black shading of regions with $R^2 \leq 0$ in the $2 \times CO_2$ simulations have been included in Figs. 6g and 6h). The integrated EPFD response in the control box is very different in the WEAK versus STRONG case. In response to CO_2 doubling, the amount of upward EP flux from the troposphere into the stratosphere across the 100-hPa control surface increases in both cases, which is a common feature in climate models (e.g., Butchart et al. 2006; McLandress and Shepherd 2009). The EP flux through the 10-hPa surface, on the other hand, remains essentially unchanged upon CO_2 doubling in both WEAK and STRONG cases. Taken together, this implies a vertical convergence of EP flux in the polar stratosphere. Because the barrier to equatorward propagation is essentially maintained upon CO_2 doubling for STRONG OGWD settings, the vertical convergence of EP flux in the control box results in a positive EPFD in this region (1.0×10^4 kg m s^{-4}). Even though an equivalent vertical convergence of EP flux in the stratosphere is realized upon CO_2 doubling in the WEAK case, the breakdown of the barrier to equatorward transport results in the opposite tendency of EPFD in the control box (the EPFD decreases by 0.8×10^4 kg m s^{-4}). The net result (Fig. 6i) is an EPFD response difference of 1.8×10^4 kg m s^{-4} in the control box between the STRONG and WEAK cases, which accounts for the weaker polar stratospheric winds in the STRONG versus WEAK response to CO_2 doubling. It can be argued that the downward extension of the circulation response difference (Fig. 3i) can be explained by a downward control response (Haynes et al. 1991) to the stratospheric EPFD differences, which may include feedbacks from tropospheric eddies (Song and Robinson 2004).

From this and the previous subsection, it is now clear that the action of resolved planetary waves and their sensitivity to the control basic state upon which the CO_2 perturbation is applied are the primary factors responsible for the sensitivity of the NAM response to the strength of OGWD. This result and its implications are discussed further in the next section.

4. Summary and discussion

Obtaining credible climate change projections in NH extratropical winter is challenging, as the current generation of coupled atmosphere–ocean models shows a wide range in the NAM response to increasing greenhouse gases. S08 showed that the NAM response is very sensitive to parameter settings of the orographic gravity wave drag scheme. A WEAK (STRONG) setting of the OGWD scheme produced a basic state with strong (weak) midlatitude winds in the wintertime lower stratosphere and a doubled CO_2 response that projected positively (neutrally) onto the NAM. This sensitivity implicitly suggests that the specific formulation and parameter settings of OGWD might directly influence the time-mean circulation response to greenhouse gas forcing. Another possibility is that the role of OGWD is limited to its influence on the basic state of the control simulation and that the sensitivity of the circulation response is due to the sensitivity of resolved planetary waves to the differing basic-state properties. If this were true, it would suggest that the circulation response to greenhouse gas forcing is indirectly influenced by OGWD and therefore relatively insensitive to the detailed formulation of the OGWD scheme used. Alternatively, the sensitivity of the circulation response to OGWD could arise from some combination of these direct and indirect mechanisms.

In the present study we have undertaken a detailed investigation of this issue. By developing a procedure to employ prescribed OGWD forcing, we were able to perform the $2 \times CO_2$ simulations with the OGWD fixed to its $1 \times CO_2$ values. In this way, for the WEAK and STRONG settings, we were able to determine the impact of differences in the OGWD forcing response on the circulation response when CO_2 is doubled. The results were conclusive. The OGWD forcing response to CO_2 doubling had essentially no impact on the NH wintertime zonal-mean zonal wind response. Stated another way, differences between the WEAK and STRONG OGWD forcing response to CO_2 doubling were not the cause but, rather, the result of the circulation response differences. The influence of OGWD is indirect and associated with determining differences in the planetary wave propagation characteristics of the control basic state.

An investigation of the resolved waves in this study clarified their central dynamical role in causing the different NAM responses in the WEAK and STRONG OGWD cases. For $1 \times CO_2$, a barrier to equatorward propagation of planetary waves occurs in the midlatitude lower stratosphere, which is larger for the STRONG OGWD settings. For $2 \times CO_2$, this barrier remains essentially unchanged in the STRONG case but is significantly reduced in the WEAK case. As a consequence, a dramatic increase in equatorward EP flux occurs in the WEAK simulation relative to both its $1 \times CO_2$ control and its $2 \times CO_2$ STRONG counterpart. The net result is a greater planetary wave driving of the mid- to high latitude stratosphere in the STRONG response compared to the WEAK response. This response difference in resolved wave driving accounts for the response difference in the lower-stratospheric zonal wind and, by downward control arguments (Haynes et al. 1991), that in the tropospheric circulation. Therefore, the action of resolved planetary waves is the primary dynamical mechanism that causes the NAM response to GHG forcing to be sensitive to the strength of OGWD.

While the present study has demonstrated the OGWD forcing response to CO_2 doubling (e.g., dipole patterns in Figs. 3g and 3h) has essentially no impact on the time-mean tropospheric circulation response, other components of the circulation response will almost certainly depend directly on the details of the CO_2-induced OGWD changes. For example, there is mounting evidence from climate change simulations employing comprehensive middle-atmosphere models that the Brewer–Dobson circulation (BDC) strengthens with increased GHG forcing (Butchart and Scaife 2001; Sigmond et al. 2004; Butchart et al. 2006; Fomichev et al. 2007; Garcia and Randel 2008; Li et al. 2008; McLandress and Shepherd 2009). A number of studies have considered the contribution of OGWD to this change in the BDC and found that the contribution is quite substantial (at least 40%, see Butchart et al. 2006; Li et al. 2008; McLandress and Shepherd 2009). The present study indicates that the OGWD forcing response that contributes to this change in the BDC depends directly on the settings of the OGWD; that is, the OGWD response difference displayed in Fig. 3i indicates that the WEAK and the STRONG settings will lead to a different contribution of OGWD to the changing BDC, which could lead to differences in the climate change response of the BDC. This issue will be the subject of a follow-on study.

The results of the prescribed forcing experiments point to something more fundamental than simply the role of OGWD in this problem. They explicitly identify differences in the control basic state upon which the CO_2 perturbation is applied as the source of the difference in the time-mean tropospheric circulation responses to climate change. The analysis presented here suggests that the property of a model's control climate that is most important to the NAM response is the zonal wind in the lower stratosphere. This would imply that any internal parameter setting or physical process that influences the circulation of the midlatitude lower stratosphere has the potential to influence the NAM response to climate change. For example, there is evidence that solar variability, like OGWD, also influences both the control basic state and the NAM response to climate change. This has been investigated by Kodera et al. (2008). Consistent with our results, they found that tropospheric circulation trends in winters with relatively weak (strong) stratospheric winds, corresponding to low (high) solar activity, projected positively (neutrally) on the NAM.

The realization that lower-stratospheric wind biases play such a pivotal role in determining the circulation response to climate change suggests that this property of a model's control climate might help explain the range of NAM responses found in existing climate projections. For example, until now, investigation of the sensitivity of the time-mean NH wintertime circulation response between Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (AR4) models has focused on the inclusion or exclusion of a well-resolved stratosphere (Miller et al. 2006). Our results would suggest that this intermodel response sensitivity might be better understood by compositing models based on the wintertime midlatitude lower-stratospheric wind biases of their control climates. Such an analysis is underway, and results will be reported in the near future.

The tacit assumption of climate change science is that, even if two models have biases relative to each other (and to observations), their perturbative response to modest external forcing (such as increased GHGs) should be independent of these biases: that is, it is assumed that each model's biases are insensitive to the external forcing and therefore will cancel out in the response. In many instances this assumption works quite well. The present study provides a clear counter example.

One of the open questions of climate science is: how can we validate future projections of climate change? Here we have identified that the NAM response is strongly sensitive to the wintertime midlatitude lower-stratospheric wind biases of the model's control climate. The physical reality of the NAM response would, therefore, seem to hinge on a strategy of reducing the wind biases with respect to observations in this region. However, wind biases and mean sea level pressure biases in this model are difficult to simultaneously eliminate (e.g., Scinocca and McFarlane 2000; Scinocca et al. 2008; section 3.3), and the reduction of wind biases is associated with a larger MSLP bias. The preferential reduction of lower-stratospheric wind biases then implicitly assumes that these wind biases are potentially more important than MSLP biases.

The idea that some climate biases might be inherently more important than others would seem to be a useful concept (cf. Sigmond et al. 2007). Knowledge of which biases are most critical to projections of climate change would be enormously useful during the development cycle of a GCM. However, it is important to realize that the importance of the wind bias illustrated in the present study is with respect to only one aspect of climate change: the time-mean NAM response during NH winter. It may well be that the MSLP biases, suffered in using the WEAK relative to STRONG OGWD settings, could result in climate change response sensitivities in other seasons or for other physical processes of the climate system. This issue remains to be investigated.

Finally, we have only investigated the role of OGWD in shaping the time-mean NAM response to increasing GHGs. This was to address the outstanding issues raised in S08. The S08 and present studies have not considered the impact of OGWD and the stratosphere on the response of interannual and intraannual variability to GHG forcing. This is left to future investigation.

Acknowledgments. The authors thank Lorenzo Polvani, Judith Perlwitz, Charles McLandress, and Nathan Gillett for constructive comments. MS and JFS are members of the CSPARC network and gratefully acknowledge the support of the Canadian Foundation for Climate and Atmospheric Sciences.

REFERENCES

Andrews, D. G., J. Holton, and C. Leovy, 1987: *Middle Atmosphere Dynamics*. Academic Press, 489 pp.

Baldwin, M. P., and T. Dunkerton, 2001: Stratospheric harbingers of anomalous weather regimes. *Science,* **244,** 581–584.

Butchart, N., and A. A. Scaife, 2001: Removal of chlorofluorocarbons by increased mass exchange between the stratosphere and troposphere in a changing climate. *Nature,* **410,** 799–802.

——, and Coauthors, 2006: Simulations of anthropogenic change in the strength of the Brewer-Dobson circulation. *Climate Dyn.,* **27,** 727–741, doi:10.1007/s00382-006-0162-4.

Chen, G., and P. Zurita-Gotor, 2008: The tropospheric jet response to prescribed zonal forcing in an idealized atmospheric model. *J. Atmos. Sci.,* **65,** 2254–2271.

Fomichev, V., A. Jonsson, J. de Grandpré, S. Beagley, C. McLandress, K. Semeniuk, and T. Shepherd, 2007: Response of the middle atmosphere to CO_2 doubling: Results from the Canadian Middle Atmosphere Model. *J. Climate,* **20,** 1121–1144.

Garcia, R. R., and W. J. Randel, 2008: Acceleration of the Brewer–Dobson circulation due to increases in greenhouse gases. *J. Atmos. Sci.,* **65,** 2731–2739.

Haynes, P., C. J. Marks, M. McIntyre, T. Shepherd, and K. Shine, 1991: On the "downward control" of extratropical diabatic circulations by eddy-induced mean zonal forces. *J. Atmos. Sci.,* **48,** 651–678.

Kodera, K., M. E. Hori, S. Yukimoto, and M. Sigmond, 2008: Solar modulation of the Northern Hemisphere winter trends and its implications with increasing CO_2. *Geophys. Res. Lett.,* **35,** L03704, doi:10.1029/2007GL031958.

Kushner, P. J., and L. M. Polvani, 2004: Stratosphere–troposphere coupling in a relatively simple AGCM: The role of eddies. *J. Climate,* **17,** 629–639.

Li, F., J. Austin, and J. Wilson, 2008: The strength of the Brewer–Dobson circulation in a changing climate: Coupled chemistry–climate model simulations. *J. Climate,* **21,** 40–57.

Matsuno, T., 1970: Vertical propagation of stationary planetary waves in the winter Northern Hemisphere. *J. Atmos. Sci.,* **27,** 871–883.

McLandress, C., and T. G. Shepherd, 2009: Simulated anthropogenic changes in the Brewer–Dobson circulation, including its extension to high latitudes. *J. Climate,* **22,** 1516–1540.

Miller, R. L., G. A. Schmidt, and D. T. Shindell, 2006: Forced annular variations in the 20th century Intergovernmental Panel on Climate Change Fourth Assessment Report models. *J. Geophys. Res.,* **111,** D18101, doi:10.1029/2005JD006323.

Scinocca, J. F., and N. A. McFarlane, 2000: The parameterization of drag induced by stratified flow over anisotropic orography. *Quart. J. Roy. Meteor. Soc.,* **126,** 2353–2393.

——, ——, M. Lazare, J. Li, and D. Plummer, 2008: Technical note: The CCCma third generation AGCM and its extension into the middle atmosphere. *Atmos. Chem. Phys.,* **8,** 7055–7074.

Shindell, D. T., R. L. Miller, G. A. Schmidt, and L. Pandolfo, 1999: Simulation of recent northern winter climate trends by greenhouse-gas forcing. *Nature,* **399,** 452–455.

Sigmond, M., P. C. Siegmund, E. Manzini, and H. Kelder, 2004: A simulation of the separate effects of middle-atmospheric and tropospheric CO_2 doubling. *J. Climate,* **17,** 2352–2367.

——, P. J. Kushner, and J. F. Scinocca, 2007: Discriminating robust and non-robust atmospheric circulation responses to global warming. *J. Geophys. Res.,* **112,** D20121, doi:10.1029/2006JD008270.

——, J. F. Scinocca, and P. J. Kushner, 2008: The impact of the stratosphere on tropospheric climate change. *Geophys. Res. Lett.,* **35,** L12706, doi:10.1029/2008GL033573.

Solomon, S., D. Qin, M. Manning, M. Marquis, K. Averyt, M. M. B. Tignor, H. L. Miller Jr., and Z. Chen, Eds., 2007: *Climate Change 2007: The Physical Science Basis*. Cambridge University Press, 996 pp.

Song, Y., and W. A. Robinson, 2004: Dynamical mechanisms for stratospheric influences on the troposphere. *J. Atmos. Sci.,* **61,** 1711–1725.

Thompson, D. W. J., and J. M. Wallace, 2000: Annular modes in the extratropical circulation. Part I: Month-to-month variability. *J. Climate,* **13,** 1000–1016.

——, and ——, 2001: Regional climate impacts of the Northern Hemisphere annular mode. *Science,* **293,** 85–89.