NOTES AND CORRESPONDENCE

Atmospheric Modes of Variability in a Changing Climate

JENNY BRANDEFELT

Department of Meteorology, Stockholm University, Stockholm, Sweden

(Manuscript received and in final form 18 October 2005)

ABSTRACT

The response of the atmospheric circulation to an enhanced radiative greenhouse gas forcing in a transient integration with a coupled global climate model is investigated. The spatial patterns of the leading modes of Northern Hemisphere atmospheric variability are shown to change in response to the enhanced forcing. An earlier study showed that the spatial patterns of the leading modes in the Southern Hemisphere changed in response to the enhanced forcing. These changes were associated with changes in the propagation conditions for barotropic Rossby waves. This is, however, not the case for the Northern Hemisphere, where the propagation conditions are unchanged. Other possible mechanisms for the changes in the spatial patterns of the leading modes are discussed.

1. Introduction

In a recent article, Brandefelt and Källén (2004, hereafter BK04) study the response of the Southern Hemisphere (SH) atmospheric circulation to an enhanced radiative greenhouse gas (GHG) forcing. They find that the spatial pattern of the leading mode of variability, the so called Pacific–South American (Mo and Ghil 1987; Lau et al. 1994) mode, changes in response to the enhanced GHG forcing in a transient integration with a coupled global climate model (CGCM). Furthermore, they find that this change is associated with changes in the propagation conditions for barotropic Rossby waves. In the present article the effect of the enhanced GHG forcing on the Northern Hemisphere (NH) leading modes of variability is studied.

The enhanced GHG forcing affects the mean state and the variability of the climate system. The variability is physically coupled to the mean state and may respond also to changes in the mean state. It has been proposed that the mean response of the climate system to an enhanced forcing projects directly onto the pre-existing natural modes of variability (Palmer 1999). This hypothesis implies that the modes of variability do not themselves change in response to the forcing. The alternative to this hypothesis of unchanged flow regimes is that the spatial structure of the flow patterns is altered in response to the forcing. BK04 find that the mean response is not only associated with changes in the frequency of occurrence of the leading modes of variability, but also with changes in the spatial patterns of the modes.

Most CGCMs predict an increasing global mean surface temperature in response to the enhanced GHG forcing (Houghton et al. 2001). However, the strength and the spatial distribution of this heating differs substantially among CGCM simulations. Differences in the model formulation (e.g., in cloud parameterizations) are a possible explanation for the large differences in the response. Another possible explanation is the large natural variability of the climate system. Selten et al. (2004) show results from a large ensemble of CGCM simulations that suggest that the observed strengthening of the westerly winds over the North Atlantic during the past decades is not due to the enhanced greenhouse effect. They find the strengthening to be an expression of a random internal climate variation driven by increased precipitation over the tropical Indian Ocean.

BK04 find that the SH response in the mean and variability is associated with changes in the mean...
propagation conditions for barotropic Rossby waves. Following BK04 we investigate a possible connection between the NH response and changes in the propagation conditions for barotropic Rossby waves. The propagation conditions for barotropic Rossby waves are determined by the horizontal wind field, or to be more precise by the curvature (Laplacian) of the wind profile. Because of the dependence on the second-order derivatives of the wind components, small differences in the wind field (e.g., between different CGCM simulations) may be associated with substantial differences in the propagation conditions. AchutaRao et al. (2004) compare the performance of 11 CGCMs, for both unperturbed “control” simulations and for transient integrations with a 1% per year increase in CO₂. They find a spread in the zonal mean wind among the control simulations on the order of 5–10 ms⁻¹. Inter-model differences in the propagation conditions are expected because of this spread, but also because of differences in the simulated response to the enhanced GHG forcing. A common signature in CGCM simulations of the response to the enhanced GHG forcing is an increase in the zonal mean upper-level wind (Räisänen 2003). The differences among present-day simulations are of the same order as this response to enhanced GHG forcing (Räisänen 2003).

A possible explanation for intermodel differences in the upper-tropospheric–lower-stratospheric winds is changes in the modeled tropopause height and slope. Increases in tropospheric GHG concentrations and decreases in stratospheric ozone are expected to result in a lifting of the tropopause during the twenty-first century (Santer et al. 2003). Increases in tropopause height are also identified in observations over the past decade (Hoinka 1998; Seidel et al. 2001; Randel et al. 2000; Highwood et al. 2000). Because clouds play an important role in determining the radiative balance at the tropopause and thus affect the thermal structure of the tropopause, the description of clouds in CGCMs is important for the modeled tropopause. Iacobellis et al. (2003) find that the surface and top-of-atmosphere fluxes are sensitive to the scheme used to specify the ice particle effective radius. They conclude that the seemingly modest flux sensitivities may have important implications for numerical climate simulations.

The focus of the present study is the effect of an increasing GHG forcing on the mean state and on the variability of the large-scale atmospheric circulation. We determine the mean response to an enhanced GHG forcing in a specific transient CGCM simulation with increasing GHG forcing. The leading modes of variability are also determined for consecutive periods of this transient integration. Changes in the leading modes of variability in response to the enhanced GHG forcing are evaluated in connection to the ideas of unchanged flow regimes (Palmer 1999). The time evolution of the leading modes is also determined to assess the robustness of the changes as compared to the natural variability. The spatial patterns of the leading modes are shown to change in response to the enhanced GHG forcing. Following BK04 we investigate a possible connection between changes in the propagation conditions for barotropic Rossby waves and these changes in the spatial patterns. The approach used by BK04 for the SH is based on the assumption that the large-scale circulation is basically zonally symmetric. This assumption is not valid for the NH. For the present study it is therefore replaced by the assumption that large-scale zonal asymmetries in the circulation are of importance for the propagation of barotropic Rossby waves.

The outline of the article is as follows. The global coupled model and the methods used to assess the atmospheric variability in this model are described in section 2. The modeled response of the Northern Hemisphere atmospheric circulation to increasing atmospheric GHG concentrations is described in section 3. The mean response and changes in the leading modes of variability are discussed. The propagation conditions for barotropic Rossby waves are determined in section 4. A possible connection between the changes in the spatial patterns of the leading modes of variability and changes in the propagation conditions is tested. Finally, the study is summarized in section 5.

2. Model, data, and methods

a. Model and data

Data from the global coupled ocean–atmosphere–sea ice–land surface climate model ECHAM4/OYP/C3 are used. The atmospheric part of this model has a spectral resolution of T42 in the horizontal dimension and 19 vertical layers. The coupling involves annual mean flux correction, restricted to heat and freshwater fluxes, in order to avoid climate drift. A 240-yr simulation of transient greenhouse warming is used. This simulation starts from equilibrium conditions with the ocean model spun up using present-day GHG concentrations (Roeckner et al. 1999). The atmospheric GHG forcing is prescribed using observations for the period from 1860 until 1990. Thereafter, the Intergovernmental Panel on Climate Change IS92a scenario for radiative GHG forcing is used (Houghton et al. 1992). The effects of sulfate aerosols and tropospheric ozone are not taken into account. Further details on the model and design of simulations used in this study are given in Roeckner et al. (1999). Also, Hu et al. (2001) describe
the impact of global warming on the interannual and interdecadal climate modes in this model for the region 20°S–90°N.

Daily data are used in this study. The spatial domain is from 20°N to the North Pole. Periods of 30 yr from the transient integration (1860–89, 1890–1919, . . ., 2070–99) are considered. The analysis is performed on anomalies defined with respect to the seasonal cycle. The seasonal cycle at each grid point is obtained using harmonic analysis (retaining the first and second harmonic) with a Lanczos window. Prior to the harmonic analysis, the grand mean and linear trend are removed.

b. Methods

The leading modes of interweekly variability (IWV) are determined using empirical orthogonal function (EOF) analysis (i.e., principal component analysis) of daily low-pass-filtered (periods greater than 10 days) data. The daily anomalies (see section 2a) are filtered using a low-pass filter with a Lanczos window (Hanning 1989) with a cutoff period of 10 days. EOFs are determined for NH winter (December–February). Prior to the EOF analysis the NH winter mean anomaly at each grid point is removed. Area weights, reflecting the area represented by each grid point, are used in computing the covariance matrix. To determine if the EOFs are clearly separated, the criterion of North et al. (1982) is used.

3. The coupled model response to an enhanced GHG forcing

The increase in atmospheric GHG concentrations is negligible during the first 30 yr of the transient integration (years 1860–89). This period is therefore used to represent present-day climate. The response to the enhanced GHG forcing is defined as the difference between the last and the first 30 yr of the transient integration (years 2070–99 minus 1860–89).

The response to the enhanced GHG forcing is described in the following in terms of (i) changes in the 30-yr average surface air temperature $T_s$, mean sea level pressure (MSLP), and geopotential height at 500 hPa ($H_{500}$) and at 200 hPa ($H_{200}$); and (ii) changes in the spatial patterns of the leading modes of IWV.

a. The mean response

The direct effect of the enhanced radiative GHG forcing is a heating of the climate system due to absorption of the increased thermal radiation. The increased thermal radiation is mainly absorbed at the earth's surface where the heating is larger over land than over ocean. The heating at the surface is redistributed in the vertical and horizontal by dynamical processes in the atmosphere (and oceans) such as convection, baroclinicity, mean winds, and ocean currents. The large-scale mean response in geopotential height is connected via hydrostatic balance to the vertically integrated mean temperature response.

The mean NH winter (December–February) response in $T_s$, MSLP, $H_{500}$, and $H_{200}$ is displayed in Fig. 1. To assess the statistical significance of the response in these variables a two-sided unpaired $t$ test is applied at each grid point. The response in $T_s$ is significant at the 99% confidence level in 100% of the grid points. The response in MSLP, $H_{500}$, and $H_{200}$ is significant at the 99% confidence level in 64%, 100%, and 100% of the grid points, respectively. Thus the response at high altitudes is stronger and more significant than in MSLP.

The primary spatial pattern of the $T_s$ response is stronger heating over land than over oceans. This pattern resembles the Cold Ocean–Warm Land (COWL) pattern first diagnosed by Wallace et al. (1996). They find that this pattern is responsible for approximately half of the temporal variance in observed monthly mean hemispherically averaged surface air temperature. The trend in the amplitude of this pattern contributes substantially to the observed warming trend over the last 25 yr (Broccoli et al. 1998). A strong increase in the heating toward the Arctic is also evident in Fig. 1. This heating is coupled to the melting of sea ice.

MSLP is decreased in the NH, with the strongest decrease found over the northern continents and the Arctic. The spatial pattern of the MSLP response shows some resemblance to the $T_s$ response. Some correspondence between the location of regions of strong response in $T_s$ and $H_{500}/H_{200}$ is also found. The pole-to-equator gradient in the $H_{200}$ response is coupled to an increase in the upper-level temperature gradient. The increase in GHG concentrations and decrease in stratospheric ozone lead to a heating of the troposphere and a cooling of the stratosphere. The result is an increase in the upper-level pole-to-equator temperature gradient.

The spatial pattern of the $H_{200}$, $H_{500}$, and MSLP response to the enhanced GHG forcing has a strong zonally asymmetric component. This zonally asymmetric component varies among different CGCM simulations (cf. Carnell and Senior 1998; Williams et al. 2001; Joseph and Ting 2004). We hypothesize that it is associated with changes in the propagation conditions for barotropic Rossby waves. The results of BK04 suggest a similar relationship for the SH.
b. The leading modes of NH variability

A changed forcing of the climate system can affect the frequency of occurrence and/or the spatial patterns of the leading modes of variability. We focus on changes in the first EOFs of MSLP, H500, and H200 in response to the enhanced GHG forcing. However, for H500 and H200 the second EOF is also discussed. We choose to study the EOFs of IWV north of 20°N. The choice of hemispheric rather than regional EOFs is motivated by a recent study by Selten et al. (2004), who find that the (62-member ensemble mean) response to an enhanced GHG forcing in a CGCM is associated with a wave train encompassing the whole NH. The spatial structure of this wave train, termed the circumglobal waveguide pattern, has a great deal in common with the structure of the North Atlantic Oscillation (NAO) over the North Atlantic (Branstator 2002).

The first and second EOFs of IWV in MSLP, H500, and H200 for the years 1860–89 are displayed in Fig. 2. The first EOF explains 13%–14% and the second EOF explains 10%–11% of the IWV in MSLP, H500, and H200, respectively. The first EOF (EOF1) of IWV in MSLP in the NH resembles the Arctic Oscillation (AO; Thompson and Wallace 1998). The North Atlantic centers of action in the AO coincide qualitatively with the centers of action of the so-called NAO (e.g., Hurrell 1995). The spatial patterns of EOF1 of H500 and H200 resemble the Pacific–North American pattern (PNA; Wallace and Gutzler 1981) over the North Pacific Ocean and North America.

Figure 3 displays the first and second EOFs of IWV in MSLP, H500, and H200 for the last 30 yr (2070–99). The total amount of IWV in the region 20°–90°N is increased by 11% as compared with the first 30 yr. The first EOF explains 14% and the second EOF explains 11%–13% of the IWV in MSLP, H500, and H200, respectively. The relative amount of variability explained by these modes is thus not changed significantly. A comparison of Figs. 2 and 3 reveals that the spatial pattern of the first AO-like EOF of MSLP is relatively unchanged in response to the enhanced GHG forcing. The pattern is however slightly displaced. Ulbrich and Christoph (1999) diagnose this shift in the centers of action of the NAO. They find that the shift is associated with changes in the storm track activity in this region.

The spatial pattern of EOF1 of H500 and H200 for the first 30 yr (1860–89; Fig. 2) resembles EOF2 for the
last 30 yr (2070–99; Fig. 3) over the northern Pacific Ocean and North America. However, over the North Atlantic Ocean EOF1 for 1860–89 differs from both EOF2 and EOF1 for 2070–99. Similar results are seen for EOF2 for 1860–89 as compared with EOF2 and EOF1 for 2070–99. We wish to test if these differences in the spatial patterns of the first two EOFs are associated with a gradual change in the EOF patterns or if it is a result of inadequate sampling. To test this, two comparisons are performed. First, the EOFs of H200 are determined for the 60-yr periods 1860–1919 and 2040–99 and compared with the 30-yr period EOFs for 1860–89 and 2070–99.

The first two EOFs of IWV in H200 for the periods 1860–1919 and 2040–99 and compared with the 30-yr period EOFs for 1860–89 and 2070–99.

The first two EOFs of IWV in H200 for the periods 1860–1919 and 2040–99 are clearly separated [according to the North et al. (1982) criterion]. We determine the pattern correlation of the EOFs for the 30-yr periods with the EOFs for the 60-yr periods. Because the sign of the EOFs is arbitrary, we report that the absolute value of the pattern correlation of EOF1 for 1860–89 with EOF1 (EOF2) for 1860–1919 is 0.18 (0.97). The pattern correlation of EOF2 for 1860–89 with EOF1 (EOF2) for 1860–1919 is 0.96 (0.18). The ordering of the first two EOFs is thus changed when a 60-yr period is used instead of a 30-yr period. The pattern correlation of EOF1 for 2070–99 with EOF1 (EOF2) for 2040–99 is 0.99 (0.08). The pattern correlation of EOF2 for 2070–99 with EOF1 (EOF2) for 2040–99 is 0.99 (0.08). For these periods, the ordering of the first two EOFs is unchanged when a 60-yr period is used instead of a 30-yr period. Based on the high correlations between the 30- and the 60-yr period EOFs we conclude that the 30-yr period EOFs are significant.

Second, the EOFs of H200 are determined for eight consecutive 30-yr periods (1860–89, 1890–1919, . . . , 2070–99). The first two EOFs of IWV in H200 are clearly separated [according to the North et al. (1982) criterion] for all periods except years 1950–79. For this period only the first EOF is clearly separated and the second EOF is therefore excluded from the analysis. The pattern correlation of EOF1 and EOF2 for 1860–89 with EOF1 and EOF2 for the seven following 30-yr periods is displayed in Fig. 4. The highest correlation for EOF1 (EOF2) for 1860–89 is found with EOF2 (EOF1) for each of the seven other 30-yr periods. Thus, the ordering of the first two EOFs differs between the

Fig. 2. The (a), (b), (c) first and (d), (e), (f) second EOF of IWV north of 20°N in (a), (d) MSLP, (b), (e) H500, and (c), (f) H200 for December–February years 1860–89.
first 30-yr period and all the other 30-yr periods. We therefore focus on the correlation of EOF1 (EOF2) for 1860–89 and EOF2 (EOF1) for the seven consecutive 30-yr periods. This correlation gradually decreases from 0.85 (0.92) to 0.72 (0.80). This gradual decrease in the correlation indicates that the differences in the spatial patterns of the first two EOFs of IWV in H200 are associated with a gradual change in the EOF patterns and not with inadequate sampling. The correlation of EOF1 (EOF2) for 1860–89 and EOF1 (EOF2) for the seven consecutive 30-yr periods increases from 0.09 (0.10) to 0.42 (0.46). These relatively high correlations indicate that the EOFs for the later part of the transient integration are combinations of the EOFs for 1860–89. This change may however not be attributed to inadequate sampling due to the gradual change in the spatial patterns.

The leading modes of IWV in H500 and H200 resemble large-scale Rossby waves. Whether there are dynamical connections between the NH regions indicated by the EOFs must however be investigated using dynamical analysis. In the following section, we test the possibility that the changes in the spatial patterns of EOF1 and EOF2 of H500 and H200 are associated with changes in the propagation conditions for barotropic Rossby waves.

4. Rossby wave dispersion

The basis of the analysis is the dispersion relation for plane barotropic Rossby waves in a horizontally varying basic-state flow. To assess the propagation of barotropic Rossby waves we consider solutions of the barotropic vorticity equation on the sphere. We linearize about a time-independent, but zonally and meridionally varying, basic state. Assuming that there is separation in scale between the mean flow and the perturbations, we look for a solution of the barotropic vorticity equation using Wentzel–Kramers–Brillouin theory. This criterion is strictly not satisfied if the wave patterns in the EOFs (Figs. 2 and 3) are regarded as the perturbations to the mean flow. The spatial scale of the EOFs is similar to that of the changes in the mean flow (Fig. 1). The result of applying this method to a relatively small-scale mean flow is a method that is more sensitive to smaller-scale properties of the basic flow than the real atmo-
spheric system. Thus, using this method we may exclude differences in the dispersion of barotropic Rossby waves as a potential explanation for the changes in the leading modes.

The zeroth-order equation gives the dispersion relation. In a coordinate system with horizontal coordinates \((x, y)\) given by \(dx = a \cos(\phi) d\phi, dy = a d\phi; \phi\) is longitude, \(\phi\) is latitude), this equation may be written as follows:

\[
\omega = U \kappa k + V \lambda l + \frac{k^2 \frac{\partial Q}{\partial y} - k \frac{\partial Q}{\partial y}}{k^2 + l^2}
\]

\[
= \mathbf{V}_\Psi \cdot \mathbf{K} + \frac{\mathbf{z} \cdot (\nabla Q \times \mathbf{K})}{k^2 + l^2},
\]

(1)

where

\[
\mathbf{V}_\Psi = (U_\Psi, V_\Psi) = \left(-\frac{\partial \Psi}{\partial y}, \frac{\partial \Psi}{\partial x}\right)
\]

is the basic-state nondivergent wind, \(\Psi(x, y)\) is the basic-state streamfunction, \(Q = \nabla^2 \Psi + f\) is the basic-state potential vorticity, \(f = 2\Omega \sin \phi\), \(\mathbf{K} = (k, l)\) is the perturbation wave vector, and \(\mathbf{z}\) is a unit vector directed vertically upward. A derivation of this equation is given by, for example, Karoly (1983).

For stationary waves \((\omega = 0)\), the total wavenumber \(|\mathbf{K}|\) may be written [using Eq. (1)] as follows:

\[
K^2 = (k^2 + l^2)^{1/2}(a \cos \phi) = \frac{\partial Q / \partial Y}{|V_\Psi|} (a \cos \phi)^2.
\]

(2)

Here, a local coordinate system with the \(X\) axis parallel to the basic-state wind is used. Furthermore, \(K_s\) represents an equivalent zonal (nondimensional) wavenumber. For stationary Rossby waves in a zonally symmetric basic state, Hoskins and Ambrizzi (1993) show that Rossby rays are always refracted toward latitudes with larger \(K_s\). We determine \(K_s\) from the December–February zonal mean wind at 200 hPa for the first and last 30 yr of the transient simulation. This zonal mean \(K_s\) is displayed in Fig. 5, where the SH has been included for reference. From this figure, we note that the zonal mean propagation conditions for Rossby waves are essentially unchanged in the NH. In the SH however, the midlatitude waveguide (at approximately 70°–40°S) is strengthened. This strengthening is associated with a significant change in the spatial patterns of the first two EOFs of IWV in SH H200 (BK04).

The mean atmospheric flow is not zonally symmetric, particularly not in the NH. Averaging over all longitudes may hide large-scale, but not zonally symmetric, differences in, for example, \(K_s\). We therefore determine \(K_s\) also as a function of longitude and latitude. In Fig. 6a \(K_s\) is displayed for 1860–89. In agreement with the Hoskins and Ambrizzi (1993) results for observations, the upper-level flow in the CGCM shows maxima in \(K_s\), along the Asian jet, along the North Atlantic jet, and along the SH jet. Regions where \(K_s^2\) is negative are shaded in white. By Eq. (2), this occurs when \(\partial Q / \partial Y\) is

---

**Fig. 4.** The absolute value of the pattern correlation of EOF1 (heavy dashed line) and EOF2 (light dashed line) of IWV in H200 for 1860–89 with EOF1 and EOF2 for seven consecutive 30-yr periods (1890–1919, ..., 2070–99). The absolute value of the pattern correlation of EOF1 (heavy line) and EOF2 (light line) of IWV in H200 for 1860–89 with EOF2 and EOF1 for seven consecutive 30-yr periods.
negative, that is, when \(\nabla Q\) is to the right of the local wind \(V_{\text{loc}}\). This occurs on the flanks of the westerly jets where the curvature of the wind exceeds the gradient of planetary vorticity. It also occurs in some parts of the Tropics where the wind is easterly.

In Fig. 6b \(K_s\) is displayed for 2070–99. Comparison of the \(K_s\) fields for the periods 1860–89 and 2070–99 shows changes in the refractive index at about 40°S and over the Caribbean. The SH minimum in \(K_s\) at about 40°S deepens, whereas the minimum over the Caribbean flattens. The strengthening of the SH jet is associated with distinct changes in the propagation conditions for barotropic Rossby waves, as concluded by BK04. For the NH, however, the changes in \(K_s\) are of small scale and difficult to couple to the changes in the spatial patterns of the leading modes of variability (Figs. 2 and 3). The possible influence of these changes in \(K_s\) on the propagation of barotropic Rossby wave perturbations is investigated following Karoly (1983). A Rossby wave ray is defined to be the integral path of the group velocity and shows the propagation of wave activity. We are however not able to distinguish between the mean flow for 1860–89 and for 2070–99 with this method. Since this method is more sensitive to smaller-scale

---

**Fig. 5.** Zonal mean \(K_s\), defined in Eq. (2), at 200 hPa for December–February 1860–89 (weak line) and 2070–99 (heavy line) in the CGCM.

---

**Fig. 6.** \(K_s\), defined in Eq. (2), at 200 hPa for December–February (a) 1860–89 and (b) 2070–99 in the CGCM, where \(K_s^2 < 0\) in the white regions.
properties of the basic flow than the real atmospheric system, we conclude that the changes in the atmospheric upper-level flow cannot explain the changes in the EOFs.

5. Discussion and conclusions

The response of the mean and the variability of the extratropical NH atmospheric circulation to an enhanced GHG forcing is investigated using data from a transient integration with the ECHAM4/OPYC3 CGCM. The response of the NH atmospheric circulation to the enhanced GHG forcing differs among different CGCMs. AchutaRao et al. (2004) compare 11 CGCMs and find that differences among models in the amount of global surface warming are substantial, spanning about a factor of 2 by the time of CO₂ doubling. The amount of global surface warming in the simulation used in this study is near the mean of the 11 models compared by AchutaRao et al. (2004). The hypothesis that the response to anthropogenic forcing projects onto modes of natural variability (Palmer 1999) implies that the modes of natural variability do not themselves change in response to the forcing. In the present study we show that the spatial patterns of the NH leading modes of variability in the transient integration with the ECHAM4/OPYC3 CGCM change in response to the enhanced GHG forcing. We investigate the possibility that these changes in the leading modes are connected to changes in the propagation conditions for barotropic Rossby waves.

The mean $T_s$ response resembles the COWL pattern (Wallace et al. 1996). The spatial pattern of the mean MSLP response shows some resemblance to the $T_s$ response. Some correspondence between the location of regions of strong response in $T_s$ and H500/H200 is also found. The mean response in geopotential height consists of two components, a zonal mean change that varies with height, and a mainly equivalent barotropic zonally asymmetric pattern. The zonal mean geopotential height response is determined by the vertically integrated temperature response. The NH lower-tropospheric meridional temperature gradient is decreased in winter whereas the upper-tropospheric meridional temperature gradient is increased. The zonally asymmetric component of the response in geopotential height varies among different CGCM simulations (cf. Carnell and Senior 1998; Williams et al. 2001; Joseph and Ting 2004). We test the possibility that it is associated with changes in the propagation conditions for barotropic Rossby waves. The results of BK04 suggest a similar relationship for the SH.

The leading mode of IWV in MSLP in the NH resembles the AO, whereas the leading modes of IWV in H500 and H200 show a closer resemblance to the PNA pattern. Episodes of persistence in the PNA-like mode are associated with equivalent barotropic persistent anticyclones and cyclones in geopotential height (not shown). The spatial patterns of the PNA-like mode in geopotential height changes in response to the enhanced GHG forcing. The changes are most prominent over the North Atlantic–European region. To test the significance of these changes in the spatial patterns we (i) determine the EOFs for the first and last 60 yr of the transient integration and (ii) determine the EOFs for eight consecutive 30-yr periods of the integration. From these tests we conclude that the changes in the spatial patterns of the EOFs occur gradually during the transient integration and may not be attributed to inadequate sampling. From this we conclude that not only the frequency of occurrence of the leading modes of variability, but also the spatial patterns of these modes can change in response to the enhanced GHG forcing.

To test the hypothesis that these changes in the spatial patterns of the EOFs are associated with changes in the propagation conditions for barotropic Rossby waves, the equivalent zonal stationary Rossby wave number $K_s$ is determined. The zonal mean propagation conditions for Rossby waves are essentially unchanged in the NH. To test for zonally symmetric, large-scale changes in the propagation conditions, $K_s$ is determined as a function of latitude and longitude. The only significant NH changes in $K_s$ in response to the enhanced forcing occur over the Caribbean. The possible influence of these changes in $K_s$ on the propagation of barotropic Rossby wave perturbations is investigated following Karoly (1983). We are however not able to distinguish between the mean flow for 1860–89 and for 2070–99 with this method. Because this method is more sensitive to smaller-scale properties of the basic flow than the real atmospheric system, we conclude that the changes in the atmospheric upper-level flow cannot explain the changes in the EOFs.

The result that the leading modes change in response to an increasing GHG forcing agrees with BK04’s results for the SH. However, the changes in the spatial patterns in the NH are not associated with changes in the propagation conditions for barotropic Rossby waves, as was the case for the SH. Another possible mechanism for the changes in the spatial patterns is changes in the occupation statistics for the leading modes. We conclude that this is not the case for the transient integration studied here. The variability explained by the leading EOFs is not changed significantly.

A possible mechanism is changes in the forcing of the
large-scale extratropical flow in terms of changes in tropical flow. Support for this idea is given by, for example, Branstator (2002). He studies the propagation of low-latitude initial disturbances. He finds a fundamental difference in the behavior of initial disturbances that reside in regions of weak and regions of strong meridional gradients in the mean wind. Investigation of the relevance of differences in the basic state for the changes in variability presented here is left to future work.

Acknowledgments. The author thanks H. Thiemann at the Deutsche Klimarechenzentrum for kindly providing the climate model data, and Dr. Jonas Nycander and Professor Erland Källén for useful suggestions and discussions. This work was performed as a part of the Swedish Regional Climate Modelling Programme, SWECCLIM, which is supported by the Strategic Funds for Environmental Research, MISTRA. The Grid Analysis Display System was used for drawing the figures.

REFERENCES


