AO/NAO response to climate change: 2. Relative importance of low- and high-latitude temperature changes

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[1] We address the issue of why different models may be getting different responses of the AO/NAO in climate change experiments. The results from part 1 (Rind et al., 2005) suggest that for substantive climate changes, the differences are likely to be found in the patterns of tropospheric climate change, rather than from the stratosphere. We assess the various tropospheric forcings through a variety of experiments. We first use extreme paleoclimate experiments (Ice Age, Paleocene) which feature large variations in the low level latitudinal temperature gradient; the results show that under these circumstances, changes in the eddy transport of sensible heat, and in situ high latitude forcing, dominate the AO response. We next test the effect of more modest SST temperature gradient changes in the current climate, and find a similar result with a model configuration that does not easily transport the low level temperature changes into the upper troposphere. We then reanalyze the results from different 2 $\times$ CO2 experiments with the GISS model and find that they can be understood by assessing: (1) the magnitude of tropical SST warming; (2) the translations of that warming into the upper troposphere; (3) the change in the extratropical low altitude temperature gradient; and (4) the change in the high latitude SST/sea ice response. We suggest that these features might explain the varying results among modeling groups, and that forecasts will not converge until these features do.

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1. Introduction

[2] The question of how the Arctic Oscillation (AO)/North Atlantic Oscillation (NAO) changes in response to climate alterations is of continuing interest, for if it can be ascertained, it can help us forecast some component of regional climate response. When the AO/NAO is more positive, Alaska and Eurasia are warmer, and northeastern North America and the North Atlantic are cooler. Storm tracks over the Atlantic are displaced to the north and east. The general temperature pattern is often referred to as COWL for cold ocean, warm land [e.g., *Wallace et al.*, 1996; *Broccoli et al.*, 1998].

[3] Over recent decades, the tendency for a more positive phase of the AO has been observed, and there are differing points of view as to what has caused it. *Shindell et al.* [1999] found in the GISS model simulations that it was caused by greenhouse warming due to anthropogenic influence, which could only be expressed in a model with a full stratosphere. Furthermore, in that paper, and work by *Rind et al.* [1998] it was concluded that this more positive phase was likely to continue as climate continued to warm. This would have numerous consequences, for example, increasing sea ice reduction in the North Atlantic [e.g., *Comiso et al.*, 2003]. However, there has been no consistent agreement among model simulations that have looked at the effect of increasing CO_2, although more of them have found an increase in the positive phase of the AO or NAO [*Paeth et al.*, 1999; *Fyfe et al.*, 1999; *Gillett et al.*, 2002, 2003a, 2003b; *Hu and Wu*, 2004; *Sigmund et al.*, 2004] than did not [*Osborn et al.*, 1999; *Zorita and Gonzalez-Rouco*, 2000]. In fact, a more positive AO phase due to increased CO_2 does not always arise even within the suite of GISS models [*Rind et al.*, 2001a, 2002].

[4] In the first part of this work [*Rind et al.*, 2005] (henceforth part 1), we investigated the relative contribution of stratospheric and tropospheric climate changes to AO/NAO variations in the GISS Global Climate Middle Atmosphere Model. It was concluded that for climate changes that are fully expressed with the usual magnitude (i.e., those that have had years to develop, allowing the sea surface temperatures (SSTs) to respond), tropospheric climate changes dominate the stratospheric influence. While most model simulations show the standard picture of stratospheric cooling in response to increased CO_2, there is much less conformity concerning the pattern or magnitude of tropospheric climate response. The results from part 1 suggest that it is the different tropospheric temperature pattern of change that is most likely producing the different AO forecasts in different models.

[5] Therefore, in this paper, we focus on how the different distribution of tropospheric temperature changes can result

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Table 1. Experiments Discussed in This Paper and in Part 1

Experiment	Purpose	Main Result
Warm and Cold Climate Experiments in part 1	To investigate the relative effect of stratospheric and tropospheric climate changes on the AO/NAO phase	Tropospheric warming and high latitude stratospheric cooling lead to a more positive AO phase, dominated by changes in eddy angular momentum transport; tropospheric changes are of greater importance
Paleoclimates	Investigate which tropospheric changes are response for the AO/NAO phase by using extreme climate changes at high and low latitudes to clarify their respective effect	Extreme low latitude temperature gradient changes make eddy sensible heat flux transports the dominant influence on the AO/NAO phase change, along with local high latitude conditions
Altered Temperature Gradients in the current climate	To investigate whether the results from the extreme paleoclimate experiments arise in more moderate climate change situations	Again, sensible heat (and hence potential vorticity) transports dominate the angular momentum change, when temperature changes are not large in the tropical upper troposphere
Altered Temperature Gradients in the Atlantic and Pacific Ocean		And again, local high latitude conditions can influence the results
2cCO2 experiments in different GISS models	To test whether the results from the previous experiments can help explain the conflicting results seen in the future climate simulations done with different GISS GCMs.	The GISS model forecast of future NAO/AO phase changes depends upon the boundary layer and cloud/convective parameterizations in the tropics, and sea ice response at high latitudes

in altered AO/NAO patterns, and what processes and model parameterizations may be responsible for producing them. In part 1 it was emphasized that the latitudinal temperature gradient plays an important role, by influencing both planetary wave refraction and eddy energy generation. The temperature gradient can be altered by both high and low latitude responses; in this paper we address more directly the effect of such temperature changes on the phase of the AO and NAO circulations.

[6] To do this, we use a variety of climate change experiments, whose description, purpose with respect to this paper, and principle results are summarized in Table 1 (for comparison, we also show in this table the same aspects for the experiments done in part 1). To clarify the respective influence of temperature changes at high and low latitudes we first present experiments in which the high and low latitude temperatures have undergone severe climate perturbations, as represented by paleoclimate simulations (section 3.1 and Table 1). We conclude that in agreement with part 1 more extreme warm climates have a more positive AO phase, while cold (Ice Age) climates a more negative phase. However, the mechanisms producing this response are somewhat different from what occurred under more moderate conditions. Due to the large alteration in low level temperature gradient in these experiments, it is the change in eddy sensible heat transports, and hence potential vorticity transport that is the determining factor for the extratropical circulations, rather than the angular momentum transport discussed in part 1. The extreme local high latitude surface changes also play a direct role.

[7] To understand the importance of low altitude temperature gradient changes under less extreme conditions more likely for future CO_2 increases, we next use a set of experiments in a finer resolution model in which the SST gradient is altered in a variety of ways (section 3.2 and Table 1). Again, the sensible heat and potential vorticity transports dominate the angular momentum transport change. An important conclusion is that changes in wave refraction and angular momentum transport are not large when low altitude temperature changes are not mixed efficiently to high altitudes, as occurred in this particular model. A subset of these experiments, changing the temperature gradient in the Atlantic or Pacific Oceans separately, emphasizes once again the importance of in situ conditions at high latitudes.

[8] In the last set of experiments (section 3.3 and Table 1), we return to the issue of the lack of consistency in extratropical circulation response in doubled CO_2 experiments. We use different simulations done with different versions of the GISS model (section 3.3 and Table 1), with results showing that the conclusions derived from the previous sets of experiments apply here as well: the AO/NAO phase change as climate warms is determined by the low level temperature gradient change unless there is large warming in the tropical upper troposphere. These model-dependent results are affected by the boundary layer and cloud/convection schemes in the tropics, and sea ice response. In the discussion section (section 4) we compare these results to those from other models, highlight how low and high latitude temperature changes influence the AO/NAO phase,

Table 2. Change of Northern Hemisphere Temperature and Dynamics in the More Extreme Paleoclimate Experiments During December–February[a]

	Ice Age				Paleocene	
	ICE AGE	ICE AGE-A[b]	ICE AGE-MW	ICE AGE A-MW[b]	PAL	PAL-A[b]
ΔSURF TEMP (Annual)	***−4.26***	***−10.2***	***−5.19***	***−10.0***	***3.33***	***7.53***
ΔSURF TEMP 4°N	**−1.7**	***−6.5***	**−0.8**	***−6.3***	**3.4**	***5.9***
ΔSURF TEMP 74°N	***−23***	***−28.3***	***−19.4***	***−25.3***	***12.3***	***17***
Δ272MB TEMP	−0.2	***−6.7***	−0.8	***−7.2***	***2.9***	***9.0***
Δ 68MB TEMP	0.2	***−1.5***	0.2	***−1.6***	−0.8	***1.2***
Δ 1.5MB TEMP	***4.8***	***4.5***	***4.6***	***4.3***	***−8.3***	***−8.0***
ΔAO INDEX SLP (MB)	***−21.6***	***−21.5***	***−16.2***	***−17.2***	**6.27**	**8.02**
ΔNAO (MB)	***−24.7***	***−22.4***	***−13.1***	***−17.8***	NA[c]	NA[c]
ΔAO INDEX 100MB (M)	***−306***	***−411***	***−222***	***−350***	***227***	***331***
ΔAO INDEX 10MB (M)	***−187***	−60	***−360***	***−250***	***1714***	***1391***
ΔTROP EKE (%)	***8.0***	***6.5***	***26.1***	***25.9***	2.4	***−17.3***
ΔTROP WAVE#1-4 (%)	1.2	***−10.5***	***26.9***	***15.3***	−0.4	***−11.9***
[STANDING EKE #1-4(%)]	***[−20.6]***	***[−40.2]***	***[−15]***	***[−34.6]***	***[−77.4]***	***[−74.3]***
ΔEDDY NT ANG MOM(%)	***−51***	***−51***	1.9	11.8	5.9	***41***
CORR: ANG MOM/EKE	−0.25	***0.82***	0.09	***0.90***	−0.55	***0.82***
CORR: ANG MOM/REFR	0.24	***0.93***	−0.29	***0.85***	0	***0.74***
ΔEDDY NT QGPV (%)	***−19.3***	***−56.6***	***−15.9***	***−49.8***	***30.7***	***39.0***

[a]Energy and transport terms are averaged over the Northern Hemisphere. Significant results at the 95% confidence level are in bold italics.
[b]Results for the "alternate" experiments.
[c]The North Atlantic was much narrower; the altered land/ocean ratio makes the NAO comparison inapplicable.

and estimate what is most likely to happen as climate warms during this century. Conclusions are presented in section 5.

2. Models and Experiments

2.1. Extreme Paleoclimate Experiments

[9] As indicated in Table 1, the extreme climate changes from these experiments are used to help clarify the relative importance of low latitude and high latitude climate response to the phase of the AO/NAO. The paleoclimate experiments used were originally published by *Rind et al.* [2001b] for the extreme cold and warm climates (8° × 10°, 23 layer model): (1) Ice Age (ICE AGE, 21k years ago) and (2) Paleocene (PAL, 55 million years ago).

[10] Both the ICE AGE and PAL experiments used two different SST data sets, a standard version, and an alternate version with more extreme tropical changes which represent realistic alternatives in line with some paleodata: colder in the tropics in the Ice Ages (ICE AGE-A), warmer in the Paleocene (PAL-A). In addition, the Ice Age experiments also varied the stratospheric gravity wave drag over the ice sheets, utilizing a rough topography and thus large mountain wave forcing in the standard experiments, and a smoother topography with less wave drag in the additional simulations (−MW); these latter experiments are used here to determine how much the stratosphere could actually influence the troposphere when the climate change is so extreme.

2.2. Experiments With Altered Latitudinal Temperature Gradients

[11] To investigate whether the results from the previous experiments are applicable in more moderate climate change experiments, we use the following simulations: (1) experiments with specified changes in latitudinal temperature gradients discussed by *Rind* [1998] (4° × 5°, 9 layer model). The experiments increased (I) or decreased (D) the tropical to high latitude temperature gradient, first with no net climate change, and then with global cooling and an increased gradient (CI), or global warming and a decreased gradient (WD). They are used here to directly assess the relative influence of temperature gradient changes on the AO/NAO. As a subset of these runs, we also use simulations with changes in latitudinal gradient in the Atlantic and Pacific oceans, separately [*Rind et al.*, 2001c] to isolate the effect of in situ temperature changes on the NAO.

2.3. Doubled CO_2 Experiments With Different GISS Models

[12] Finally, to determine whether the results from the two previous sections are actually applicable to doubled CO_2 climate simulations, we use different doubled CO_2 experiments that have been done with GISS models: (1) the 2 × CO2 experiments described in *Rind et al.* [2002] with the 4° × 5°, 53 layer model; (2) the 2 × CO2 experiment of *Rind et al.* [2001a] with a 4° × 5° 31 layer model; (3) the doubled CO_2 experiment of *Rind et al.* [1998] with the 8° × 10°, 23 layer model. Experiments were performed using two different sets of sea surface temperature changes, distinguished primarily by the magnitude of tropical warming, hence 2CO2, and 2CO2WT for Warmer Tropics, in the different models. Simulations were also done with (+O3) and without ozone response.

[13] Most of these experiments utilized specified SSTs, and ran for 20 years. Significance of the changes is judged via a Student T test with respect to the interannual variations in the respective control runs.

3. Results

3.1. Extreme Paleoclimate Experiments: Influence of High-Latitude Temperatures

[14] In part 1, we found that warm climates (or a cold polar stratosphere) produced a more positive AO/NAO phase, while cold climates (or a warm polar stratosphere) resulted in a more negative phase. To investigate whether that result holds for more extreme paleoclimate conditions, we use the experiments for the last Ice Age (~21kya) and the Paleocene (58Mya) described above. Shown in Table 2 are the temperature changes at various levels in the troposphere and stratosphere. The temperature changes *are*

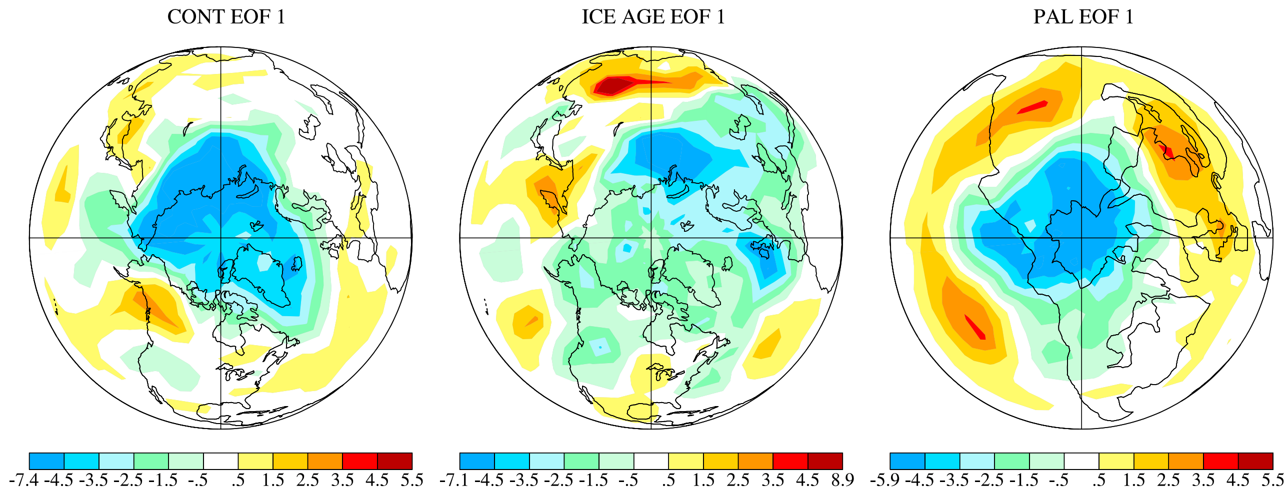

Figure 1. The leading EOF of monthly sea level pressure variations for December–February in the control run (CONT, top), the standard Ice Age experiment (IA, middle) and the standard Paleocene run (PAL, bottom). Results are similar for the other Ice Age and Paleocene simulations. The integrated value of the EOF pattern poleward of 60° is scaled to be −1.

extreme, ranging from a global annual surface air temperature increase of 7°C (PAL-A) to a decrease of 10°C (ICE AGE-A). In December–February, surface temperature changes at 74°N were even larger, with cooling of close to 30°C at high northern latitudes associated with large continental ice sheets (ICE AGE-A), or warming of 15–20°C in conjunction with little sea ice (PAL-A) and warm polar SSTs. The response in the upper troposphere was dependent on the experiment, with strong cooling or warming in the alternate experiments (ICE AGE-A, PAL-A) associated with the larger equatorial surface temperature changes. While stratospheric temperatures as a whole varied by 1–2°C (warmer in the Last Glacial Maximum with less CO_2, cooling in the Paleocene with more CO_2), in the lower stratosphere polar regions Ice Age temperatures increased by up to 10°C, and Paleocene temperatures decreased by some 20°C. The experiments also involve more than just climate (temperature) changes, as the topography was very different, and in the Paleocene, the continental positions were altered.

[15] The first question to ask about these experiments is whether they are suitable for addressing questions concerning the AO; is the AO in fact the leading mode of variability in climates as extremely different as these, and is its proportion of the total variability similar to that currently? The 1st EOF of monthly (Dec., Jan., Feb.) sea level pressure for the control run and experiments ICE AGE and PAL are shown in Figure 1, and it can be seen that the leading mode of variability in each case features pressure changes at high latitudes (e.g., 60–80°N) of opposite sign to those at mid-latitudes (e.g., 30–50°N). (There are some differences between 50–60°N, in particular that region is more associated with high latitudes in IA than in the other experiments.) The percentage of variance explained by this leading mode of variability is also similar for each climate regime (Control: 13.6%; ICE AGE: 13.8%; PAL: 12.6%) (in comparison, the first leading mode in the observations represents 22% of the variance [e.g., *Shindell et al.*, 1999]). The results are similar for the other Ice Age and Paleocene experiments (ICE AGE-A, PAL-A, etc.). The AO therefore does seem to be a comparable feature in these extreme climates, which is an interesting result in itself.

[16] The Ice Age simulation was both colder in the troposphere, and had a warm polar stratosphere, so the results in part 1 would suggest it should have a more negative AO/NAO phase. Inversely, with the Paleocene being a warmer climate with a colder polar lower stratosphere, there should be a more positive phase. We also show in Table 2 the resulting circulation index changes from the variety of experiments performed for each of the time periods. Clearly the expectations from part 1 are achieved, and the differences in AO/NAO values at the surface greatly exceed those from the experiments discussed in part 1. (The circulation indices are calculated as discussed in part 1, from the differences in sea level pressure at mid versus high latitudes for the AO, and from sea level pressure at grid-points surrounding Portugal and Iceland for the NAO.) The effects are also highly barotropic in nature, with the same sign of the change extending from the surface through the middle stratosphere. Partly this is the result of the extreme nature of the lower level changes, but also there is no extreme climate forcing in the lower stratosphere – the changes in effect are being initiated at the surface, a situation which previously was shown to favor a more barotropic response in this model (part 1). The Northern Hemisphere sea level pressure field during winter for the standard and alternate experiments are shown in Figure 2, with the changes of extremely high significance. The standard and alternate experiments have very similar sea level pressure responses, indicating that alterations in boundary conditions other than the SSTs and sea ice dominate the result. This is not the case for the doubled CO_2 experiments discussed later in this paper.

[17] There are some differences among the experiments in the degree of AO phase change. The Ice Age experiments with reduced gravity wave drag over the ice sheets (labeled −MW) have somewhat smaller negative phase change responses at the surface. Given that the direct effect of the reduced drag is an increase in the west wind velocity in the upper troposphere/lower stratosphere, this is an example of

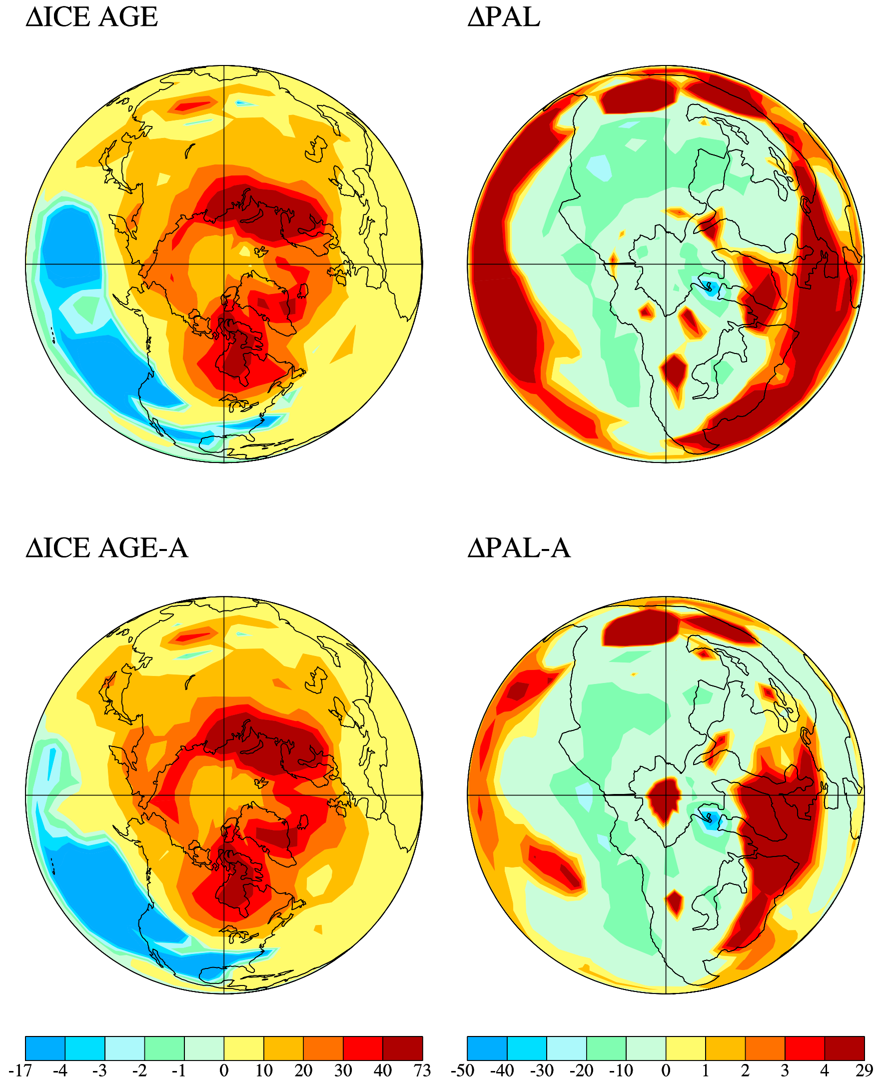

Figure 2. Change in sea level pressure in Dec–Feb for the Ice Age experiments (left) and Paleocene experiments (right). The top row shows the results with more modest tropical SST changes (cooling in the LGM, warming in PAL), while the bottom row shows the results with greater cooling/warming in the tropics. Note that the sea level pressure field in both cases has been corrected uniformly for the change in mean sea level relative to the current climate (correction of −11.5mb for the Ice Age runs, +17mb for the Paleocene).

how changes in those layers can affect the surface circulation, as discussed in part 1.

[18] In part 1, the circulation changes at high latitudes were closely coupled to changes in angular momentum transport, wave refraction and eddy energy; in particular, variations in the meridional wave refraction and corresponding poleward eddy momentum transports were largely responsible for determining the phase of the AO/NAO. Is that true in these experiments? In the last section of Table 2 we indicate the changes in the Northern Hemisphere average eddy (TROP EKE) and planetary wave (TROP WAVE# 1-4) energy, the changes in the Northern Hemisphere vertically-integrated eddy northward transport of angular momentum (EDDY NT ANG MOM), and its correlation with the change in Northern Hemisphere eddy energy (CORR: ANG MOM/EKE) and the wave refraction (CORR: ANG MOM/REFR), calculated following the same procedure as in part 1. Unlike the results in part 1, the relationship between the circulation index changes and these diagnostics is now ambiguous. For the Ice Age experiments with the rough ice sheets, reduced momentum transport is consistent with the more negative AO index, but when the ice sheets were smoothed, the angular momentum transport change was actually slightly positive. While both Paleocene experiments do feature increased momentum transport in concert with their more positive AO phase, for the standard experiment the change was not significant, while the AO change was very large. The results from these six experiments indicate that the hemispheric-average angular momentum transports do not correlate significantly (at the 95% level) with the AO response. (Note that using hemispheric averages for momentum transport might conceal important changes in the meridional structure of this flux, but all these experiments have a similar latitudinal

profile as the control run, with maximum vertically-integrated eddy momentum transport centered at around 40°N, and magnitudes falling rapidly further poleward.)

[19] Furthermore, in part 1, the eddy momentum transport changes were correlated with changes in refraction, and to some extent eddy energy. The results, as shown in Table 2, indicate that only in the (alternate) experiments in which large temperature changes occurred in the upper troposphere were such correlations maintained. Nevertheless, all the experiments had very large AO changes; how did that occur?

[20] The actual eddy forcing of the zonal wind flow is associated with the divergence of the Eliassen-Palm (EP) flux, the sum of the latitudinal convergence of eddy momentum transport plus the altitudinal change in eddy heat transport. The EP flux divergence (divided by the density) is proportional to the northward transport of quasi-geostrophic potential vorticity (QGPV) ($v'q'$) which can be written as [e.g., *Holton*, 2004, p.329]

$$\overline{v'q'} = -\partial\frac{\overline{u'v'}}{\partial y} + \frac{f_0}{\rho_0}\frac{\partial}{\partial z}\left(\frac{\rho_0}{N^2}\overline{v'\frac{\partial\Phi'}{\partial z}}\right)$$

where u' and v' represent the zonal and meridional wind perturbations, ρ is the density, N the Brunt-Vaisala frequency, f the coriolis force and $\frac{\partial\Phi'}{\partial z}$ is proportional to temperature. The first term is associated with the latitudinal convergence of the eddy angular momentum transport, and the second term on the right represents the change with altitude of the eddy sensible heat flux. In part 1 we noted that the change in AO/NAO was strongly correlated with the eddy angular momentum transport, although the correlation was actually stronger with eddy transport of QGPV. In these experiments, while the relationship with the angular momentum transport is inconsistent, as shown in Table 2 the relationship with the Northern Hemisphere vertically-integrated QGPV transport is always of the proper sign, an increase in transport being associated with strengthening of the high latitude west winds and a more positive AO phase. (The results are insensitve to whether the vertically-integrated hemispheric average of $\overline{v'q'}$ is used, as in this case, or the peak latitudinal value or some other measure of the extratropical transport is employed.)

[21] The difference in these experiments from the ones discussed in part 1 is the very large change in the low level atmospheric temperature gradient. The second term in the equation for QGPV is normally negative, for sensible heat transports decrease with altitude as the latitudinal gradient decreases (the eddy sensible heat transport decrease is a factor of three, per unit mass, between the lower and upper troposphere in this model for the control run). In the Ice Age experiments, this decrease with altitude is very large (a factor of five per unit mass in the alternate ice age runs), providing for even more negative QGPV transport, and a more negative AO phase. In the alternate Paleocene experiment the decrease with altitude is smaller than in the control run, due to the reduced low level latitudinal temperature gradient, so the QGPV transport is more positive, and eddy forcing at high latitudes intensifies the circulation. Hence one difference from part 1 is the importance of eddy sensible heat transports on determining the AO/NAO phase.

[22] This, however, is not the complete story, because the AO/NAO index changes are equally as large in the ice age runs with smaller changes in QGPV transport. In these cases, the altered AO phase is strongly affected by the surface boundary conditions. During the Ice Ages, the presence of the cold high altitude ice sheets in the region of 50–70°N results in massive high pressure cells dominating the high latitudes, as air descends from the cold ice sheets (at various longitudes, and in particular over the North Atlantic). The presence of expanded sea ice helps limit heat fluxes from the ocean, which then works to stabilize the atmosphere and weaken low pressure systems at the higher latitudes. At the same time, the increased temperature gradient south of the ice sheets (and sea ice) results in strong zonal winds and more zonal (as opposed to northeastward) storm tracks.

[23] For the Paleocene, the situation is reversed: there is no ice on Greenland, very little sea ice at high northern latitudes, and relatively warm SSTs which can act to destabilize the atmosphere. As a result, cold air does not build up at high latitudes, and a relatively more positive phase of the AO predominates.

[24] Altered surface boundary conditions are thus another way that tropospheric climate changes influence the sea level and tropospheric pressure/height fields. To the extent that the surface changes influence the tropospheric wind field at higher altitudes, their influence on planetary wave propagation can then extend up into the stratosphere. With surface conditions this extreme compared with today, the sea level pressure results are relatively insensitive to the different SST fields used in the standard and alternate experiments.

3.2. Experiments With Altered Latitudinal Temperature Gradients

3.2.1. Zonal Average Changes

[25] The runs discussed in section 3.1 emphasized the importance of the low level temperature gradient and high latitude in situ influence, but they were from climate simulations that had extreme changes to surface features. To investigate whether these influences are important in more moderate climate change situations, we utilize the experiments discussed in *Rind* [1998], in a 4° by 5°, 9 layer model. While the stratosphere is poorly resolved in this model, the results from part 1 suggest that this should not greatly affect results from experiments with strong SST gradient changes. In these experiments, the latitudinal temperature gradient in SSTs was altered by warming the tropics and cooling high latitudes, or the reverse (unlike the experiments in part 1, in which the surface temperature changes were generally of similar sign at all latitudes). In one set of experiments, the different changes of temperature at the different latitudes compensate for one another, so there is no global temperature change (i.e., high latitudes were cooled to the same degree the tropics was warmed; EXP I for increased gradient), or high latitudes were warmed to the same degree the tropics were cooled (EXP D for decreased gradient). The increased latitudinal SST gradient of EXP I was maintained in the next experiment but all temperatures were decreased by 4°C (hence EXP CI for colder with increased gradient) as colder climates are expected to have increased gradients. Finally, the decreased

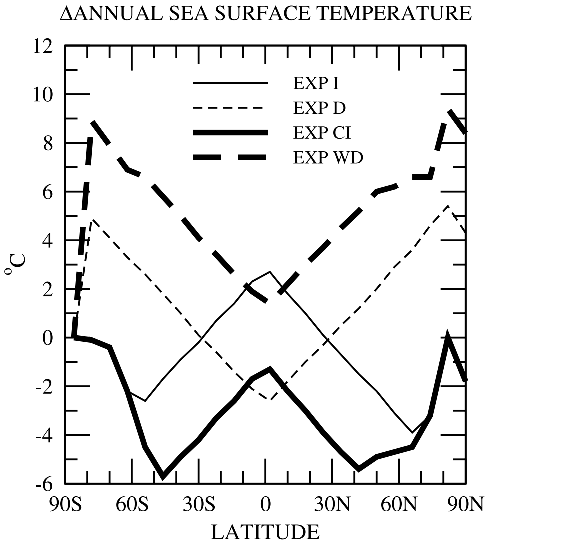

Figure 3. Change in SSTs in the experiments with specified gradients.

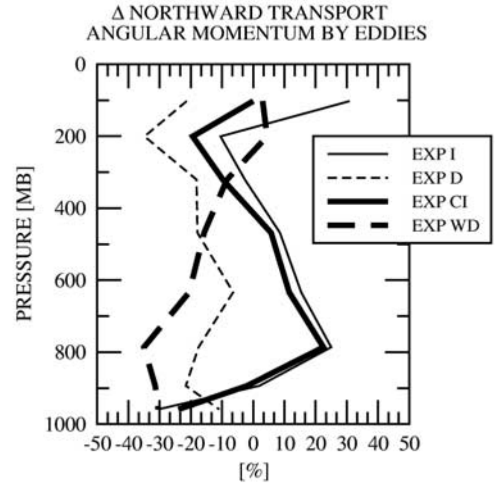

Figure 4. Change in Northern Hemisphere eddy transports of angular momentum for December–February in the different temperature gradient experiments. Differences greater than 5–10% are significant at the 95% confidence level.

SST gradient of EXP D was maintained but all the temperatures were increased by 4°C (WD) as warmer climates are expected to have decreased gradients. The zonal mean sea surface temperature changes from the control run for these experiments is reproduced in Figure 3, and the resulting temperature and circulation changes indicated in Table 3.

[26] The latitudinal temperature gradient increased in EXP I while it decreased in EXP D. In part 1 it was shown that an increased gradient alters both wave refraction and eddy generation so as to favor more poleward angular momentum transport and a more positive phase of the AO/NAO. The experiments with the increased gradient (EXP I, CI) do have more eddy and planetary wave energy as expected (see Table 3, TROP EKE, and TROP WAVE #1-4), while with the decreased gradient, eddy energy is less (EXP D, WD). The decreased gradient is also associated with decreased angular momentum transport. However, as indicated in Table 3, the AO/NAO phase change is more positive in EXP D with this model, and while the individual changes are not significant, the difference between EXP I and EXP D is. Comparing the results from EXP CI and EXP WD, again it is the run with the reduced latitudinal gradient that has the more positive AO/NAO phase change.

Table 3. Northern Hemisphere Changes in SST Gradient Experiments

	EXP I[a]	EXP D	EXP CI[a]	EXP WD
ΔSURF TEMP	0.24	−0.06	***−4.94***	***6.17***
ΔSURF TEMP 4N	***2.7***	***−2.6***	***−1.8***	***1.9***
ΔSURF TEMP 74N	***−2.0***	***2.2***	***−13.4***	***22.1***
Δ320MB	***2.9***	***−2.5***	***−2.7***	***3.3***
Δ 26MB TEMP	***−0.7***	***0.5***	***−0.5***	***1.0***
ΔAO INDEX SLP	−1.16	1.16	***−6.20***	***5.58***
ΔNAO (MB)	***−4.67***	−1.33	***−5.90***	***3.34***
ΔAO INDEX 100MB (M)	***111***	−38	−29	49
ΔAO INDEX 30MB (M)	***267***	24	21	***185***
ΔTROP EKE (%)	***10.9***	−2.2	***9.2***	***−9.4***
ΔTROP WAVE#1-4 (%)	***7.6***	1.0	3.1	***−6.0***
ΔEDDY NT ANG MOM(%)	12.7	***−20.9***	−1.5	−6.0
ΔEDDY NT QGPV (%)	−9.4	3.6	***−28.6***	***35.1***

[a]Results for the experiments using an increased latitudinal SST gradient.

[27] We can again assess how the circulation change is arising by comparing the eddy transports of angular momentum (Figure 4) and QGPV (Figure 5). The angular momentum transport changes do not match the AO/NAO phase changes in these experiments but the QGPV transport changes provide the right sign of the forcing in all the experiments. Clearly it is the heat transport changes that are dominating, with the largest effects at low levels where the altered temperature gradient is most pronounced. In particular, the reduction in the latitudinal temperature gradient in EXP D and EXP WD results in a strong reduction in the eddy heat transport at low levels throughout the extratropics, hence the second term in the equation for QGPV transport is much less negative, increasing the

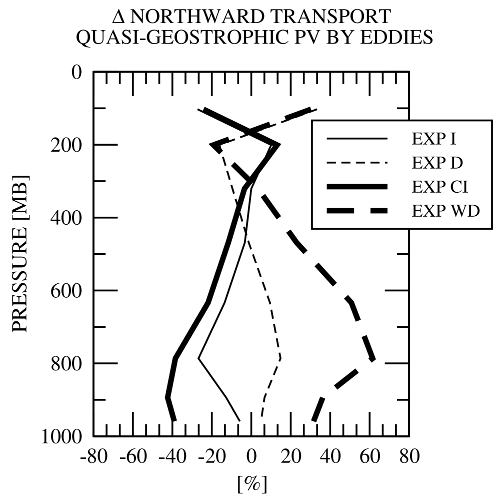

Figure 5. Change in Northern Hemisphere eddy transport of quasi-geostrophic potential vorticity for December–February in the different temperature gradient experiments. Differences greater than 5% are significant at the 95% confidence level.

Table 4. Temperature and Circulation Index Changes for Experiments in Which the SST Gradient is Altered Separately in the Atlantic and Pacific Oceans

	ΔTemp 958Mb Eq/50N	ΔTemp 320Mb Eq/50N	ΔNAO	ΔAO
Atlantic Ocean Increased gradient minus Decreased gradient (Ai − Ad)	**2.5/−3.5**	***4.0/−0.6***	***−3.77***	−0.40
Pacific Ocean Increased gradient minus Decreased gradient (Pi − Pd)	***4.1/−5.1***	***7.6/−3.6***	***4.47***	−1.24
Decreased gradient in Atlantic, increased gradient in Pacific minus the reverse (AdPi − AiPd)	***2/−0.9***	***4.5/−0.6***	***11.11***	−0.13

northward transport of QGPV and providing for more eddy forcing and for a more westerly circulation index at high latitudes. This result is now occurring without the other surface/topography variations found in the more extreme paleoclimate experiments.

[28] Why is this result different from that found in part 1? As shown in Table 3, the low level temperature changes are not strongly amplified in the upper troposphere in this model; the temperature change at 320mb is similar to that at the surface. Therefore the change in wave refraction and angular momentum transport is minimized (as shown by the generally insignificant momentum transport changes given in Table 3). Hence the low level temperature gradient effect dominates and produces QGPV transport changes opposite in sign to the momentum transport changes. In contrast, in part 1, in the model that effectively translated low level temperature responses into the upper troposphere, it was noted that the QGPV and momentum transport changes were correlated with a coefficient of 0.9.

[29] The results therefore suggest that both tropical warming (if transported to high altitudes) and high latitude warming (at low levels) can lead to a more positive AO phase via their effects on QGPV transport and wave forcing. As noted for the Paleocene run, high latitude warming can have a direct impact as well, by altering the stability via increased ocean heat fluxes.

3.2.2. Changes in Individual Ocean Basins

[30] We test the "in situ" influence in an additional series of temperature gradient experiments, utilizing the runs discussed in *Rind et al.* [2001c]. Here the temperature gradients in EXP I and EXP D were applied to the Atlantic and Pacific individually; following the nomenclature of that paper, Ai or Ad for increased/decreased gradient in the Atlantic, Pi or Pd for increased/decreased gradient in the Pacific.

[31] Experiment Ai minus Experiment Ad (i.e., Ai-Ad) for the Atlantic then forced the climate with warmer tropical SSTs and colder high latitude SSTs in the Atlantic alone; the temperature changes in that ocean basin at high and low latitudes and at high and low altitudes in the troposphere are shown in Table 4. The effect on the extratropical circulation, also indicated in Table 4, was a more negative NAO index, as can also be seen in the sea level pressure differences shown in Figure 6. Hence the tropical warming was less important than the effect of the cooling that was occurring at high latitudes.

[32] In contrast, for the same experiment in the Pacific (Pi-Pd), the result was a more positive NAO index (Figure 6, middle, and Table 4). Here the tropical warming was more important, as the high latitude cooling was not in situ for the NAO; it can be seen in the table that there is a significant

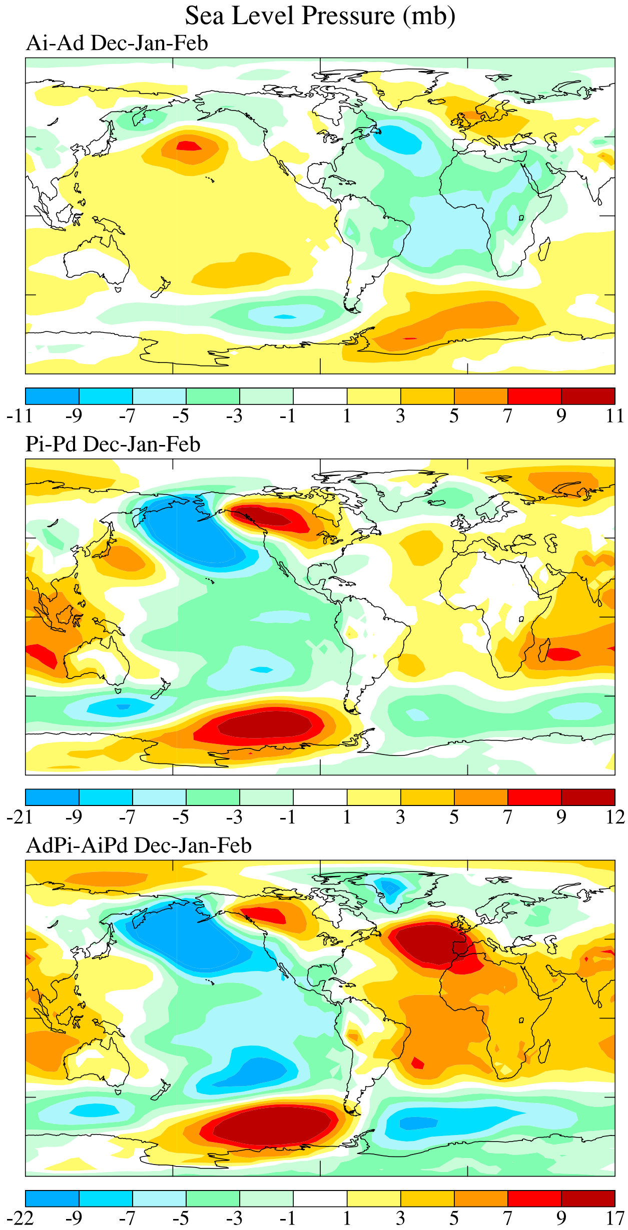

Figure 6. Change in seal level pressure between different pairs of sensitivity experiments with the sea surface temperature gradient altered in the Atlantic Ocean, Pacific Ocean, or in combinations in both oceans. (top) Increased gradient experiment (Ai) (warmer tropics, colder poles in the Atlantic Ocean) minus decreased gradient (Ad). (middle) Increased gradient in the Pacific Ocean (Pi) minus decreased gradient (Pd). (bottom) Difference between experiments AdPi and AiPd. Most changes outside of the lowest ranges shown (−2.2 to 2.2 mb) are significant at the 95% level.

Table 5. Northern Hemisphere Changes in 2 × CO2 Experiments With Different GISS GCMAMs

	4 × 5, 53 Layer					
	CT[a]	CT + O_3[a]	WT	WT + O_3	4 × 5 31L CT[a]	8 × 10 23L WT
ΔSURF TEMP	***4.1***	***4.1***	***5.1***	***5.1***	***4.0***	***6.0***
ΔSURF TEMP 4N	***2.9***	***2.8***	***5.1***	***4.9***	***2.5***	***5.4***
ΔSURF TEMP 74N	***7.9***	***8.1***	***9.3***	***9.4***	***9.1***	***11.6***
Δ270MB	***2.7***	***2.8***	***4.9***	***4.8***	***2.3***	***8.2***
Δ 68MB TEMP	***1.1***	*0.5*	***1.9***	0.4	−0.1	***2.4***
Δ 1.5MB TEMP	***−10.3***	***−9.3***	***−10.5***	***−8.9***	***−10.9***	***−8.9***
ΔAO INDEX SLP	***2.28***	***1.98***	***4.91***	***5.43***	−1.14	***4.47***
ΔNAO (MB)	−0.22	***5.67***	***9.22***	***10.45***	0.8	***5.77***
ΔAO INDEX 100MB (M)	76	12	103	45	−19	***231***
ΔAO INDEX 10MB (M)	196	−57	−96	***−220***	58	***224***
ΔTROP EKE (%)	0.8	***−3.0***	−0.4	−0.9	−0.7	***−3.3***
ΔTROP WAVE#1-4 (%)	***6.4***	0.7	***4.7***	***3.1***	−1.4	***12.0***
ΔEDDY NT ANG MOM(%)	−5.8	0.6	−3.5	−7.1	−7.9	13.4
ΔEDDY NT QGPV (%)	***5.8***	1.7	***20.7***	***14.4***	***8.8***	***16.4***

[a]Results for the experiments using a smaller tropical warming.

amplification of the warming in the tropical upper troposphere. Finally, when we combine Atlantic high latitude warming (and tropical cooling) with Pacific tropical warming (and high latitude cooling), and compare that experiment with the reverse gradient alterations (AdPi-AiPd), the NAO change is very large and positive (Figure 6, bottom, and Table 4). It is this configuration, of tropical and high latitude response, warm tropical Pacific and warm northern North Atlantic, that is optimal for forcing a more intense NAO. The warm tropical Pacific through its expression in the upper troposphere provides for more equatorward wave refraction and poleward angular momentum transport; the warm northern North Atlantic provides for decreased low altitude northward sensible heat transport, as well as more in situ destabilization of the local atmosphere. (One can also see in Figure 6 that there is considerable response in the Southern Hemisphere as well, even during summer, which relates to sea ice variations seen in that hemisphere [*Rind et al.*, 2001c].) As shown in the table, the effect on the AO, made up of both Atlantic and Pacific circulations, was muted in all cases.

[33] In summary, these experiments indicate the importance of the high and low latitude temperature response for the AO/NAO phase change, and suggest that without strong amplification of the upper tropospheric temperature response, the local high latitude temperature change can have a greater influence than that in the tropics.

3.3. Doubled CO_2 Experiments With Different GISS Models

[34] We are now in a position to investigate the AO/NAO response to the doubled CO_2 climate as found in several different GISS models. A key determinant for altering the AO phase in the variable SST experiments described in part 1 was the response of the tropical SSTs; as their change was amplified in the upper troposphere, wave propagation and planetary wave generation were strongly affected. How general is this conclusion?

[35] We test that result with the different GISS models. We employ the simulations reported by *Rind et al.* [2001a, 2002], for doubled CO_2 experiments with different magnitudes of SST warming, especially in the tropics. Three different versions of the GISS model were involved, 4° × 5°, 31 layers, 4° × 5° 53 layers, and 8° × 10°, 23 layers; some of the physics also differed in the models, in particular the cloud/convection and boundary layer schemes. All the models have tops at the mesopause, and all use parameterized gravity wave drag. Analysis has been done in the previous publications of the effects of the resolution and physics changes, and we will make use of some of that here.

[36] Given in Table 5 are the results from those experiments, including the temperature changes at various levels, the surface temperature response at low and high latitudes, changes in the relevant AO/NAO indices, and in eddy energy and eddy transports. The experiments labeled WT for 'warm tropics' all utilize the same input SST field (although at different resolutions), as do the runs labeled CT for 'cool tropics'.

[37] Consistent with the expectations from part 1, the experiment with the warmer tropical SSTs had the greatest positive AO and NAO phase changes at the surface. The simulations without as much tropical warming do not produce a consistent response; while the 53 layer model did have a (smaller) positive phase change, the 31 layer model phase change was not significant. As shown in the table, the 31 layer model had the smallest tropical warming, and the weakest warming in the tropical upper troposphere. This result shows that an increasing positive phase of the AO/NAO indices does not necessarily arise in a warming climate; the results depend on the particulars of the climate response, in particular the warmth of the tropical SSTs.

[38] However, in the different WT runs, despite the similar tropical SST changes, there are significant differences at higher altitudes: the warming in the upper troposphere (270mb) is much reduced in the finer resolution model, due to more limited transport of surface heat through the boundary layer, and the reduced convective fluxes into the upper troposphere [*Rind et al.*, 2001a, 2002]. The positive nature of the AO phase change away from the surface is also weaker in these models (at 100 mb). We noted in part 1 (Figure 4a) that with the coarser grid model, warming experiments produced increased northward momentum transport at most altitudes from the middle troposphere through the lower stratosphere. Shown in Figure 7 is the change in northward momentum transport by eddies in these experiments. Except for the coarse grid model, all the others actually have reduced momentum transport throughout the troposphere, with positive values occurring only in

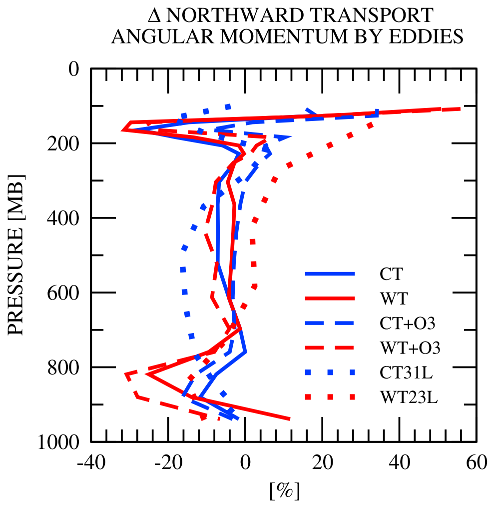

Figure 7. As in Figure 4, except for the different 2 × CO2 experiments.

the lower stratosphere. Without the strong heating in the (tropical) upper troposphere, the zonal wind change in the troposphere in the finer resolution models does not lead to extensive equatorward wave refraction changes (the normalized index of refraction change for the troposphere as a whole, calculated by comparing eddy momentum and heat transport changes varies by less than 10% in the fine resolution runs, while it increased by 50% in the coarse grid model). Hence the troposphere as a whole does not receive added angular momentum convergence at high latitudes, and the AO phase change is not nearly as barotropic as it was in the coarse grid model; as shown in Table 4, the change in tropospheric eddy transport of angular momentum is generally negative.

[39] We show in Figure 8 the change in the northward transport of QGPV and now the WT experiments have a more northward QGPV flux (which increases zonal winds) in the troposphere below 500mb, and also near the tropopause. In fact, the big difference between the WT and CT runs is in the lower troposphere, so the positive AO/NAO surface effect could very well be produced by changes in that region primarily. Note that the coarse grid model response has much greater uniformity in altitude in this diagnostic as well. Overall, the WT runs have the largest percentage change in this transport (and thus eddy forcing) (Table 4).

[40] Thus, while the warmer tropical SSTs did lead to a more positive AO phase response, it was for a very different reason in the 4 × 5, 53 layer model than in the model used in part 1. In the 4 × 5, 53 layer model, the WT runs also had strong warming at high latitudes, a reduced low level latitudinal temperature gradient, and reduced eddy sensible heat transports at low elevations, allowing the QGPV transports to be more positive. In the 8 × 10, 23 layer model the WT run had strong warming in the tropical upper troposphere, altered planetary wave refraction and greater poleward angular momentum transport.

[41] In summary, the different GISS model responses are due to (1) different magnitudes of tropical SST warming; (2) different translations of that warming into the upper troposphere; (3) different changes in the extratropical low altitude temperature gradient; and (4) different changes in the high latitude SST/sea ice responses.

4. Discussion

4.1. Comparison With Other Models

[42] A primary goal of this investigation has been to help explain why different models produce different AO/NAO phase change responses to increasing atmospheric CO_2. The summary given above at the end of section 3 may very well be applicable to these other models as well, and differences in their tropospheric climate change with respect to those features should determine the extratropical circulation change they produce. Only if the tropospheric effect is relatively neutral, then the results from part 1 suggest the stratospheric influence could be decisive, with cooling in the high latitude lower stratosphere (or warming in the low latitude lower stratosphere) producing a more positive phase response.

[43] As noted in the introduction, most, but not all, of the model simulations published concerning the change in AO or NAO phase due to greenhouse gas warming have found an increase in the positive phase in the troposphere. Few of the models, however, show the detailed temperature distribution that would allow for a determination of the different contributing terms to wave forcing at high latitudes. *Zorita and Gonzalez-Rouco* [2000] compared the extratropical circulation response in the HadCM2 and ECHAM models for changes between 2100 and today. The ECHAM model had somewhat greater high latitude warming and a more positive AO response, while results were ambiguous with HadCM2. One would have to compare all the different aspects of the simulations to determine how the various features of their warming influence the QGPV or angular momentum transport. *Sigmund et al.* [2004] used the ECHAM middle atmosphere model which produced warming due to doubled CO_2 of greater than 6°C in the tropical upper troposphere, as well as strong warming (of some 12°C) at high latitudes. Their model response featured what

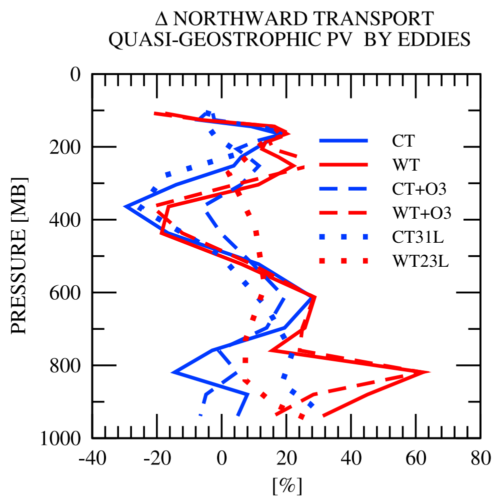

Figure 8. As in Figure 5, except for the different 2 × CO2 experiments.

appeared to be a strengthened NAO, and increased west winds in the Southern Hemisphere extratropics, but only the latter was directly associated with tropospheric forcing in their analysis.

[44] The impact of the low level temperature gradient primarily concerns QGPV transports at those levels, and the term is of less consequence for influencing the lower stratospheric response. Whether the stratospheric polar vortex weakens or strengthens is then dominated by the wave refraction/wave generation effects associated with heating of the tropical upper troposphere, and it could be opposite the surface response if that is due to changes in sensible heat transports. For example, *Butchart et al.* [2000] in a model with just 3°C warming in the tropical upper troposphere by the 2050s found that there was no consistent change in polar vortex strength among the different simulations performed. *Gillett et al.* [2002] with the HadSM3 model found a more positive AO response at the surface while at the same time the stratospheric winds weakened (negative AO response), as did *Sigmund et al.* [2004] with the ECHAM model.

4.2. Importance of Tropical and High-Latitude Temperature Changes

[45] The model results studied in this paper emphasize the importance of both low and high latitude surface temperature changes, as well as changes in the low level latitudinal temperature gradient. The influence of low latitude SST anomalies on the AO phase can be seen in observations, in part, because El Ninos (warm Pacific SSTs) are known to be associated with a deeper Aleutian Low, hence a more positive AO phase for the Pacific component [e.g., *Schneider et al.*, 2003]. A more positive NAO in response to warmer tropical SSTs has also been concluded, either from the Pacific/Indian Ocean region [*Hoerling et al.*, 2001, 2004; *Hurrell et al.*, 2004] or from the Atlantic [*Terray and Cassou*, 2002]. While the precise mechanism for such connections is still not well understood, the model results here and in part 1 would suggest planetary wave refraction as an important component.

[46] Concerning high latitudes, *Watanabe and Nitta* [1999] and *Cohen et al.* [2001, 2002] presented observations which suggest that the winter AO signal originates in the lower troposphere in eastern Siberia during fall, and is forced by snow cover variations, similar to the impact of the boundary forcings described here. In addition, observations show a decrease in Arctic sea ice occurring in conjunction with the more positive AO/NAO phase [e.g., *Comiso et al.*, 2003]. While much attention has been focused on the atmospheric circulation change helping to reduce sea ice through advection, the model results suggest the feedback would also be the other way, with reduced sea ice leading to a more positive AO phase by helping to destabilize the atmosphere. *Rodwell et al.* [1999] and *Latif et al.* [2000] concluded that warmer high latitude SSTs were forcing lower pressures at high latitudes, with increased latent heat flux leading to increased precipitation, and condensational heating destabilizing the atmosphere. Note that because of this direct, local high latitude influence, it would be difficult to deduce the tropical temperature response from high latitude circulation anomalies, as in the *Mann et al.* [1998] temperature reconstruction for the past 1000 years.

4.3. Predicting Future Changes in AO/NAO

[47] To predict future phase changes in the AO/NAO, both at the surface and in the stratosphere, we have to determine how tropical SSTs will change, how well the change will be imparted to the upper troposphere, how SSTs will change at high latitudes, and what will happen to the low level temperature gradient in the extratropics. If these effects are all relatively neutral, than changes in the low and high latitude lower stratospheric temperatures could influence the result.

[48] What do current trends and model projections indicate about each of these aspects? First, it is well-established that there has been a positive trend in the phase of the AO/NAO over the last 30 years [e.g., *Hartmann et al.*, 2000; *Thompson et al.*, 2000]. How much of this trend is due to global warming and how much to natural variability cannot yet be ascertained; simulations with GCMs often produce the observed pattern of response, but most have failed to produce the proper magnitude [e.g., *Gillett et al.*, 2003b].

[49] The degree of tropical warming that has occurred over the past century is on the order of 60% of that in the extratropics [*Rind*, 2000]; if this continues it would suggest substantial tropical warming during the 21st century, which would favor a more positive AO/NAO phase if that warming was dispersed and amplified at higher altitudes. However, mixed layer depths are smallest in the tropics, and this smaller heat capacity would allow the tropics to warm more rapidly than other latitudes initially. Furthermore, complex non-linear feedbacks with cloud systems might dampen (or amplify) tropical warming as greenhouse gases increase further. Model simulations for the future climate have little consistency when it comes to tropical warming predictions, part of the overall uncertainty (of a factor of three) in total climate sensitivity. For the tropics in general it is not yet possible to know how much warming will occur.

[50] A subset of tropical warming, an increase in El Nino frequency, would also favor a more positive AO/NAO response. Current indications of El Nino frequency change are mixed, and may be related to the Pacific Decadal Oscillation [e.g., *Trenberth and Hoar*, 1996]. For future projections, models in general tend to produce an increase in an "El Nino-like" state (with the eastern tropical Pacific warming more than the western tropical Pacific in the mean) [e.g., *Jin et al.*, 2001]; this would favor a more positive NAO response, as shown by the results in section 3.2.2. As for the El Nino itself, there is no agreement among models, partly because the models are not able to generate accurate El Ninos in the control run [*Latif et al.*, 2001].

[51] How well will any tropical warming be translated into the upper troposphere, a crucial component for influencing planetary wave propagation changes? In the GISS GCM, that depends on the convection and boundary layer schemes [*Rind et al.*, 2001a], and as result differ greatly between models (also related to the overall question of tropical sensitivity). Current trends of upper tropospheric tropical temperatures are very uncertain, and may be influenced by ozone trends. At this point we don't know how large the response in the upper troposphere will be in conjunction with surface warming.

[52] How much will high latitude temperatures warm? Current trends show substantial warming in the Arctic [*Comiso et al.*, 2003], and high latitude warming over Eurasia and Alaska is a prominent feature of surface temperature change records. At least some of this is correlated with the change in the AO/NAO itself [*Thompson and Wallace*, 1998]. Models show significant amplification of climate warming at high latitudes during winter, and one can expect that would occur if there is substantial global warming; by itself, this would tend to reduce the low level latitudinal temperature gradient, and provide more in situ destabalization to the atmosphere, both aspects favoring a more positive AO/NAO phase. One caveat relates to changes in North Atlantic Deep Water (NADW) production, found by most models as climate warms [e.g., *IPCC*, 2001]. Cooling over the North Atlantic favors a more negative NAO phase (as in section 3.2.2); if the cooling is sufficiently substantial, a much more negative NAO phase could result [*Rind et al.*, 2001d]. The response of most future projections does not suggest a drastic reduction in NADW production, but, as in the case of El Ninos, lack of confidence in the ability of coupled models to generate NADW accurately makes all such projections somewhat questionable.

[53] Finally, cooling of the stratosphere in general is the expected result from increased CO_2, but for the Northern Hemisphere lower polar stratosphere during winter, the response is more uncertain. Ozone recovery by itself should provide some degree of warming in this region, but the overall result depends primarily on the stratospheric dynamical changes; with significant tropical warming, the stratospheric residual circulation intensifies and provides warming in the polar region from 100–10 mb [*Rind et al.*, 2002]. As noted in that paper, the same tropical warming induces a more positive AO/NAO phase irrespective of the stratospheric response. It is likely from a climate change perspective, that the patterns of tropospheric climate change will be the determinant of future AO/NAO phases.

[54] In summary, if there is significant tropical and high latitude warming, it is likely that the current tendency for increased positive phase of the AO/NAO will continue. Major questions relate to future changes in ENSO, the ability of the atmosphere to amplify low level changes in the tropical upper troposphere, and possible NADW reductions.

5. Conclusions

[55] In this study we investigated the aspects of tropospheric climate changes that would affect the AO/NAO indices. The overall goal was to elucidate possible mechanisms to explain why different models get different AO or NAO responses to a warming climate. The principal results are as follows:

[56] 1. Consistent with the results in part 1, the AO/NAO is influenced by tropical SST changes if their effects extend to high levels in the troposphere, at which point they influence planetary wave refraction and generation, and eddy angular momentum transport.

[57] 2. The AO/NAO is also influenced by high-latitude surface temperature changes that affect the atmospheric stability and pressure field.

[58] 3. High-latitude temperature changes also may alter the low level extratropical latitudinal temperature gradient, which then affects eddy sensible heat transports, and total eddy forcing of the zonal mean circulation.

[59] 4. This latter effect can dominate wave propagation and eddy momentum transport changes, especially if the tropical upper troposphere temperatures are not large.

[60] 5. The configuration that produces the largest changes in the NAO would involve tropical Pacific warming extending to high altitudes; high latitude warming in the North Atlantic; and cooling in the polar lower stratosphere.

[61] 6. To ascertain what will happen to the future AO/NAO, and understand the differences between models, we need to know how high and low latitude SSTs will change and whether the warming will be transported and amplified into the upper troposphere.

[62] 7. The tropospheric response, especially the degree of tropical upper tropospheric warming, will have consequences for NAM phase changes in the stratosphere, as the stratospheric response did for the AO/NAO discussed in part 1.

[63] Published research to date has generally not included sufficient information to evaluate the various influences discussed in this paper. We encourage such assessments with other models. Until model predictions converge on the important aspects of tropospheric climate change, it is unlikely that the predictions of changes in the AO/NAO will be consistent, or if they are, whether they will be occurring for similar reasons.

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