AO/NAO response to climate change: 1. Respective influences of stratospheric and tropospheric climate changes

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[1] We utilize the GISS Global Climate Middle Atmosphere Model and eight different climate change experiments, many of them focused on stratospheric climate forcings, to assess the relative influence of tropospheric and stratospheric climate change on the extratropical circulation indices (Arctic Oscillation, AO; North Atlantic Oscillation, NAO). The experiments are run in two different ways: with variable sea surface temperatures (SSTs) to allow for a full tropospheric climate response, and with specified SSTs to minimize the tropospheric change. The results show that experiments with tropospheric warming or stratospheric cooling produce more positive AO/NAO indices. Experiments with tropospheric cooling or stratospheric warming produce a negative AO/NAO response. For the typical magnitudes of tropospheric and stratospheric climate changes, the tropospheric response dominates; results are strongest when the tropospheric and stratospheric influences are producing similar phase changes. Both regions produce their effect primarily by altering wave propagation and angular momentum transports, but planetary wave energy changes accompanying tropospheric climate change are also important. Stratospheric forcing has a larger impact on the NAO than on the AO, and the angular momentum transport changes associated with it peak in the upper troposphere, affecting all wavenumbers. Tropospheric climate changes influence both the AO and NAO with effects that extend throughout the troposphere. For both forcings there is often vertical consistency in the sign of the momentum transport changes, obscuring the difference between direct and indirect mechanisms for influencing the surface circulation.

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1. Introduction

[2] The potential influence of the stratosphere on tropospheric extratropical circulation patterns has been the subject of numerous papers in the last decade, covering a wide variety of time-scales. Considering interannual variations, *Baldwin et al.* [1994] noted the apparent barotropic relationship between circulation anomalies in the troposphere and stratosphere. *Perlwitz and Graf* [1995] confirmed this feature, and emphasized the connection between the strength of the stratospheric cyclonic winter vortex and the tropospheric circulation over the North Atlantic. *Thompson and Wallace* [1998, 2000] showed that the Arctic Oscillation (AO), the leading Empirical Orthogonal Function of monthly sea level pressure anomalies during winter poleward of 20°N is structurally similar to the dominant mode of circulation variability in the lower stratosphere (referred to as the Northern Annular Mode, NAM). They also noted that there has been a recent upward trend (implying stronger west winds and lower pressure at high latitudes) in the index of these leading modes at altitudes ranging from sea level to the lower stratosphere.

[3] For short time-scales (days to a month), *Baldwin and Dunkerton* [1999, 2001], *Baldwin et al.* [2003], and *Thompson et al.* [2002] found that stratospheric NAM phase changes located just above the tropopause had often migrated downward from the middle stratosphere. At times this was followed by persistent sea level pressure AO anomalies, raising the possibility of using stratospheric changes to forecast subsequent tropospheric responses. *Limpasuvan et al.* [2004] have related stratospheric warming events to tropospheric pressure patterns, again finding a similarity in high altitude and surface pressure responses.

[4] On climate time-scales, radiative forcings affecting the stratosphere-troposphere system have been related to surface-level circulation variations. *Kodera* [2002, 2003] found that during solar minimum conditions, the NAO signal is confined to the North Atlantic while during solar maximum it extends over the Northern Hemisphere. *Ruzmaikin and Feynman* [2002] found that in early winter for the west QBO, and in late winter for the east QBO, solar

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forcing affected the NAM in both the troposphere and stratosphere. Numerous authors have noted the tendency for (in effect) a positive AO phase in the winters following volcanic aerosol eruptions [e.g., *Groisman*, 1992; *Graf et al.*, 1993; *Kodera*, 1994; *Kirchner et al.*, 1999; *Stenchikov et al.*, 2002, 2004]. The influence of polar ozone losses on inducing positive phase changes in the NAO has been studied by *Kindem and Christiansen* [2001] and *Schnadt and Dameris* [2003], while a trend toward the positive phase of the Antarctic Oscillation (AAO) has been modeled by *Gillett and Thompson* [2003]. The influence of increasing atmospheric CO_2 on the AO/NAO has been studied in models by *Shindell et al.* [1999], *Butchart et al.* [2000], *Zorita and Gonzalez-Rouco* [2000], and *Gillett et al.* [2002] (among others), with somewhat conflicting results (discussed further in *Rind et al.* [2005], henceforth part 2). *Shindell et al.* [2001a] summarized the GISS GCM response to these various climate perturbations, while *Rind et al.* [2004] reported on the AO/NAO changes that resulted from climate and ozone changes between the Maunder Minimum time period (late 1600s) and today: the cooler climate produced a more positive AO/NAO phase.

[5] The exact reason(s) for the relationships on these various time-scales remains unclear. Planetary wave-mean flow interaction is a leading candidate, as in the studies by *Kodera* [1994], *Rind et al.* [1998], *Shindell et al.* [1999], etc, with stratospheric thermal contrasts altering the zonal wind profile, and influencing the upward wave activity flux from the troposphere. This then leads to alterations in the zonal wind flow at lower levels, further altering upward wave energy propagation, until the effect extends down to the surface. On a climate time-scale what would be apparent is the time-averaged effect of such a process. In a study by *Baldwin et al.* [2003], there was some evidence that the NAM in the lower stratosphere may influence waves in the upper troposphere, whose transient momentum flux anomalies then induce circulation in the meridional plane that affects sea level pressure. The possible importance of synoptic-scale momentum fluxes or wave drag near the tropopause has also been emphasized by *Thorncroft et al.* [1993], *Shepherd* [2002], *Limpasuvan et al.* [2004], and *Wittman et al.* [2004]. A direct stratospheric forcing mechanism has been suggested by *Hartley et al.* [1998], *Ambaum and Hoskins* [2002], and *Black* [2002], in which potential vorticity anomalies associated with variations in the stratospheric polar vortex strength (initiated, for example, by EP flux convergences within the stratosphere) induce circulation anomalies in the troposphere via nonlocal geostrophic and hydrostatic adjustment. (This is similar to the *Baldwin et al.* [2003] mechanism except initiated above the tropopause.) Another option suggested is the importance of wave drag and reflection, often from the upper stratosphere [*Perlwitz and Harnik*, 2003, 2004]. Different processes may very well be more dominant on different time-scales.

[6] In attempting to isolate the stratospheric influence on tropospheric circulation when climate time scales are concerned, there is a basic problem: the climate forcing also directly affects the troposphere, causing tropospheric climate changes which themselves can affect the annular mode [e.g., *Hoerling et al.*, 2001]. This is true even for climate perturbations that originate within the stratosphere, such as stratospheric ozone or water vapor changes, and volcanic aerosol injections. It is less true if the perturbation acts over a shorter period of time, when its effect is reduced to that of an interannual forcing, since the sea surface temperatures do not have a chance to fully adjust (like the first year after a volcano, or the QBO).

[7] In this paper (part 1) we attempt to separate the various influences on the tropospheric annular mode that may be produced by climate perturbations, especially when the climate forcing is in the stratosphere. In effect we will address the question of the relative importance of the stratospheric-induced changes compared to tropospheric climate changes on influencing the high latitude/extratropical circulation. The results shown below indicate that both regions influence the surface response, although when tropospheric climate changes are large they dominate the stratospheric effect on the AO/NAO. In a subsequent paper (part 2), we show the relative importance of low and high latitude tropospheric climate changes on the phase of the dominant modes of variability; this will also address issues of why different climate models may get different AO responses in doubled CO_2 experiments. Together, answers to the altitudinal and latitudinal influences should allow us to better estimate how climate changes may alter the AO/NAO.

2. Model and Experiments

[8] The experiments used for this paper all employ the coarse grid GISS Global Climate Middle Atmosphere Model ($8^\circ \times 10^\circ$ resolution; 23 layers), extending from the surface to the mesopause. The following experiments are used: (1) doubled CO_2 in the troposphere and stratosphere (2CO2); (2) doubled CO_2 in the stratosphere alone (2CO2 Strat); (3) doubled water vapor in the stratosphere (2H2O Strat); (4) ±2% change in total solar irradiance (+2%TSI; −2% TSI); (5) ozone completely removed between 147 and 67mb (lower stratosphere) (No O3 LS); (6) volcanic aerosols (sulfate) with optical thickness of 0.15 (Volc); (7) stratospheric aerosols (soot + sulfate), each with an optical thickness of 0.15 (Sulf + Soot).

[9] Some of these experiments had been run previously [e.g., *Rind et al.*, 1998], but all were rerun or extended for the sake of this discussion. The soot + sulfate experiment is included mainly to provide a large impact in the lower stratosphere, but it might represent a particular type of volcanic eruption with large amounts of carbon in the earth's crustal structure, or extensive burning that has been hypothesized to accompany a 'nuclear winter' scenario, or even possibly a bolide impact.

[10] To distinguish between the tropospheric and stratospheric forcings of the AO/NAO, the model is run in two different ways. The first approach is to use specified sea surface temperatures at present day climatological values (specified as the mean for each month occurring in the middle of the month, with an annual sinusoidal variation between months); by limiting the tropospheric response, the stratospheric effect is amplified relative to that of the troposphere. (*Braesicke and Pyle* [2004] found that use of interannually varying SSTs from observations amplifies the ability of tropospheric forcing to influence stratospheric variability, relative to the use of the climatological SSTs, as was done here.) The second approach involves using a

Table 1a. Annual Temperature Changes in the Variable SST Experiments[a]

Experiment	SURF	346–203 MB (D-F 50°–90°N)	Trop	68 Mb (16N–16S)	1.5 MB	Strat
2CO2	***5.15***	***7.9 (3.4)***	***6.13***	***1.7 (3)***	***−8.6***	***−0.71***
2CO2 Strat	***0.54***	***1.1*** (1.0)	***0.77***	0.5 ***(1)***	***−8.7***	***−1.32***
2H20 Strat	***0.46***	***0.8*** (0.4)	***0.45***	−0.6 (0)	***−1.7***	***−1.8***
+2% TSI[b]	***4.73***	***7.8 (3.0)***	***5.94***	***3.6 (4)***	***1.4***	***2.78***
Volc	***−2.55***	***−3.6 (−1.6)***	***−2.91***	0.10 ***(2.3)***	−0.10	−0.01
Sulf+Soot	***−4.59***	***−1.7 (−1.6)***	***−1.51***	***34.8 (42)***	***6.6***	***31.28***
−2% TSI[b]	***−4.09***	***−5.6 (−1.6)***	***−4.68***	***−1.8*** (0)	***−1.4***	***−1.91***
No O3 LS[b]	***−1.15***	***−2.4 (−1.2)***	***−2.04***	***−8.8 (−12)***	0.1	***−5.71***

[a]Values in parenthesis refer to the latitudinal slices listed in the column heading. Significant results at the 95% confidence level are in bold italics.

[b]Experiments in which the temperature changes in the troposphere and stratosphere are of the same sign.

mixed layer ocean incorporating constant ocean heat transports (a "q-flux" ocean), allowing the ocean temperatures to adjust to the altered radiative forcing, and thus freeing up the tropospheric response, particularly in the low-to-mid troposphere. For the runs with specified sea surface temperatures, the integrations in general are carried out for at least 20 model years; with calculated sea surface temperatures the integrations extend for at least 50 years, with results averaged over the last 20 years (during which the model is in equilibrium with its climate forcing).

3. Results

[11] We discuss first the temperature changes in the different experiments, both in the troposphere and stratosphere. We then report on the AO/NAO responses, at the surface and at various atmospheric heights, including the relationship between changes in the troposphere, lower stratosphere and mid-stratosphere. We investigate the reason(s) for the AO/NAO changes, including changes in eddy energy, eddy transports, and wave propagation.

3.1. Temperature Response

[12] The first question to ask is, does this procedure work, does it really distinguish between stratospheric and tropospheric forcing? With the exception of the ±2% solar irradiance and CO_2 change throughout the atmosphere, the other experiments specifically target stratospheric perturbations of climate. Nevertheless, with varying SSTs, even stratospheric forcing experiments produce a significant tropospheric temperature response (as indicated in the following tables). In Tables 1a and 1b, we show the temperature changes at different levels in the atmosphere (at the surface, in the upper troposphere, averaged over the troposphere, in the lower and upper stratosphere, and averaged over the stratosphere). In Tables 1a and 1b, we separate the results between the experiments with variable SSTs (Tables 1a) and specified SSTs (Table 1b). Shown first are the experiments that cool the stratosphere as a whole and warm the troposphere (2CO2, 2CO2 Strat, 2H20 Strat); then the experiment that warms the stratosphere and warms the troposphere (+2% TSI); experiments that warm the stratosphere and cool the troposphere (Volc; Sulfur+Soot); and finally those that cool the stratosphere and cool the troposphere (−2% TSI; No O3 LS).

[13] Note that in the experiments with variable SSTs (Table 1a) the stratospheric response is not necessarily of the same sign throughout the region: with increased CO_2, the lower stratosphere warms, while the stratosphere in general cools significantly. Also, the volcanic experiment is listed under the stratospheric 'warming' category even though there is little overall stratospheric temperature change, because it does warm the tropical lower stratosphere (value in parenthesis at 68 mb). The sulfur plus soot experiment produces extreme warming in the lower stratosphere. All of these experiments, including those with only stratospheric forcing, also produce a significant tropospheric response (significant values in bold italics; 95% significance determined by a t-test with respect to interannual variability in the control run).

[14] What happens when the sea surface temperatures are not allowed to respond (i.e., they are fixed at climatological values, Table 1b); does this approach in fact allow us to separate the stratospheric from tropospheric forcing? Restraining the sea surface temperatures reduces the surface temperature response by greater than a factor of 10 in all cases, and the tropospheric response by a similar magnitude except in the case of the volcanic aerosol + soot where the

Table 1b. As in Table 1a, but for the Specified SST Experiments

Experiment	SURF	346–203 MB (D-F 50°–90°N)	Trop	68 MB (16N–16S)	1.5 MB	Strat
2CO2	***0.36***	***0.26*** (−0.7)	***0.23***	***−1.5 (−1.5)***	***−8.6***	***−3.1***
2CO2 Strat	0.0	***0.30*** (0.4)	0.08	−0.4 (1)	***−8.3***	***−1.4***
2H20 Strat	0	−0.1 (0.0)	0.03	***−0.8*** (−1)	***−1.7***	***−1.3***
+2% TSI[a]	***0.16***	***0.4*** (0.0)	***0.25***	0.5 (0)	***1.1***	***0.58***
Volc	***−0.10***	0.1 (0.0)	0	***1.2 (2.2)***	0.0	***0.8***
Sulf+Soot	−0.04	***3.8*** (−0.2)	***1.06***	***36.3 (41)***	***5.9***	***32.0***
−2% TSI[a]	***−0.15***	***−0.3*** (0.8)	***−0.25***	−0.4 (−1)	***−1.1***	***−0.54***
No O3 LS[a]	−0.03	***−0.5*** (−0.6)	−0.06	***−8.8 (−11)***	−0.1	***−4.36***

[a]Experiments in which the temperature changes in the troposphere and stratosphere are of the same sign.

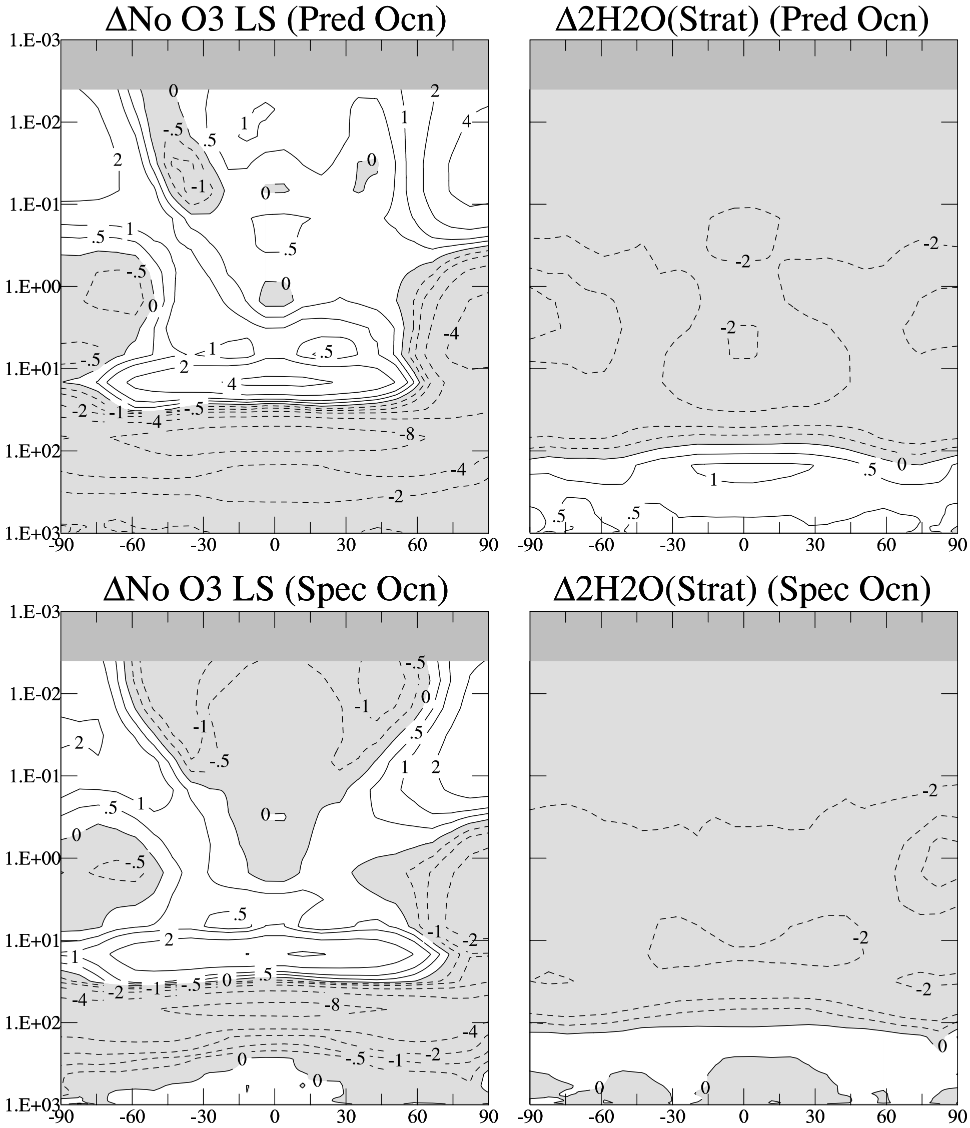

Figure 1. Annual temperature changes for the experiment with no ozone in the lower stratosphere (left) and doubled stratospheric water vapor (right). Top row shows the results with full sea surface temperature (SST) response, and in the bottom row when SSTs are kept at climatological values.

extreme lower stratospheric warming extends down into the upper troposphere. The stratospheric response remains high, and the ratio of tropospheric to stratospheric temperature change is reduced by a factor of 10 with the specified SSTs. From this perspective, the procedure has strongly minimized the tropospheric contribution to climate and thus circulation response.

[15] Examples of the temperature responses to stratospheric climate forcing under the two sets of SST conditions are given in Figure 1. Shown are the annual temperature changes for the experiment with no ozone in the lower stratosphere (left) and doubled stratospheric water vapor (right). The top row shows the response when SSTs are allowed to change; the troposphere features significant temperature changes, although the stratospheric temperature changes are still somewhat larger. The bottom row shows the results when sea surface temperatures are specified at current day values. The complete dominance of stratospheric over tropospheric temperature changes is obvious.

[16] The upper tropospheric response, while generally not large, is often significant globally even in the specified SST experiments. This result raises the epistemological question: if the forcing is only in the stratosphere, but the direct effect extends down into the upper troposphere, does any subsequent surface circulation response count as "stratospheric forcing"? One answer is to determine whether the altered tropospheric circulation is initiated from the stratosphere as opposed to the upper troposphere. For the specified SST experiments, the high latitude extratropical upper troposphere temperature changes for Dec–Feb (values in parenthesis in Table 1b) are not significant. In contrast, most of the temperature changes in the lower stratosphere extratropics are significant, as shown in Figures 2a and 2b (the interannual standard deviation at this altitude from a

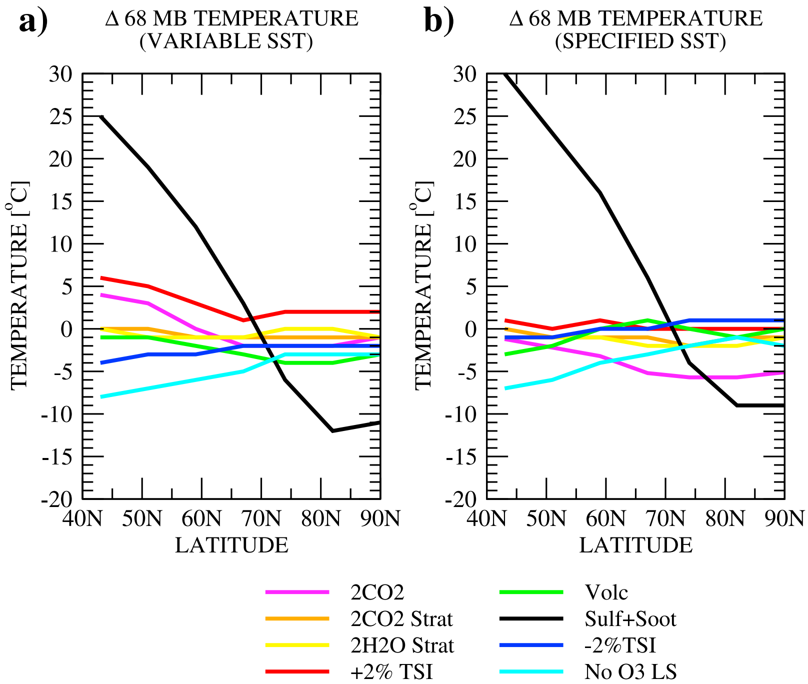

Figure 2. Zonal mean 68 mb temperature change during December-February in the different experiments when SSTs are allowed to respond (Figure 2a) and when they are not (Figure 2b). For ease in interpretation, the green/blue/purple colors refer to experiments in which the global mean temperature has cooled, while the yellow/orange/red colors refer to those in which it has warmed, as demarcated in Tables 1a and 1b.

100 year control run is less than 1°C except right at the pole). Therefore, the extratropical thermal perturbation is being initiated from the lower stratosphere, although as we show below, dynamic responses have much continuity across the tropopause region.

3.2. AO/NAO and NAM Response

[17] To explore the extratropical circulation response to climate change, we use a representation associated with the largest degree of variability, the AO/NAM. We employ as an estimate of the AO/NAM response the anomalous difference in pressure or geopotential height between mid latitudes (30–50°N) and high latitudes (60–80°N) at various levels, an index which is generally similar to AO changes (both qualitatively, and, when normalized by the interannual standard deviation, quantitatively [e.g., *Rind et al.*, 2004]). While the actual meaning of the AO has been debated [e.g., *Ambaum et al.*, 2001; *Christiansen*, 2002; *Dommenget and Latif*, 2002], our use here is simply to characterize the extratropical circulation in a zonal average sense. Similarly, we also report the NAO changes, calculated from the anomalous difference in sea level pressures using nine grid points averaged around the vicinity of the Azores and Iceland, to characterize the atmospheric circulation in the North Atlantic.

[18] Results for the change in these indices at different levels are given in Table 2 (all differences are relative to the respective control run values). About one-half of the experiments produce significant changes in the AO or NAO index when the SSTs are allowed to change, with the global warming experiments (top four rows in Table 2) showing a more positive index, and the global cooling experiments a more negative one. When the SSTs are not allowed to change, the NAO response has a similar proportion of

Table 2. Change in AO-Index (30–50°N Minus 60–80°N) and NAO During December–February

	Variable SSTS				Specified SSTS			
Experiments	SLP/AO (MB)	NAO (MB)	100 MB (M)	10 MB (M)	SLP/AO (MB)	NAO (MB)	100 MB (M)	10 MB (M)
2CO2	***4.47***	***5.77***	***231***	***224***	0.51	***4.95***	***131***	*222*
2CO2 Strat	***1.84***	−1.64	46	53	0.90	1.00	***82***	***259***
2H2O Strat	1.33	−0.10	19	−55	0.76	***3.50***	33	57
+2% TSI[a]	***4.44***	1.84	***180***	***193***	−0.95	−1.16	−21	***326***
Volc	***−3.39***	***−3.44***	−11	***189***	***1.79***	***3.00***	14	−45
Sulf+Soot	−0.30	−1.66	***533***	***1962***	***3.8***	***3.55***	***589***	***1937***
−2% TSI[a]	***−3.84***	***−5.99***	***−74***	66	0.02	0.84	−46	***196***
NO O3 LS[a]	−1.36	***−2.25***	***−94***	79	0.89	0.17	−66	***318***

[a]Experiments in which the temperature changes in the troposphere and stratosphere are of the same sign.

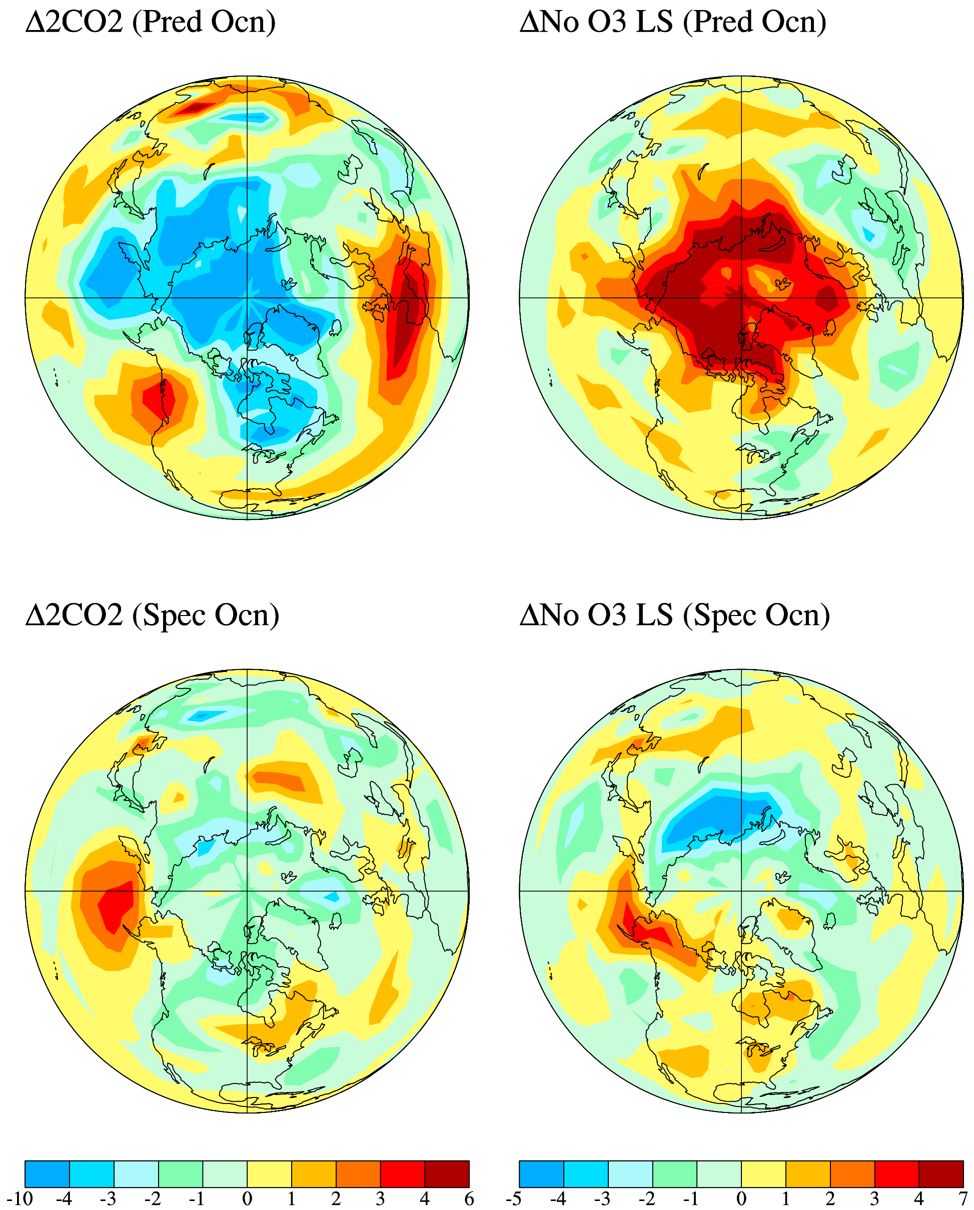

Figure 3a. Sea level pressure changes in December–February for the 2 × CO_2 results (left) and no ozone in the lower stratosphere (right). Results in the top row are for the simulations when SSTs are allowed to change, and in the bottom row when they are not.

significant changes as in the case of variable SSTs (but often of opposite sign), and while the AO responses are in general not significant, they are always of the same sign as the NAO change. The results suggest that in this model, the AO responds more strongly to tropospheric climate changes than to stratospheric ones, and the NAO is more responsive than the AO to stratospheric forcing.

[19] With the tropospheric influence minimized in the specified SST experiments, for the runs in which the lower stratosphere cools at high latitudes (e.g., 2H2O Strat, Sulf + Soot, 2CO2), the index changes are positive. The only experiment included here with a significantly positive temperature response in the extratropical lower stratosphere (+2% TSI) produced a negative index change (although not significant). A positive temperature response in this region was also found in the QBO East – QBO West experiments under solar minimum conditions [*Rind and Balachandran*, 1995], and that experiment also produced a negative index change.

[20] For comparison of the different effects, we show in Figure 3a the Dec-Feb sea level pressure changes for the 2CO2 results (left) and No O3 LS (right); the zonal average changes at most latitudes are significant. (The model's zonal mean standard deviation in the control run is less than 2 mb south of 60°N, rising to near 4 mb near the pole, which is similar to the observed [e.g., *Thompson and Wallace*, 2000].) In the top row, with variable SSTs, the global warming experiment produces lower pressure over the pole, while the global cooling experiment produces higher pressure. When SSTs are not allowed to change (bottom row), in the 2CO2 climate experiment the polar lower stratosphere still cools (Figure 2b) and the index change (especially over

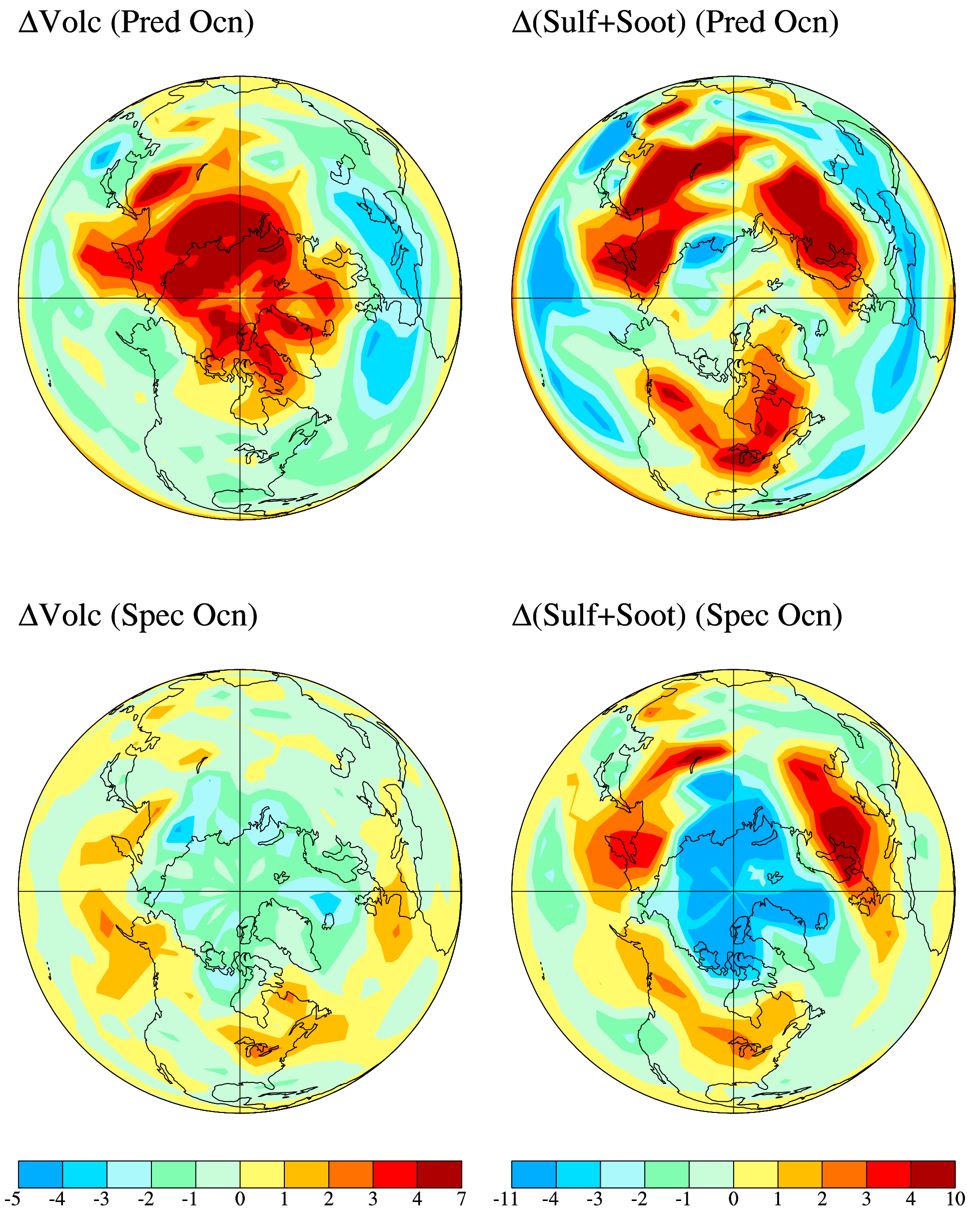

Figure 3b. As in Figure 3a, except for the experiments with volcanic aerosols (left) and a mixture of soot and sulfur aerosols (right).

the North Atlantic) is still positive. However, without the global warming component, the response is smaller. In the No O3 LS experiment with specified SSTs, temperatures are cooler in the lower stratosphere, and the index changes are slightly positive. Yet without the global cooling component, the index changes are not significant. (Note that this is not a 'polar ozone' loss experiment, which strongly alters the latitudinal temperature gradient in the lower stratosphere [e.g., *Gillett and Thompson*, 2003].) These experiments emphasize that with similar magnitudes of global tropospheric temperature response and polar lower stratospheric temperature response, the tropospheric climate change produces the dominant impact on the AO/NAO at the surface.

[21] Given that conclusion, we can understand what happens if the tendencies induced by the tropospheric climate change and stratospheric temperature change are opposed, for example a global cooling climate (more negative AO tendency) with cooling in the polar stratosphere as well (more positive AO tendency). The experiments with added volcanic (sulfur) aerosols and added sulfur/soot aerosols help to illustrate the net effect; results are shown in Figure 3b. Both forcings produce colder climates in the troposphere (Tables 1a and 1b) especially when SSTs are allowed to change. However, with variable SSTs only the volcanic aerosol experiment produced the higher pressure over the pole and more negative AO/NAO indices. When SSTs are specified (hence unchanging), the index change becomes positive in the volcanic aerosol experiment. This is consistent with observations of the first winter response after a tropical volcano; by the second winter, the associated warming over land at high latitudes is weaker, and it is gone by the third winter [*Robock and Mao*, 1995], as SSTs cool (and the stratospheric latitudinal temperature gradient weakens). With SSTs specified in the

sulfur/soot experiment, a much more positive AO response arises. This latter effect is consistent with the large negative cooling in the lower stratosphere associated with that experiment. This is a result, as we shall see, of altered planetary wave propagation so that much less energy is advected poleward in the lower stratosphere; the northward transport of sensible heat in the lower stratosphere drops by 33%. This stratospheric polar cooling influence cancels out the tropospheric cooling influence in the sulfur/soot variable SST experiment. The excessive nature of the stratospheric temperature change in the sulfur/soot experiment (or, mechanistically, the large planetary wave refraction response) allows it to successfully 'compete' with the tropospheric climate change. Appropriately, the tropospheric warming experiment which also features cooling in the lower stratosphere ($2CO_2$) has the largest index change, presumably aided by both factors working together.

[22] Our results on the climatic time scale are consistent with those of *Black and McDaniel* [2004] for 10 day time-scales. They noted that the ability of stratospheric potential vorticity anomalies to influence the troposphere was limited when the troposphere had anomalies of the opposite sign; the stratospheric influence was still operating as a tendency, but it could not overcome the tropospheric anomaly.

3.3. Relationship Between the Responses at Different Levels

[23] One aspect of understanding whether the stratosphere is influencing the troposphere is to determine how events at one level are related to those above or below.

[24] On short time scales, *Baldwin et al.* [2003] found the most significant (but only occasional) relationship between NAM changes just above the tropopause ($\sim$150 mb) and AO changes in sea level pressure; does such a relationship exist for the climate change experiments in general? It will, if they are forced by similar processes, or if the stratospheric change is forcing a tropospheric response. In the variable SST experiments, the change in the index that occurs at the 100 mb level (Table 2) is of a similar nature to that change at the surface, with a correlation of about 0.9 between the levels (99% significant) considering either the AO or NAO. With specified SSTs, while the changes at the two levels are generally of similar sign, the correlation is significant only for the NAO (0.7). The difference between these results suggests that the tropospheric climate changes are influencing the lower stratospheric circulation index in a similar manner to their influence on the high latitude surface index. *Zhou et al.* [2002] and *Black and McDaniel* [2004] noted that negative anomalies (weak vortex) were more likely to propagate from the lower stratosphere to the troposphere on the short time-scales, but the results in Table 2 do not show any such discrimination on climatic time-scales.

[25] *Perlwitz and Graf* [1995], *Deser* [2000], and *Charlton et al.* [2003] reported that the interannual relationship between stratospheric AO-like indices and the troposphere was most consistent with respect to Atlantic storm tracks, while *Ambaum and Hoskins* [2002] emphasized the NAO-stratospheric relationship on synoptic time-scales. The relationship with the synoptic situation over the Pacific appeared to be more variable, in part because of ENSO influence on the Aleutian Low [e.g., *Zhang et al.*, 1997], and as this component contributes to the AO, the NAO appeared to provide a better representation of tropospheric response. In the experiments performed here, this result was obviously true for the specified SSTs experiments and again suggests that the sea level pressure field over the Atlantic may be more responsive to stratospheric climate perturbations. The relationship between Atlantic and Pacific SST changes and the NAO is explored further in part 2.

[26] *Baldwin and Dunkerton* [1999, 2001] discuss a downward propagation of the stratospheric circulation changes from the middle to lower stratosphere. Their results were for daily to weekly time-scales; for the climate changes resulting from the perturbations used in this model, there is no significant correlation between middle stratosphere AO-like changes and either the AO or NAO at the surface (and for the specified runs, the correlation is negative). One reason for this is that many of the perturbations are associated with heating or cooling directly in the lower stratosphere which then changes the strength of the polar vortex at levels above 100 mb. An example of this effect is the "no ozone" experiment in the lower stratosphere, especially with specified SSTs. Reduced thermal absorption due to the ozone removal cooled the lower stratosphere at all latitudes, while the mid and upper stratosphere warmed due to both increased thermal absorption (from longwave energy no longer absorbed in the lower stratosphere), and reduced tropical upwelling. This latter effect only extended to mid-latitudes; its counterpart at high latitudes, reduced downwelling then resulted in cooling. The combination of lower and mid-stratospheric cooling at higher latitudes converted a weaker vortex at 100 mb to a stronger vortex by 10 mb (and hence changed a negative AO-index to a positive one) due to its impact on the local thickness, not via wave propagation anomalies. In fact, for the experiments with moderate climate perturbations in the lower stratosphere (Volc, NO O3 LS, 2H2O Strat) there is no significant correlation between the AO-like changes in the lower stratosphere and those in the middle stratosphere with either variable or specified SSTs. However, for some experiments where the forcing is primarily at the surface, e.g., 2CO2 and +2% TSI, the AO-like change is more barotropic in nature within the stratosphere in this model.

3.4. AO Changes and Eddy Momentum Transports

[27] AO changes result in increases (positive phase) or decreases (negative phase) in zonal winds at the higher extratropical latitudes, and so they are thought to be associated with changes in angular momentum transport. The difference in eddy momentum transport is shown as a function of altitude from 950–150 mb (Figures 4a and 4b). The eddy northward transport of angular momentum is calculated as equal to $(v'M)(2\pi r \cos \phi)(u'r \cos \phi)$, i.e., the northward transport of mass per unit area (first parenthesis) through the circumference of the specific latitude circle (second parenthesis) multiplied by the zonal velocity times the moment arm (third parenthesis). The term itself has units of kg m^2 s^{-2}, hence joules. Results are shown for the Northern Hemisphere average; values were also calculated for specified latitudinal bands, with results similar to those discussed below.

[28] As expected, when the AO change was positive, angular momentum transport (and ultimately convergence) increased producing stronger west wind flow, while a negative AO change resulted from decreased momentum trans-

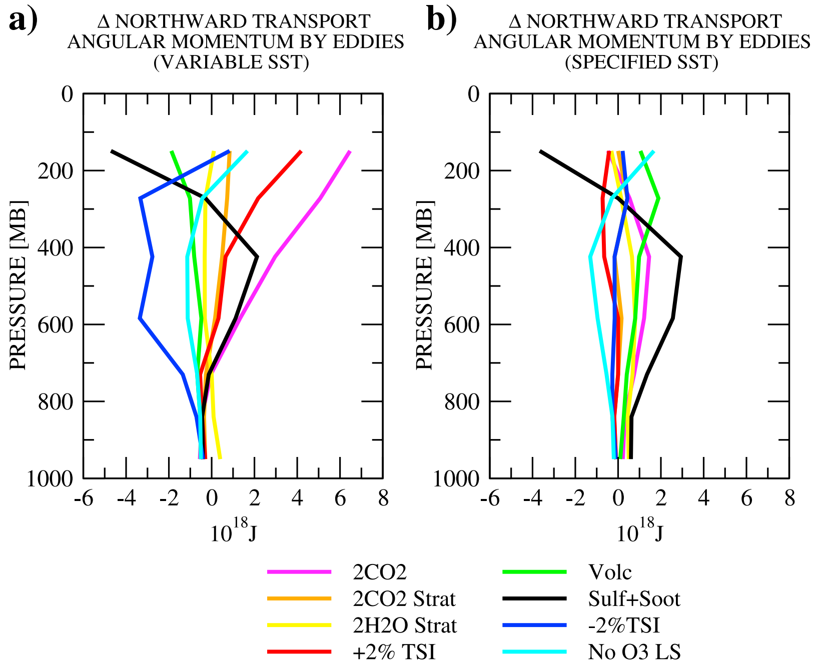

Figure 4. Change in Northern Hemisphere eddy transports of momentum in December–February as a function of altitude in the different experiments with variable (Figure 4a) and specified (Figure 4b) SSTs. For reference, the vertically integrated change in the −2%TSI experiment (variable SSTs) was about 20% of the current climate value.

port. Overall, the change in (vertically-integrated) eddy transport of angular momentum in the different experiments correlates with the change in AO at the surface at 0.7, and with the change in NAO at 0.64, both significant at 99%. The correlation is even higher between the vertically-integrated change in the northward transport of quasi-geostrophic potential vorticity (QGPV) and these two indices (0.85 and 0.76, respectively); QGPV represents an approximation of the total eddy forcing, the EP flux divergence, due to both eddy momentum transport convergences and the change in eddy sensible heat flux with altitude. For these experiments the eddy transports of angular momentum and QGPV are themselves highly correlated (correlation coefficient of 0.9).

[29] It has been suggested that the "stratospheric" influence is primarily in the upper troposphere; at what level was the altered transport occurring in these experiments? With the specified SSTs (Figure 4b), the changes peak in the region between 450 and 200 mb in 7 of the 8 experiments, as if the stratosphere is influencing that region the most. With variable SSTs (Figure 4a), when the lower troposphere has more freedom to adjust, there is less uniformity in the level of maximum transport change (only 4 of the 8 clearly peak in the upper troposphere), consistent with a greater influence of the troposphere as a whole, including the levels below 500 mb. Nevertheless, in both cases the transport changes often have strong vertical consistency, indicating that the altered meridional wave propagation, implied by these momentum transport changes, appears to be occurring throughout the troposphere.

3.5. Eddy Momentum Transport Changes and Eddy Energy

[30] The angular momentum transport changes result either from altered eddy energy and/or from altered wave propagation. Eddy energy is $(\frac{1}{2}[(u')^2 + (v')^2])$, where u′ and v′ represent deviations of the zonal and meridional wind from the zonal average, and the values used here represent both transient plus stationary eddy components. Changes in eddy energy will affect the momentum transport because the control run (and observations) feature poleward transport of angular momentum at mid-latitudes, and equatorward transport at the highest latitudes, resulting in angular momentum convergences at upper mid-latitudes. Shown in Table 3 are the differences in both total and longwave tropospheric eddy energy in the different experiments. With variable SSTs, the warming experiments often have decreases in total eddy energy, consistent with the idea that high latitude amplification of the warming magnitude will reduce the latitudinal temperature gradient and baroclinic instability. The cooling experiments have increased eddy energy via changes in the same processes. This aspect is verified in Table 3, with values in parenthesis showing the

Table 3. Change of Northern Hemisphere Winter Tropospheric Eddy Kinetic Energy[a]

	Variable SSTS		Specified SSTS	
Experiments	TROP EKE	WAVE#1–4	TROP EKE	WAVE#1–4
2CO2	***−3.3 (−3.3)***	***12 (34)***	***1.8 (1.0)***	***5.6 (3.0)***
2CO2 Strat	***−1.4*** (−0.5)	−1.3 ***(6.2)***	***−2.8 (−5.4)***	***−3.3(−7.2)***
2H2O Strat	***−3.1 (−3.2)***	***−5 (−4.9)***	***−1.7 (1.5)***	−2 (2.0)
+2% TSI[b]	***−6.9 (−1.6)***	***7.4 (29)***	0.3 ***(−1.1)***	−0.4 (−0.4)
Volc	−0.3 (0.9)	***−6.1 (−15)***	−0.9 ***(−5.1)***	0.3 ***(−8.6)***
Sulf+Soot	***−12.7 (−7.4)***	***−13.3 (−12)***	***−13.8 (−13.1)***	***−10.5(−20.6)***
−2% TSI[b]	***1.5 (2.8)***	***−7.1 (−16)***	0.7 ***(−1.1)***	0.8 (−0.4)
NO O3 LS[b]	***6.1 (1.9)***	***4.3 (3.7)***	***3.4*** (0.2)	***5.0 (14.9)***

[a]Change is in percent. Changes in baroclinic energy conversion (%) are shown in parenthesis.
[b]Experiments in which the temperature changes in the troposphere and stratosphere are of the same sign.

changes in eddy potential to kinetic energy conversion, generally decreasing in the warming experiments and increasing in the cooling experiments.

[31] Also given in Table 3 is the change in longwave energy (planetary waves 1–4), and the result for variable SSTs is now very different. There is no consistency between the sign of the climate change and either the sign of the planetary wave energy change or the change in baroclinic energy generation. The alteration of the latitudinal temperature gradient due to high latitude amplification of climate warming is basically a low-altitude phenomenon. It is associated partly with the differences in stability between low and high latitudes; as low latitudes are able to distribute warming to greater altitudes, the change in latitudinal temperature gradient often reverses sign by the upper troposphere, an effect which is felt more strongly by planetary waves with their greater vertical and latitudinal reach, than by synoptic scale waves. Hence the changes in total tropospheric eddy kinetic energy (EKE) and planetary wave energy for waves 1 to 4 are uncorrelated in the variable SST experiments, with the planetary wave energy changes often being larger percentage-wise. This is one reason why warming experiments with the GISS model can have reduced tropospheric eddy energy yet increased stratospheric eddy energy [e.g., *Rind et al.*, 1998], as the planetary wave change in the troposphere propagates into the stratosphere, altering its dynamics.

[32] For the specified SST experiments, in which changes in the latitudinal gradient in the troposphere are minimal, stratospheric forcing appears to play more obvious a role, and now there is little difference between planetary and shorter-scale waves in their qualitative or quantitative response. In experiments in which the lower stratosphere is warming, stability increases in the upper troposphere, and eddy energy decreases; conversely, lower stratospheric cooling is associated with eddy energy increases. Specific examples of this relationship are presented in Figure 5 using the specified SST experiments that have the strongest temperature response in the extratropical lower stratosphere. Results are shown for the no ozone in the lower stratosphere and the sulfate/soot experiment. Removing the ozone cools the lower stratosphere at mid-latitudes (Figure 5a, shown for 50°N); associated with the reduction in vertical stability, baroclinic generation increases in the Northern Hemisphere between 900 and 200 mb (Figure 5b), and Northern Hemisphere (total) eddy energy increases throughout the troposphere (Figure 5c). In contrast, in the sulfate/soot experiment simulation, temperatures increase in the upper troposphere/lower stratosphere, stability increases, baroclinic generation decreases, as does eddy energy.

[33] It is important to note that even though these changes are initiated by forcing in the lower stratosphere, the temperature response extends down into the upper troposphere (Figure 5), due to a combination of advection and radiation processes. The effect on baroclinic instability extends even lower in the troposphere, influencing eddy energy at most levels, including the planetary wave energy. As calculated by *Saltzman* [1970], planetary waves obtain significant energy through the baroclinic process.

[34] The variation of eddy kinetic energy with altitude in the different experiments is presented in Figures 6a and 6b. With the variable SSTs (Figure 6a), the changes at altitudes in the low and mid-troposphere are occasionally of opposite sign to those in the upper troposphere where the largest planetary wave responses are dominating (for example, in the 2 × CO2 experiment, eddy energy decreases from the surface to 300 mb, then increases above). In the specified SST runs (Figure 6b), with all wavelength eddies respond-

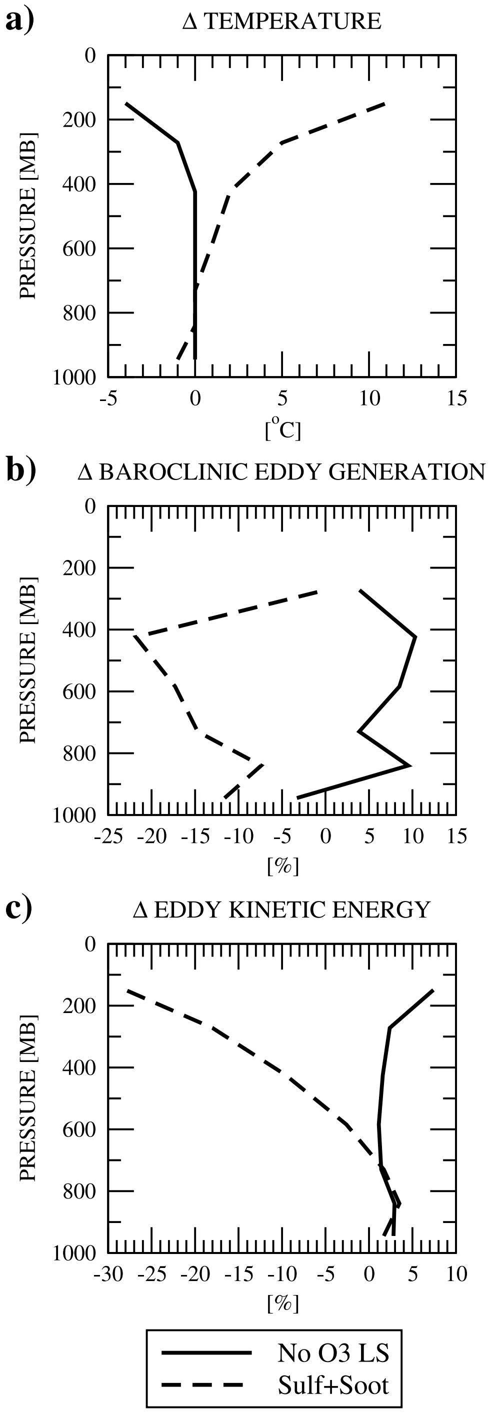

Figure 5. Change in December-February temperature at 51°N (Figure 5a), in baroclinic energy generation (Figure 5b) and in eddy kinetic energy (Figure 5c) for the "no ozone in the lower stratosphere" experiment (solid line) and the sulfur/soot experiment (dashed line). Both experiments use specified SSTs and hence emphasize stratospheric forcing.

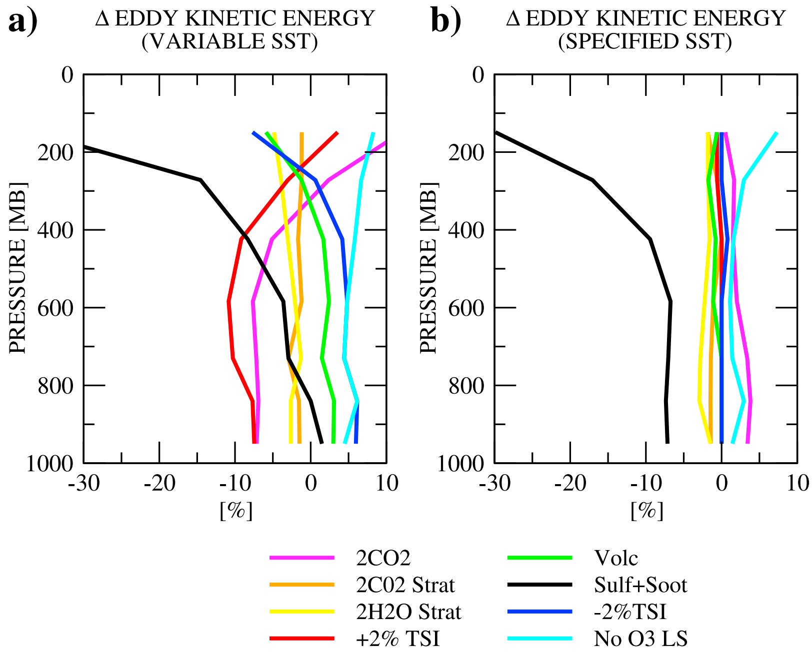

Figure 6. As in Figure 4, but for the variation of eddy kinetic energy with altitude.

ing similarly, the changes with altitude show a great deal of consistency; for these runs, the change in total tropospheric EKE and the change for planetary waves 1 to 4 are highly correlated (0.9, >99% significance).

[35] Do these eddy energy changes help account for the angular momentum transport changes? Using the Northern Hemisphere eddy angular momentum transport and total eddy energy values at each level in the troposphere for each experiment, we correlate the change in angular momentum transport with the change in total eddy energy. In both the variable and specified SST runs, the correlation is significant at greater than the 95% level; however, with a correlation coefficient of about 0.25 (similar in both sets of experiments), the eddy energy change accounts for only about 6% of the variance. When a similar correlation is performed with respect to the change in planetary wave energy the correlation in the variable SST runs is 0.46, accounting for about 20% of the variance; there is no significant correlation in the specified SST runs. Hence the impact of climate change in the troposphere on planetary wave energy has a significant influence on the angular momentum transport. In contrast, stratospheric forcing affects shorter wavelength energy in the troposphere in a similar manner to planetary scale waves, with a minimal impact on angular momentum transport.

3.6. Eddy Momentum Transport Changes and Wave Propagation

[36] How is the propagation of planetary wave energy altered in the various experiments? The change in eddy angular momentum transport was shown in Figure 4; this is proportional to the meridional propagation of wave energy. The absolute magnitude of the meridional propagation can change due to alterations in the background wind structure, or simply due to a change in wave energy propagating up to a particular altitude. We normalize for this latter effect, i.e., differences in the vertically propagating planetary wave energy, by dividing the eddy momentum transports by the eddy northward heat transport, a quantity which is approximately proportional to the upward wave energy flux [e.g., *Holton*, 1992, p. 323]. In terms of the northward and vertical Eliassen-Palm flux (Fy, Fz), which are proportional to the meridional and vertical zonal-average wave energy flux, the quasi-geostrophic approximation is

$$F_y = -\rho\overline{u'v'}, \quad F_z = \rho f R\overline{v'T'}/\left(N^2 H\right)$$

where T′ is the temperature perturbation, N the Brunt-Vaisala frequency, R the dry air gas constant, H the scale height, f the coriolis parameter, and ρ the density. By using the sensible heat transport to normalize for changes in upward energy propagation, we are calculating the meridional 'turning' relative to the upward energy flux.

[37] The change in the ratio of this northward to upward propagation, or refraction, is shown in Figures 7a and 7b. Positive values for both eddy heat and momentum transport represent planetary wave energy propagating upward and equatorward, so a value greater than one indicates increased propagation from the extratropics to low latitudes relative to the amount of energy propagating upwards.

[38] Two opposing tendencies are evident in Figure 7. Experiments that cool the troposphere tend to produce more poleward wave refraction (e.g., with variable SSTs: −2%TSI, No O3, Volc), while global warming experiments result in more equatorward refraction (with 2CO2 and +2%TSI). Experiments that cool the tropical stratosphere more than other latitudes (or warm the polar stratosphere relative to other latitudes) produce more poleward wave refraction (e.g., with specified SSTs, No O3) or alternatively, with excessive warming of the tropical lower stratosphere/cooling of the polar lower stratosphere, more equatorward refraction (Sulfur + Soot). This outcome results from the effect of the latitudinal temperature gradient change on wave energy propagation. With an increased latitudinal

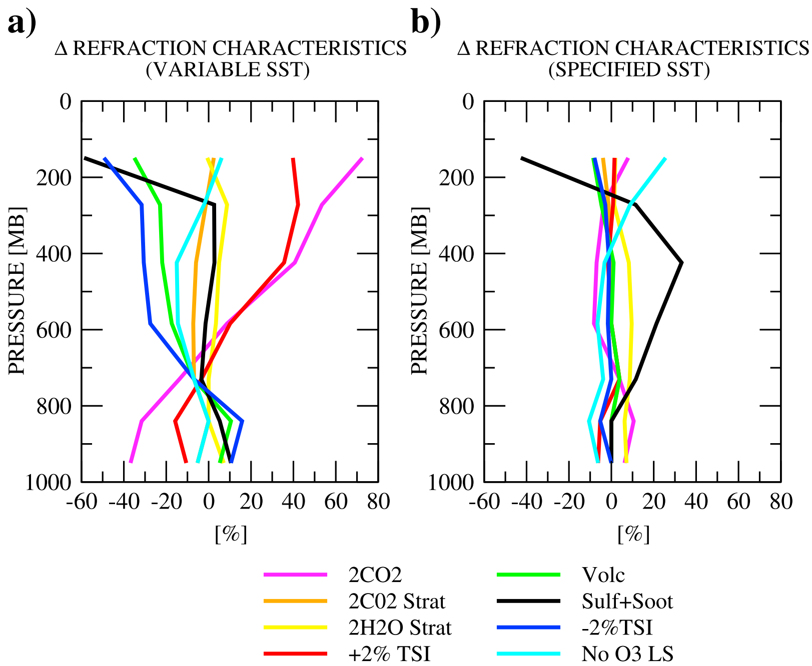

Figure 7. As in Figure 4, but for the change in eddy momentum transport normalized by the change in eddy sensible heat transport, as an indication of the change in meridional refraction characteristics of the atmosphere. A more positive value indicates more equatorward wave propagation.

gradient, from the thermal wind relationship, west winds increase at higher elevations; this has the effect of maximizing the zonal winds in the region that they are normally a maximum (mid-latitudes near the tropopause), which increases the second derivative of the zonal wind shear both latitudinally and longitudinally. It increases the refractive index of the atmosphere and planetary wave energy is preferentially reflected equatorward. In the global cooling experiments, in this model the tropical upper troposphere cools strongly, associated with a significant tropical SST sensitivity and transference of that response to higher levels in the troposphere. This reduces the latitudinal temperature gradient, produces an east wind tendency, which alters the latitudinal gradient of potential vorticity in such a way as to promote more poleward wave refraction (and a more negative AO phase). The change in latitudinal temperature gradient and hence background wind can be initiated in the stratosphere as well, as occurs when the tropical stratosphere warms, increasing the latitudinal gradient just above the tropopause (in the tropics), increasing west winds in the lower stratosphere, and promoting more equatorial wave refraction (and a more positive AO phase).

[39] Again, the tropospheric and stratospheric effects often work against one another: the stratospheric soot experiment with variable SSTs (Figure 3b, top) has more equatorward refraction despite its tropospheric cooling, because the magnitude of lower stratospheric tropical warming was so extreme (Tables 1a and 1b). With the normal magnitude of stratospheric temperature changes, of a few °C, variations of that magnitude in SSTs end up dominating the refraction change, as in the case of the volcanic aerosols with variable SSTs, which now has more poleward propagation.

[40] In comparing the changes in angular momentum transport to these changes in refraction characteristics, one would expect that the correlation would be positive: greater refraction toward low latitudes would be associated with greater poleward momentum transport, unless the changes in eddy energy intervened. We again correlate the two phenomena, in this case, relating the change in angular momentum transport at each altitude with the change in refraction characteristic. The resulting correlation is positive for both sets of experiments, significant at the 99% level. With specified SSTs the change in refraction characteristics accounts for about 50% of the variance in the momentum transports, while with variable SSTs it accounts for about 36% (hence to the extent that the correlations are not higher, the eddy energy changes are important; we show in part 2 that surface conditions at high latitudes also influence the result). Adding these results to those of the previous section, we find that the sum of energy and refraction changes thus accounts for 56% of the observed variance in both sets of experiments. With stratospheric forcing (specified SSTs) it is predominantly due to refraction changes, while when the tropospheric climate is altered, both energy and refraction changes are important. Note that changes in the latitudinal temperature gradient affect both wave energy generation and propagation, in the sense that an increased gradient will produce more eddy energy as well as more equatorward wave refraction. Both then end up increasing angular momentum transport poleward from mid-latitudes, leading to a more positive AO phase.

4. Discussion

4.1. Evaluation of the Different Mechanisms

[41] Of the mechanisms discussed in the introduction concerning how the stratosphere might influence the tropospheric response, clearly wave-mean flow interaction is the dominant response seen from these experiments. The largest

influence relates to the change in the refraction properties of the atmosphere, which affects all waves. With respect to the impact on eddy energy, when the tropospheric climate is changing in the variable SST experiments, the higher correlation between the change in momentum transport (and AO index) with the change in planetary wave energy (waves 1 to 4) (correlation coefficient of 0.46), compared with the change in total EKE (0.25) indicates that it is not the synoptic-scale waves that are influencing the results. In contrast, with the stratospheric forcing and specified (unchanging) SSTs, there is a weak effect due to the influence on eddy energy in general, but not in particular with the largest scale waves. Therefore the model does suggest that synoptic scale wave energy appears to be responding to the stratospheric forcing, although this impact on the AO is small. The coarse grid model used for these experiments might not be the best tool for determining the full magnitude of synoptic-scale influence.

[42] Can we distinguish, on the climate time scales, between the so-called ''indirect mechanism'', in which stratospheric changes are altering planetary wave energy propagating up from the troposphere, hence affecting the tropospheric circulation, and the ''direct mechanism'', in which the stratospheric vortex is being directly affected by altered wave propagation, and producing circulation anomalies in the troposphere via non-local hydrostatic and geostrophic adjustment? The basic question concerns the level at which the altered wave energy fluxes are occurring. As shown in Figure 4, much of the change in eddy transports of angular momentum is occurring within the troposphere, which would suggest that the ''indirect forcing'' is the dominant mechanism. *Black and McDaniel* [2004] lump the upper troposphere and lower stratosphere together for the ''direct effect''; with that interpretation, the stratospheric forcings (runs with specified SSTs), by having their peak angular momentum transport change in the upper troposphere (Figure 4a), can be said to be 'directly forcing' the vortex near the tropopause. There is also often a vertical consistency, so that transport changes are occurring in a similar fashion in the upper troposphere and lower stratosphere, obscuring the difference between the two mechanisms.

[43] However, the ''no ozone'' experiment with specified SSTs provides an interesting example. Here the reduced poleward momentum transport in the troposphere should have produced a more negative AO/NAO response, while the increased transport tendency in the lower stratosphere would have given a more positive response. The net result was a slightly more positive (though not significant) response. As in the daily time-scale cases studied by *Black and McDaniel* [2004], the direct stratospheric forcing mechanism appeared to be acting, but its tendency was opposed by the tropospheric forcing, resulting in an insignificant change. A similar example is that of the sulfur/soot aerosols with variable SSTs, where again the tropospheric and stratospheric influences are acting in opposite directions, and the stratospheric effect (forcing a negative AO/NAO index change in this case) predominates but with a non-significant net result.

4.2. Comparison With Observations and Other Models

[44] Is there any evidence that the model results are at all related to how the actual atmosphere AO/NAO responds? *Shindell et al.* [2001a] reviewed some of the results with respect to observations, and we expand upon it here. Concerning forcing in the lower stratosphere, the more positive phase of the NAO when volcanic aerosol eruptions occur (without allowing for the SST response), agrees well with many observations [e.g., *Robock and Mao*, 1995] indicating a more positive NAO phase in the first winter following a tropical eruption (i.e., before there has been substantial cooling of the system), and with modeling studies of the same phenomenon [e.g., *Stenchikov et al.*, 2002, 2004]. The volcanic experiments of *Graf et al.* [1993], while run for 60 years in a perpetual January mode, kept the ocean temperatures fixed, thus their positive AO response was consistent with the specified SST runs here. The positive AO phase change associated with cooling of the polar lower stratosphere is consistent with observations of the effect of the ozone hole change on the AAO phase and a modeled response [*Thompson and Solomon*, 2002; *Gillett and Thompson*, 2003]. The result in the ozone experiment conducted here is relatively weak, but the reduced ozone was not confined to the polar region, and this study focuses on the Northern Hemisphere with greater tropospheric planetary wave energy, so the comparison is not perfect. (Since the latitudinal temperature gradient in the lower stratosphere affects extratropical circulation via its influence on planetary wave propagation and angular momentum transport, the 'indirect mechanism' may be less effective in forcing a high latitude response in the Southern Hemisphere.) The stronger relationship between stratospheric forcing and the NAO than with the AO, is in agreement with observations of their interannual relationship [*Perlwitz and Graf*, 1995; *Deser*, 2000; *Charlton et al.*, 2003]. And the relatively weak influence of the stratosphere compared to the troposphere with the typical magnitude of temperature responses is in agreement with results found for the short time-scales, for which *Charlton et al.* [2003] concluded that the stratospheric contribution forced only about 5% of the variance of the 1000 mb AO anomalies.

[45] Concerning tropospheric climate changes, *Shindell et al.* [2001b] showed how the estimated reduced solar irradiance during the Maunder Minimum generated a more negative AO/NAO phase, in approximate agreement with implications from the paleo-temperature record, and *Rind et al.* [2004] showed how the colder climates for the Maunder Minimum and Little Ice Age in general produced a negative NAO phase, again in agreement with reconstructions [*Luterbacher et al.*, 1999]. The more positive AO phase in the Northern Hemisphere associated with greenhouse warming in the model is in approximate qualitative and quantitative agreement with the observed trends (0.8 mb/decade in the model compared with 1 mb/decade in observations) [*Shindell et al.*, 1999; *Thompson et al.*, 2000], including the increased west winds in the stratosphere and altered planetary wave propagation [*Hartmann et al.*, 2000]. However, as noted in the introduction, the effect of increasing trace gases and global warming on modeled circulation anomalies at high latitudes is inconsistent [*Butchart et al.*, 2000; *Zorita and Gonzalez-Rouco*, 2000; *Gillett et al.*, 2002]. In part 2 of this study, we discuss other influences on the AO/NAO that might result in such disagreements, such as the

magnitude of temperature response at high and low latitudes and in the tropical upper troposphere.

5. Conclusions

[46] In this study we explore the influence of stratospheric-induced climate changes on the AO/NAO relative to that of tropospheric climate changes. We utilize 8 different climate change experiments, each run with both varying SSTs to allow for a full tropospheric climate response, and with specified (unchanging) SSTs to focus the primary forcing in the stratosphere. The primary results from this study are as follows:

[47] 1. Both tropospheric and stratospheric forcings contribute to the the AO/NAO change.

[48] 2. The AO responds more strongly to tropospheric climate forcing, and the NAO is more responsive than the AO to stratospheric forcing.

[49] 3. In global warming experiments, the AO and NAO phase change is positive, while it is negative in global cooling experiments.

[50] 4. When the lower stratosphere cools at high latitudes, the index changes are positive, and they are negative when it warms.

[51] 5. If the tropospheric and stratospheric forcings tend to produce opposing tendencies, given the usual magnitude of climate changes, the tropospheric response dominates.

[52] 6. The AO/NAO index changes are closely related to changes in eddy transports of angular momentum, peaking in the upper troposphere from stratospheric forcing, but throughout the troposphere when tropospheric climate changes occur.

[53] 7. Climate changes affect both eddy energy generation and planetary wave refraction; both influence the eddy angular momentum transport.

[54] 8. With tropospheric climate changes, refraction effects account for 36% of the variance of eddy angular momentum transport, while altered planetary wave energy (in waves 1 to 4) is responsible for 20%.

[55] 9. With stratospheric climate changes, refraction effects account for 50% of the variance of eddy angular momentum transport, while altered (primarily synoptic-scale) eddy energy, due to the change in vertical stability, accounts for only 6%.

[56] 10. Experiments that cool the troposphere, in particular the tropical upper troposphere, tend to produce more poleward wave refraction (with equatorward angular momentum transport), while warming experiments produce the reverse effect.

[57] 11. Experiments that warm the tropical lower stratosphere produce more equatorward wave refraction; again the tropospheric and stratospheric changes can work against one another.

[58] 12. Some of the angular momentum forcing is directly within the troposphere and some is in the stratosphere, affecting the polar vortex directly; however, there is often vertical consistency obscuring the difference between the "direct" and "indirect" mechanisms for influencing the surface circulation.

[59] 13. The model results are in general agreement with observations for climate and inter-annual time-scales, and with most, but not all, other model simulations.

[60] As emphasized by this last point, while these experiments were specifically tailored to discuss 'climate' forcings, the specified SST runs are also relevant to the interannual time-scale for which SSTs have not had sufficient time to adjust to radiative perturbations. Results from the QBO/solar UV experiments discussed in *Rind and Balachandran* [1995] also fit in this category, e.g., the difference between solar maximum and solar minimum during the east phase of the QBO results in a positive AO index, and the difference between the east and west QBO phases during solar minimum produce a negative AO phase (at the surface and at 100 mb), a result also reported from observations [*Ruzmaikin and Feynman*, 2002]. These additional experiments illustrate that the stratospheric forcing can occur from higher levels, working all the way down via wave-mean flow interaction (the momentum transport changes associated with the QBO/solar UV experiments occurred in the upper troposphere).

[61] Not all models produce the AO/NAO responses seen here in the global warming/cooling experiments. Given the likely dominance of tropospheric climate change over stratospheric forcing shown in these experiments, we may expect at least part of the difference to be related to the particulars of the climate changes generated in the individual models. The relative influence of low latitude and high latitude climate change on the AO/NAO is the subject of the second part of this study.

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