Arctic Oscillation response to the 1991 Mount Pinatubo eruption: Effects of volcanic aerosols and ozone depletion

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[1] Observations show that strong equatorial volcanic eruptions have been followed by a pronounced positive phase of the Arctic Oscillation (AO) for one or two Northern Hemisphere winters. It has been previously assumed that this effect is forced by strengthening of the equator-to-pole temperature gradient in the lower stratosphere, caused by aerosol radiative heating in the tropics. To understand atmospheric processes that cause the AO response, we studied the impact of the 1991 Mount Pinatubo eruption, which produced the largest global volcanic aerosol cloud in the twentieth century. A series of control and perturbation experiments were conducted with the GFDL SKYHI general circulation model to examine the evolution of the circulation in the 2 years following the Pinatubo eruption. In one set of perturbation experiments, the full radiative effects of the observed Pinatubo aerosol cloud were included, while in another only the effects of the aerosols in reducing the solar flux in the troposphere were included, and the aerosol heating effects in the stratosphere were suppressed. A third set of perturbation experiments imposed the stratospheric ozone losses observed in the post-Pinatubo period. We conducted ensembles of four to eight realizations for each case. Forced by aerosols, SKYHI produces a statistically significant positive phase of the AO in winter, as observed. Ozone depletion causes a positive phase of the AO in late winter and early spring by cooling the lower stratosphere in high latitudes, strengthening the polar night jet, and delaying the final warming. A positive phase of the AO was also produced in the experiment with only the tropospheric effect of aerosols, showing that aerosol heating in the lower tropical stratosphere is not necessary to force positive AO response, as was previously assumed. Aerosol-induced tropospheric cooling in the subtropics decreases the meridional temperature gradient in the winter troposphere between 30°N and 60°N. The corresponding reduction of mean zonal energy and amplitudes of planetary waves in the troposphere decreases wave activity flux into the lower stratosphere. The resulting strengthening of the polar vortex forces a positive phase of the AO. We suggest that this mechanism can also contribute to the observed long-term AO trend being caused by greenhouse gas increases because they also weaken the tropospheric meridional temperature gradient due to polar amplification of warming. *INDEX TERMS:* 1620 Global Change: Climate dynamics (3309); 3319 Meteorology and Atmospheric Dynamics: General circulation; 3309 Meteorology and Atmospheric Dynamics: Climatology (1620); 3362 Meteorology and Atmospheric Dynamics: Stratosphere/troposphere interactions; 8409 Volcanology: Atmospheric effects (0370); *KEYWORDS:* aerosol, ozone, climate, Pinatubo, Arctic Oscillation

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1. Introduction

[2] It has been long known that strong volcanic eruptions cause radiative, chemical, dynamic, and thermal perturbations in the climate system [*Humphreys*, 1913, 1940; *Mitchell*, 1961; *Lamb*, 1970; *Toon and Pollack*, 1980]. One phenomenon that has generally been observed in the two boreal winters following major explosive eruptions is an anomalously positive phase of the Arctic Oscillation (AO) [*Groisman*, 1992; *Robock and Mao*, 1992, 1995; *Perlwitz and Graf*, 1995]. The AO, defined

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as the first hemispheric empirical orthogonal function of sea level pressure variability [*Thompson and Wallace*, 1998, 2000], is closely related to the North Atlantic Oscillation (NAO) index [*Walker*, 1924; *Walker and Bliss*, 1932; *Rossby*, 1939; *Hurrell*, 1995]. To study this phenomenon we simulated the impact of the Mount Pinatubo eruption on 15 June 1991 in the Philippines, which produced the largest and best observed global volcanic aerosol cloud in the twentieth century. We used the comprehensive stratosphere-troposphere-mesosphere SKYHI general circulation model (GCM) and focused on the atmospheric response to aerosol and ozone forcings, which involves large-scale stratosphere-troposphere dynamic interaction. By examining the mechanisms by which the aerosols and other forcings affect the atmosphere, we also aim to better clarify the AO mechanisms that play an important role in climate trends and variability [*Hurrell and Van Loon*, 1997; *Kodera and Koide*, 1997; *Thompson et al.*, 2000].

[3] Mt. Pinatubo erupted about 20 Mt of SO_2 into the lower stratosphere [*McCormick and Veiga*, 1992; *Bluth et al.*, 1992; *Stowe et al.*, 1992; *Lambert et al.*, 1993], which in a few weeks was converted into sulfate aerosols. These aerosols in the lower stratosphere have an *e*-folding lifetime of about a year and were distributed globally a few months after the eruption. In our previous studies using SAGE II and CLAES/ISAMS satellite observations we developed a spectral-, space-, and time-dependent set of aerosol parameters for 2 years after the Mt. Pinatubo eruption [*Stenchikov et al.*, 1998]. This aerosol data set was used in earlier studies of the aerosol radiative forcing and climate response using the ECHAM4 [*Stenchikov et al.*, 1998; *Kirchner et al.*, 1999] and GFDL SKYHI GCMs [*Ramachandran et al.*, 2000]. *Andronova et al.* [1999] and *Yang and Schlesinger* [2002] utilized these data in the UIUC stratosphere-troposphere GCM. In this study we continue to use this aerosol data set.

[4] Pinatubo aerosols had radiative effects on both the troposphere and stratosphere. The aerosol cloud decreased the net solar surface flux in the tropics by 5–6 W/m^2 as a result of reflection of solar radiation, and increased heating in the lower tropical stratosphere by about 0.3 K/d because of absorption of thermal IR and solar near-IR radiation [*Kinne et al.*, 1992; *Dutton and Christy*, 1992; *Lacis et al.*, 1992; *Hansen et al.*, 1992; *Minnis et al.*, 1993; *Stenchikov et al.*, 1998; *Andronova et al.*, 1999; *Andersen et al.*, 2001]. Volcanic aerosols also perturb stratospheric temperatures indirectly through their heterogeneous and radiative effect on the ozone photochemistry [*Hofmann and Solomon*, 1989; *Hofmann and Oltmans*, 1993; *Turco et al.*, 1993; *Tie et al.*, 1994; *Schoeberl et al.*, 1993; *Chandra*, 1993; *Kinnison et al.*, 1994; *Randel et al.*, 1995; *Angell*, 1997a; *Rosenfield et al.*, 1997; *Solomon et al.*, 1998; *Solomon*, 1999]. As a result of the Pinatubo eruption, globally averaged surface temperature decreased by about 0.3 K for 2 years after the eruption and the temperature in the tropical lower stratosphere increased by about 2–3 K [*Labitzke*, 1994; *Randel et al.*, 1995; *Angell*, 1997b]. The tropospheric response over most land areas in the Northern Hemisphere is characterized by summer cooling and winter warming [*Robock and Mao*, 1992, 1995; *Graf et al.*, 1993].

[5] In this paper we focus on the boreal winter response, because the circulation is more vigorous than other times of the year and the AO explains a larger portion of the large winter total variance. A positive phase of the AO is expressed in winter as a zonally symmetric negative anomaly of sea level pressure (or 500 hPa geopotential height) in the Arctic and as a positive sea level pressure (or 500 hPa geopotential height) anomaly in the midlatitudes. It is also associated with a zonally asymmetric warming over North America and Eurasia in high and midlatitudes, and cooling over Greenland and the eastern Mediterranean. During recent decades the AO has had a long-term upward trend [*Labitzke and van Loon*, 1995; *Graf et al.*, 1995; *Hurrell and van Loon*, 1997; *Kodera and Koide*, 1997; *Kelly and Jones*, 1999; *Hansen et al.*, 1999; *Thompson et al.*, 2000], possibly forced by greenhouse gas (GHG) increases and ozone depletion [*Ramaswamy et al.*, 1996; *Graf et al.*, 1998; *Volodin and Galin*, 1998; *World Meteorological Organization* (*WMO*), 1999; *Randel and Wu*, 1999; *Randel et al.*, 1999; *Langematz*, 2000; *Shindell et al.*, 2001]. *Hoerling et al.* [2001] argued that the AO change could be partly caused by SST warming in the equatorial Indian and Pacific oceans observed in the course of global warming but did not provide any specific dynamic mechanism. Periods with a positive AO phase in the troposphere are associated with strengthening of the polar night jet in the lower stratosphere and growing negative anomaly of the 50 hPa geopotential height in the North Pole region [*Graf et al.*, 1995; *Kodera and Koide*, 1997]. Empirical evidence of interaction of stratospheric and tropospheric processes leads to the characterization of this type of circulation anomaly as a coupled troposphere-stratosphere mode [*Perlwitz and Graf*, 1995, 2001; *Baldwin and Dunkerton*, 1999; *Black*, 2002]. However, in GCM simulations that poorly resolve the stratospheric circulation, the stratosphere-troposphere interaction is not obvious [*Limpasuvan and Hartmann*, 1999; *Fyfe et al.*, 1999].

[6] The AO is defined statistically, and mechanisms of its excitation are not well understood [*Marshall et al.*, 2001]. However, all large equatorial volcanic eruptions during the period of instrumental observations, including the Mt. Pinatubo eruption on 15 June 1991 (Figures 1–3), have been followed by a positive phase of the AO for 1 or 2 years [*Groisman*, 1992; *Robock and Mao*, 1992, 1995; *Robock*, 2000]. Figures 1–3 show anomalies (calculated using reanalysis data [*Kalnay et al.*, 1996; *Kistler et al.*, 2001] relative to the period of 1985–1990) of atmospheric quantities in each of two boreal winters following the Pinatubo eruption. We chose 1985–1990 so as to exclude the period influenced by the 1982 El Chichón eruption and made the period short enough so that the long-term trends would not be important. The observed negative geopotential height anomaly at 50 hPa in the polar region (Figures 1a and 1b) indicates a strengthening of the polar vortex during two winters of 1991/1992 and 1992/1993 following the Pinatubo eruption. The geopotential height anomaly was strongest in the second winter, reaching −360 m, although aerosol radiative forcing decreased in the second year. The observed geopotential height anomaly at 500 hPa

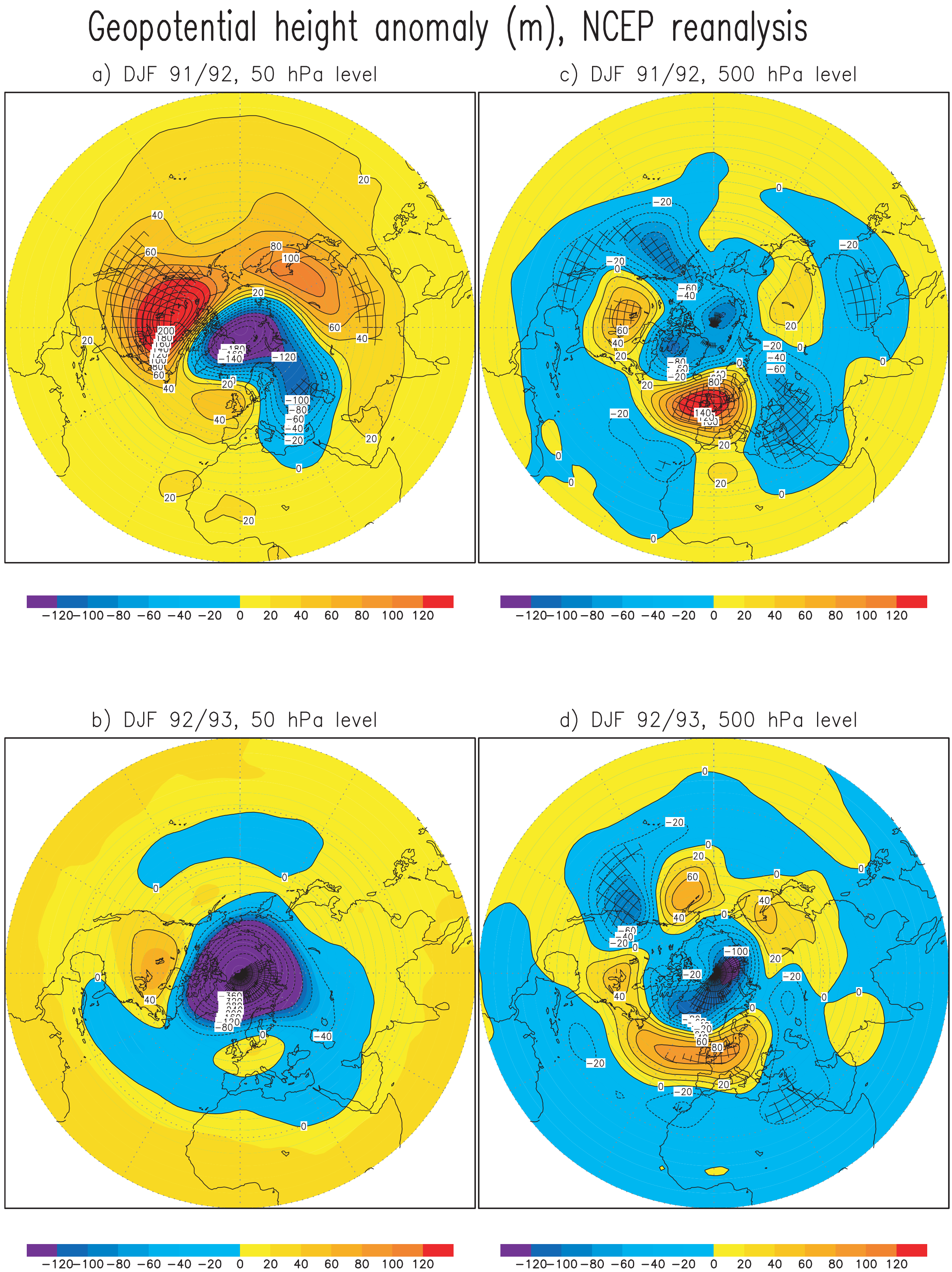

Figure 1. Seasonally averaged anomalies of geopotential height (m) at 50 hPa (a, b) and 500 hPa (c, d) obtained from NCEP reanalysis data for the winters (DJF) of 1991/1992 and 1992/1993, respectively. Anomalies are calculated with respect to the mean for the years 1985–1990. The hatching corresponds to the 90% confidence level obtained assuming a normal distribution at each grid point and no autocorrelation, with standard deviation calculated using seasonally mean quantities for the period 1951–1990.

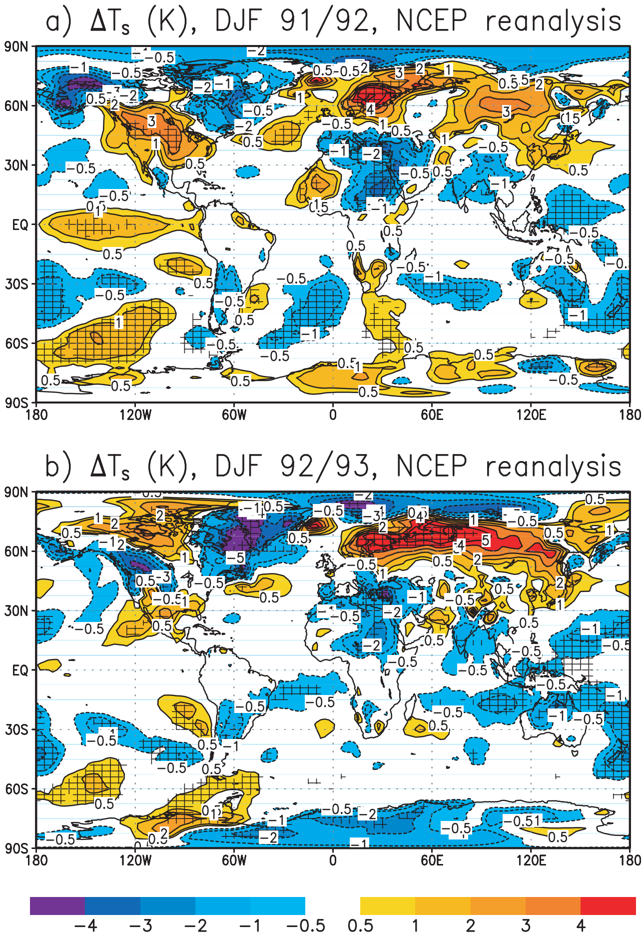

Figure 2. Seasonally averaged anomalies of the surface air temperature (K) obtained from NCEP reanalysis data for the winters (DJF) of 1991/1992 (a) and 1992/1993 (b). Anomalies are calculated with respect to the mean for the years 1985–1990. The hatching corresponds to the 90% confidence level obtained as in Figure 1.

exhibited a structure typical of the positive phase of the AO with a negative anomaly over Greenland and at the pole (>−100 m), and positive anomalies in midlatitudes over North Atlantic and Europe (<140 m), Eastern Eurasia (<40 m), and North America (<60 m) (Figures 1c and 1d). In both winters warming near the surface was observed over North America and Eurasia with cooling over southwest Asia (>−2 K) and Greenland (>−5 K) (Figures 2a and 2b). The surface temperature response over Eurasia was also stronger in the second winter, reaching 5 K (Figure 2b), than in the first one (Figure 2a). The moderate El Niño event of 1991/1992 presumably contributed to the warming over North America in the winter of 1991/1992.

[7] The zonal mean temperature anomaly in the lower stratosphere at 50 hPa showed a complex spatial-temporal structure (Figure 3a). The warming reached 2 K in the tropics a few months after the eruption. In the winter and spring of 1992, the strongest warming (3 K) was observed at 30°N. In the summer of 1992 the tropical temperature anomaly went to zero. A positive anomaly reappeared in the tropics in the winter of 1992/1993 enhanced by the westerly phase of the QBO. In high latitudes in the Northern Hemisphere in the winter of 1991/1992 polar cooling was interrupted by stratospheric warming in January associated with the breaking down of the vortex in the middle of the winter. In the second winter, 1992/1993, a stronger polar cooling of 5–6 K was observed, associated with the stable and strong polar vortex and consistent with Figure 1b. The large anomalies seen in the Southern Hemisphere springtime high latitudes are presumably related to the depth and timing of the ozone hole in these 2 years relative to the 1985–1990 comparison period. Few of the observed geopotential

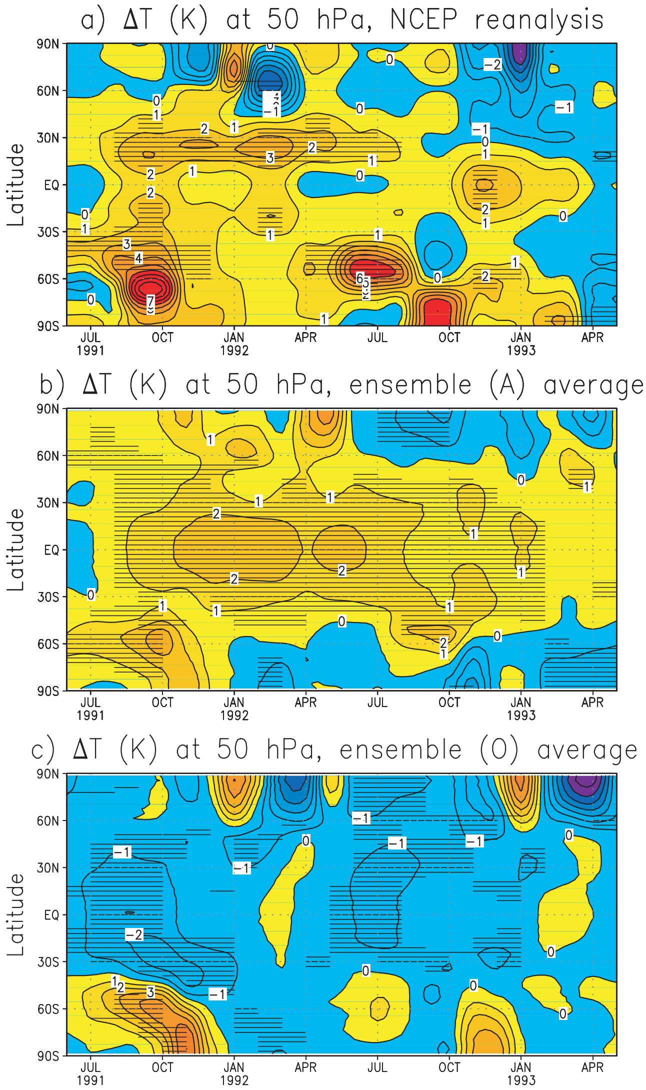

Figure 3. Zonal mean lower stratospheric temperature anomalies (K) at 50 hPa following the Mt. Pinatubo eruption on 15 June 1991. (a) Calculated from NCEP reanalysis with respect to the mean for the years 1985–1990. (b) Calculated from the ensemble (A) with respect to a 40-year mean from the control run. (c) Calculated from ensemble (O) with respect to a 10-year mean from the control run with prescribed ozone. The hatching corresponds to the 90% confidence level obtained for observations as in Figure 1, and for simulations by a local Student's *t* test.

height anomalies (Figure 1) are significant at the 90% confidence level, assuming a normal distribution at each grid point and no autocorrelation, with the standard deviation calculated using seasonally mean quantities for the period 1951–1990. However, surface warming over Eurasia and cooling in the tropics (Figure 2), as well as changes of stratospheric temperature (Figure 3a), are statistically significant over large areas at a 90% confidence level.

[8] It has been suggested that AO changes observed after volcanic eruptions were produced by aerosol heating in the tropical lower stratosphere (Figure 3a) that strengthens the equator-pole temperature gradient in the lower stratosphere [*Kodera*, 1994; *Ohhashi and Yamazaki*, 1999; *Perlwitz and Graf*, 1995; *Kirchner et al.*, 1999; *Kodera and Kuroda*, 2000a, 2000b; *Shindell et al.*, 2001]. Hereafter, we will refer to this mechanism as the "stratospheric gradient" mechanism. It states that a stronger stratospheric temperature gradient produces stronger zonal winds that tend to inhibit propagation of planetary waves into the polar stratosphere. This further strengthens the polar vortex, causes propagation of strong zonal winds southward and toward the surface, cools the polar stratosphere, and further increases the meridional temperature gradient. We will refer to this chain of processes as "wave feedback." The shift from a more transparent to a more reflecting stratosphere (for planetary waves) changes the "top boundary condition" for the tropospheric flow and affects tropospheric circulation. According to the "stratospheric gradient" mechanism, temperature variations in the lower stratosphere are of particular importance.

[9] In the study reported in the present paper we focused on atmospheric mechanisms of the AO response to volcanic forcing that presumably play a dominant role on short timescales from a season to a year. The possible sensitivity of the AO to sea surface temperature variations was not considered. We accounted for the effects of volcanic aerosols and ozone changes. Each of these forcings directly influenced the meridional temperature gradient in the lower stratosphere and could activate "stratospheric gradient" mechanism. We also found that there was an important tropospheric influence on the lower stratosphere. In this paper we report on the effects of observed aerosols and post-Pinatubo ozone changes to quantify their contribution to the AO response and test AO mechanisms. The influence of stratospheric quasi-biennial oscillation (QBO) on the aerosol effects on circulation will be considered in a forthcoming paper.

[10] This paper is organized as follows. Section 2 briefly describes the SKYHI GCM and the design of the numerical experiments. The results of simulations, including a test of the stratospheric gradient mechanism, are presented in section 3. The conclusions are summarized in section 4.

2. SKYHI Model and Experimental Setup

[11] SKYHI is a comprehensive finite difference GCM designed at the Geophysical Fluid Dynamics Laboratory to simulate the general circulation of troposphere, stratosphere, and mesosphere [*Fels et al.*, 1980]. It has been used successfully in studies of middle atmospheric phenomena, planetary wave propagation, and climate change [*Mahlman and Umscheid*, 1984; *Mahlman et al.*, 1994; *Ramaswamy et al.*, 1996; *Hamilton*, 1998]. A complete description of SKYHI climatology with prescribed climatological cloudiness is provided by *Hamilton et al.* [1995].

[12] This study uses a version of the SKYHI model with a $3^\circ \times 3.6^\circ$ latitude-longitude horizontal spatial resolution and 40-level hybrid sigma-pressure grid from the ground to 0.0096 hPa or about 80 km [*Fels et al.*, 1980]. The vertical spacing increases in the model from approximately 1 km in the middle troposphere to about 2 km in the stratosphere and over 3 km in the mesosphere. This version of the SKYHI uses predicted clouds [*Wetherald and Manabe*, 1988]. The model incorporates the improved shortwave radiative transfer algorithm of *Freidenreich and Ramaswamy* [1999], with 25 spectral bands calibrated against "benchmark" calculations. It accounts for absorption by CO_2, H_2O, O_3, and O_2, and Rayleigh scattering. The improved longwave radiative transfer scheme [*Schwarzkopf and Fels*, 1991; *Schwarzkopf and Ramaswamy*, 1999] uses eight spectral bands and accounts for absorption by H_2O, CO_2, O_3, CH_4, N_2O, and CFCs. In addition, the scheme incorporates the effects of aerosol scattering in the shortwave and aerosol absorption in the shortwave and longwave, as in [*Ramachandran et al.*, 2000]. The aerosol spatial distribution and optical characteristics were calculated following *Stenchikov et al.* [1998] for the SKYHI spatial and spectral resolution. A climatological zonal and monthly mean ozone distribution, adjusted interactively to stratospheric temperature variations using an O_3 chemistry parameterization above 10 hPa [*Fels et al.*, 1980], is used for most of the SKYHI experiments. We switched off this parameterization and used a completely specified ozone distribution when we examined the sensitivity to ozone changes. In common with most other GCMs the SKYHI model run at this moderate resolution does not simulate a spontaneous QBO in the tropical stratospheric winds.

[13] Two-year Pinatubo aerosol forcing is not sufficient to produce a strong ocean feedback [*Rind et al.*, 1992]. However, interannual sea surface temperature variability (e.g., El Niño or La Niña events) can influence the circulation and the AO [*Kirchner et al.*, 1999]. In this study, however, we focus on atmospheric mechanisms of AO response, so calculations were conducted with prescribed climatological sea surface temperatures [*Reynolds*, 1988].

[14] We calculated ensembles of experiments for 2 years following the Pinatubo eruption from June 1991 to May 1993 with prescribed aerosols or ozone changes and analyzed them with respect to multiyear control runs. The experiments, presented in this paper, are as follows:

1. *Aerosols only* (A): Ensemble of eight experiments calculated from different initial conditions as given by *Ramachandran et al.* [2000] with observed aerosols [*Stenchikov et al.*, 1998]. The anomalies are computed with respect to climatological mean from a 50-year control run without aerosols.
2. *Ozone only* (O): Ensemble of six 2-year experiments with the post Pinatubo ozone anomalies derived from the ozonesonde observations [*Grant et al.*, 1994; *Grant*, 1996;

Angell, 1997a]. These runs are analyzed with respect to a 10-year control run with prescribed climatological ozone profiles. Parameterized O_3 chemistry was not used either in the ensemble or in the control runs.

3. *Surface cooling only* (C): Ensemble of 4 experiments with the same distribution of aerosol shortwave extinction, but with longwave optical depth equal to 0 and shortwave single scattering albedo equal to 1. This almost completely eliminates stratospheric heating and limits the aerosol radiative forcing to cooling of the Earth's surface and the troposphere. The results are analyzed with respect to the control run from series (A).

3. Results

3.1. Aerosols

[15] Figures 4 and 5 show the ensemble mean geopotential height, and temperature responses to volcanic aerosols from ensemble (A). The model reproduced the strengthening of the polar vortex in the lower stratosphere (as in the observations in Figure 1); the geopotential height anomaly at 50 hPa in the first and second winters after the eruption reached -40 and -100 m, respectively (Figures 4a and 4b). In the middle troposphere at 500 hPa the geopotential height at the North Pole in the first winter decreased by 40 m and in midlatitudes increased by 20–30 m (Figures 4c and 4d). Winter warming over Eurasia and North America (Figure 5) reached 1 K in both winters. Cooling over the Labrador-Greenland region was about 0.5 K, and over southwest Asia and North Africa was about 1 K. Both dynamic and surface air temperature responses to aerosol-only forcing look similar to observations although the anomalies are weaker (compared with Figures 1 and 2). The El Niño of 1991/1992 contributed to the observed temperature change over North America although El Niño warming is centered in northwest North America and its amplitude is only about 30% of volcanic signal [*Mao and Robock*, 1998; *Robock*, 2001; *Yang and Schlesinger*, 2001; *Yang et al.*, 2000]. The El Niño effect, however, is not present in the simulations. The calculated heating of 2 K in the tropical lower stratosphere at 50 hPa was in good agreement with observations (Figure 3b), but the space-time structure of the lower stratospheric temperature response was different than observed (Figure 3a), presumably because of a lack of the QBO in the model. Figure 3b shows warming of 2 K in the 60°S–60°N latitude belt from December 1991 to June 1992. There are also regions of strong warming at 60°S in austral spring. The cooling at the North Pole in boreal winter, which was caused by a decrease of meridional wave heat transport [*Newman et al.*, 2001] because of the strengthening of the polar vortex, is weaker than observed. However, neither the observed or simulated changes are statistically significant.

[16] Large atmospheric winter variability, both in observations and in simulations, is a major problem in understanding the physical mechanisms of climate change in high latitudes [*Kodera et al.*, 1996; *Hamilton*, 2000]. The unique observed natural response to the Mt. Pinatubo eruption (Figures 1–3) is strong enough to be statistically significant at the 90% confidence level in some regions. In simulations, averaging over an ensemble of realizations reduces the largest variations but helps to retrieve a deterministic portion of system response caused by external forcing. The simulated ensemble mean surface air temperature responses are statistically significant at the 90% confidence level obtained by a local Student's t test over Northern Eurasia but not over North America (Figure 5). Temperature anomalies in the lower stratosphere are statistically significant everywhere in the tropics and midlatitudes but not in polar regions (Figure 3b), as could be expected because of high natural variability in this region [*Hamilton*, 2000; *Ramaswamy et al.*, 2002]. Simulated geopotential height anomalies at 50 hPa (Figures 4a and 4b) are statistically significant in low and middle latitudes but not in the polar region. Negative 500 hPa geopotential height anomalies are statistically significant at the North Pole, and positive anomalies are statistically significant in some regions in midlatitudes.

[17] As with any comparison of observations from one particular year with an ensemble of model simulations, we cannot expect a perfect match, as we only have one realization of the real atmosphere. In situations where the response to forcing is strong, this will be less of a problem. The simulated AO response to aerosol-only forcing may be different than in observations partly because we did not account for other important forcings caused by volcanic eruptions.

3.2. Ozone

[18] Volcanic aerosols affect stratospheric circulation and temperature, provide surfaces for chemical heterogeneous reactions, and disturb photolysis rates. All these factors are critically important for stratospheric ozone photochemistry and therefore result in substantial ozone changes following strong explosive volcanic eruptions [*Hofmann and Solomon*, 1989; *Hofmann and Oltmans*, 1993; *Schoeberl et al.*, 1993; *Herman and Larko*, 1994; *Angell*, 1997a; *Solomon et al.*, 1998; *Solomon*, 1999].

[19] Unfortunately heavy aerosol loading contaminates retrievals from most remote sensing instruments, especially from satellites, and degrades the quality of ozone data during the most dramatic period after eruptions. However, total column Total Ozone Mapping Spectrometer (TOMS) ozone data appear to be fairly reliable because of consistent aerosol corrections [*Herman and Larko*, 1994]. Figure 6b shows that the ozone loss exceeded 50 DU in middle and high latitude in the Northern Hemisphere in the boreal winter and spring of 1992 and 1993, caused primarily by Cl activation on aerosol surfaces [*Turco et al.*, 1993; *Kinnison et al.*, 1994; *Tie et al.*, 1994; *Rosenfield et al.*, 1997; *Solomon*, 1999]. At the South Pole, the spring ozone depletion of 1992 observed by *Hofmann and Oltmans* [1993] was clearly measured. In the equatorial region, the ozone amount went down by 15 DU in the first year after the eruption, because of the change of temperature, transport, radiation, and the easterly QBO phase, and went up by 5 DU in January of 1993, as a result of decreasing aerosol abundance and switching of the QBO from easterly to westerly phase in the fall of 1992 [*Angell*, 1997a; *Bruhwiler and Hamilton*, 1999].

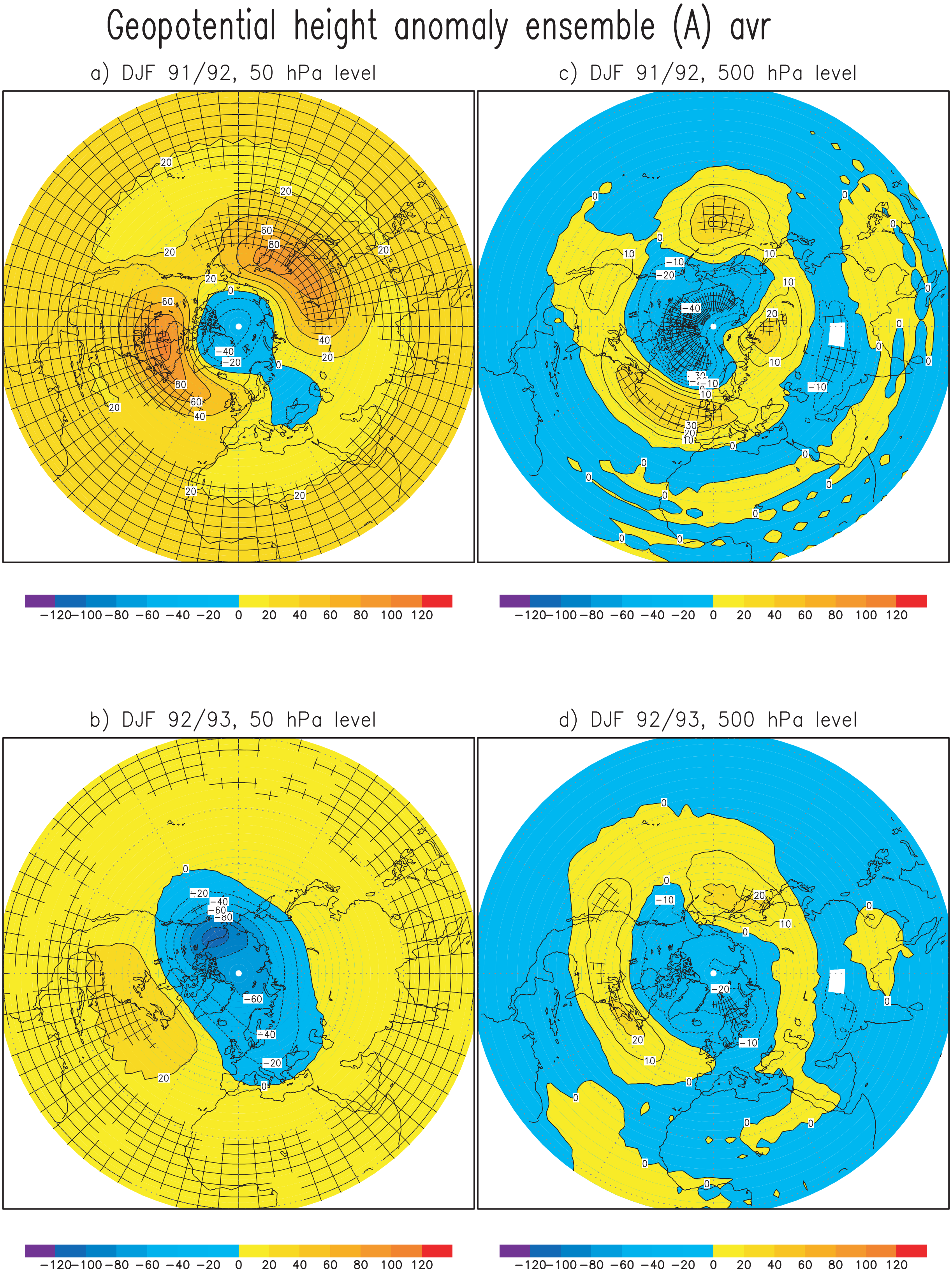

Figure 4. Seasonally averaged anomalies of geopotential height (m) at 50 hPa (a, b) and 500 hPa (c, d) obtained from ensemble (A) for the winters (DJF) of 1991/1992 and 1992/1993, respectively. Anomalies are calculated with respect to a 40-year mean from the control run. The hatching corresponds to a 90% confidence level obtained by a local Student's *t* test.

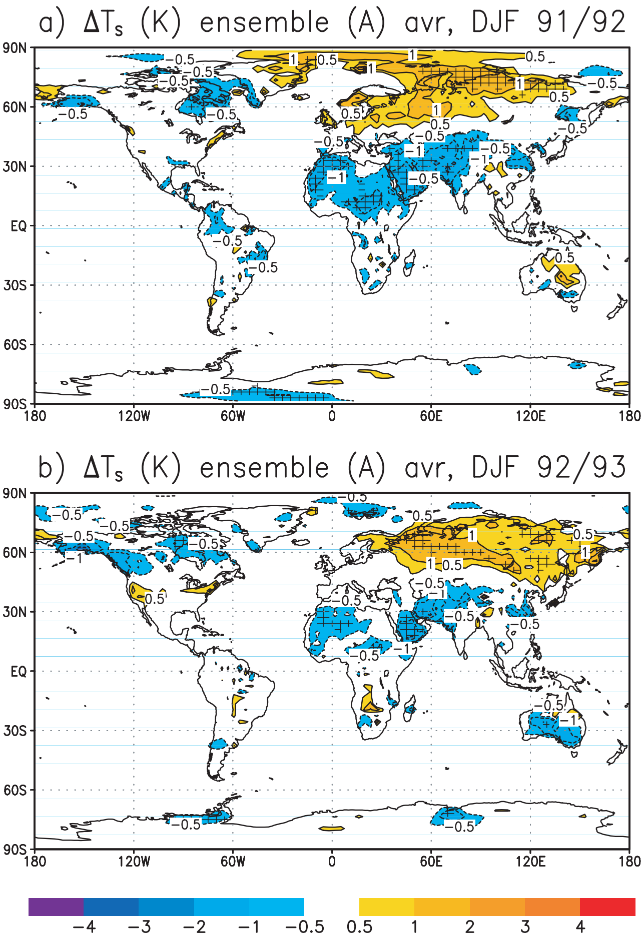

Figure 5. Seasonally averaged anomalies of surface air temperature (K) from ensemble (A) for the winters (DJF) of (a) 1991/1992 and (b) 1992/1993. Anomalies are calculated with respect to a 40-year mean from the control run. The hatching corresponds to a 90% confidence level obtained by a local Student's *t* test.

[20] To calculate vertically resolved zonal mean ozone anomalies, required for our GCM experiments, we used in-situ ozonesonde observations [*Angell*, 1997a]. The ozonesonde network consisted of three stations in the north polar zone; 12 stations in the north temperate zone; three stations in the south temperate zone; and one station in the south polar zone. Regular ozone observations have been carried out since 1968. Zonal mean relative ozone anomalies with respect to the 1969–1990 average were determined for each of the above zones [*Angell*, 1997a]. In the tropics we used ozonesonde measurements conducted from June 1990 until October 1992 at Brazzaville, Congo (4°S, 15°E), and Ascension Island (8°S, 14°W). Relative ozone anomalies in the tropics in this case were calculated with respect to SAGE II climatology [*Grant et al.*, 1994; *Grant*, 1996].

[21] To produce the ozone data set, we interpolated observed relative ozone deviations to the SKYHI spatial grid, and applied these perturbations to the zonal mean ozone distribution used in the SKYHI control run. To validate this procedure, the calculated column absolute ozone anomalies were tested with the TOMS column observations [*Randell et al.*, 1995]. The vertically integrated ozone amounts calculated using ozonesonde data (Figure 6a) lack the fine structure seen in the TOMS data (Figure 6b), but exhibit the main features discussed above and compare fairly well with the TOMS column data. Ozone loss of 40–50 DU in the north polar region in late winter and early spring were about as large as in observations. In summer, ozone loss in the polar region was present but weaker than in winter. In the tropics the ozone anomaly changed from −15 DU in 1991 to +5 DU in 1993, as in TOMS observations.

[22] The vertical structure of the ozone anomalies (Figure 7) reveals ozone loss in the equatorial lower stratosphere in the 10- to 40-hPa layer after July 1992. In the middle and high latitudes the maximum ozone losses in 1992 are located at pressure greater than 30 hPa. Positive ozone anomalies in the summer hemisphere and in the Southern Hemisphere in July 1992 at pressure less than 30 hPa were due to an increase in photolysis rates because of UV reflection from the aerosol layer and NO_x consumption [*Solomon*, 1999] in the course of chemical ozone destruction in the aerosol layer. This increase of ozone amount correlates with the intrusion of positive ozone anomalies in the period from June 1991 to July 1992 detected by TOMS (Figure 6b), but is not as pronounced

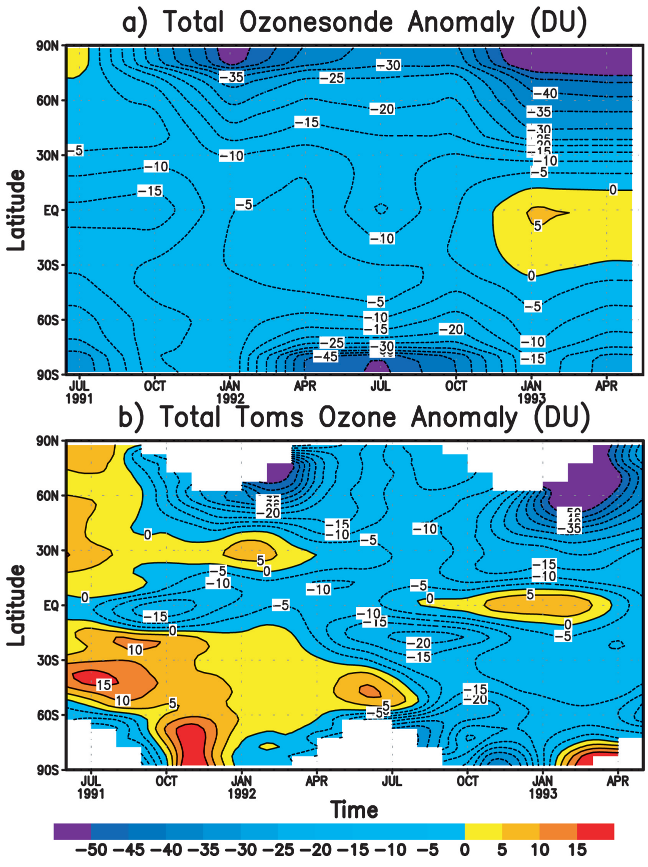

Figure 6. Zonal mean vertically integrated post-Pinatubo ozone anomalies (DU) calculated from (a) ozonesonde measurements provided by James Angell, and (b) TOMS observations, processed by William Randel and Fei Wu.

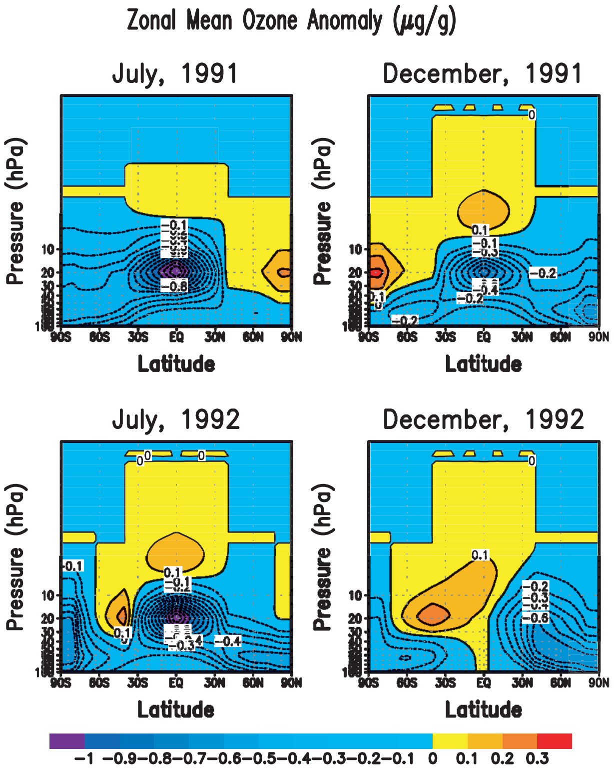

Figure 7. Zonal mean cross-section of post-Pinatubo ozone mixing ratio anomalies (μg/g) calculated using ozonesonde data.

in the ozonesonde observations (Figure 6a). The ozone amount tends to increase in the equatorial region above 10 hPa, especially during the westerly phase of the QBO from December, 1992 through May, 1993.

[23] This fairly coarse reconstruction of ozone anomalies is imperfect, but was based on available ozonesonde data. The advantage of this data set over using a calculated ozone loss [*Turco et al.*, 1993; *Tie et al.*, 1994; *Kinnison et al.*, 1994; *Rosenfield et al.*, 1997] is that it is consistent with TOMS column data and accounts for all contributing processes, including QBO. It exhibits coarse but realistic features and, we believe is good enough to calculate impact on stratospheric temperature and circulation.

[24] Ozone absorbs solar UV radiation, and radiates and absorbs thermal IR radiation. In Figure 6b the polar night regions correspond to blank regions with missing TOMS data. Beyond those regions, in midlatitudes and later in the winter and spring closer to North Pole, ozone depletion will cause cooling because of a decrease of UV absorption. The net contribution from the IR effect is smaller. So the strongest cooling from the ozone loss might be expected in the late winter and spring. Its most pronounced effect could be to delay the final warming [*Shindell et al.*, 2001].

[25] In the ensemble (O) experiments, as a response to ozone loss, the stratospheric zonal mean temperature at 50 hPa (Figure 3c) cools by about 1 K in tropics, and by more than 6 K in the North Pole region in late winter and spring, as expected. The simulated stratospheric cooling is statistically significant at the 90% confidence level in the tropics during the summers of 1991 and 1992. Polar stratospheric cooling increased the meridional temperature gradient in the lower stratosphere north of 60°N, strengthened the polar vortex, and activated the wave feedback (defined in the Introduction) in late winter and spring (Figures 8a and 8b). The negative anomaly of geopotential height in the polar region, averaged over February, March, and April, reached −180 m in 1992 and −200 m in 1993. In 1992 the anomaly stretches from the North Pole to Alaska. In 1993 it is centered on the North Pole although it was slightly zonally asymmetric. These strong anomalies are statistically significant at the 90% confidence level and affect the timing of the final warming compared to the control run. At 500 hPa, the geopotential height anomaly varied from −30 m in the polar region to +30 m in the midlatitudes. Regions with the largest changes are statistically significant. A stronger AO pattern in the lower troposphere (Figures 8c and 8d) during February, March, and April of 1992 and 1993 causes warming of 1 K over Siberia, and 0.5–1 K over North America (Figure 9). In our set of six experiments this effect is statistically significant at the 90% confidence level and is present in both years following the Pinatubo eruption, although the polar vortex is more stable in the spring of 1993 (Figure 8b).

[26] Thus, results from the (O) ensemble show that high latitude radiative cooling caused by ozone loss after an eruption produces a positive AO phase by itself. Accounting for the ozone effect in addition to the aerosol effect would strengthen and prolong the overall AO response to Pinatubo forcing, assuming that nonlinearity does not reverse the signs of ozone- and aerosol-induced contributions.

3.3. Stratospheric Gradient Mechanism

[27] Wave feedback plays a key role in stratosphere-troposphere dynamic interaction that affects the AO. It is always present in the system and is controlled by the strength of the polar vortex. The stratospheric gradient mechanism explains how changes of temperature in the lower stratosphere, caused either by high latitude ozone cooling or aerosol warming in the tropics, could affect the intensity of the wave feedback. The ozone radiative effect is confined to the lower stratosphere and, as we saw, affects the AO in late winter and spring. The results of the (A) and (O) experiments allow us to quantify the sensitivity of the polar vortex to changes of lower stratospheric temperature in polar or tropical regions. Aerosols radiatively force both the stratosphere and the troposphere. It has been assumed that the stratospheric portion of aerosol forcing is solely responsible for its impact on the AO. Here we challenge this opinion and specifically test the AO sensitivity to aerosol heating in the tropical lower stratosphere.

3.4. Tropospheric Gradient Mechanism

[28] In the ensemble (C) runs we used an aerosol forcing designed to produce no stratospheric heating, but only tropospheric and surface cooling. According to the stratospheric gradient mechanism, the AO should be less affected in the (C) experiment than in the (A) experiment, but it turns

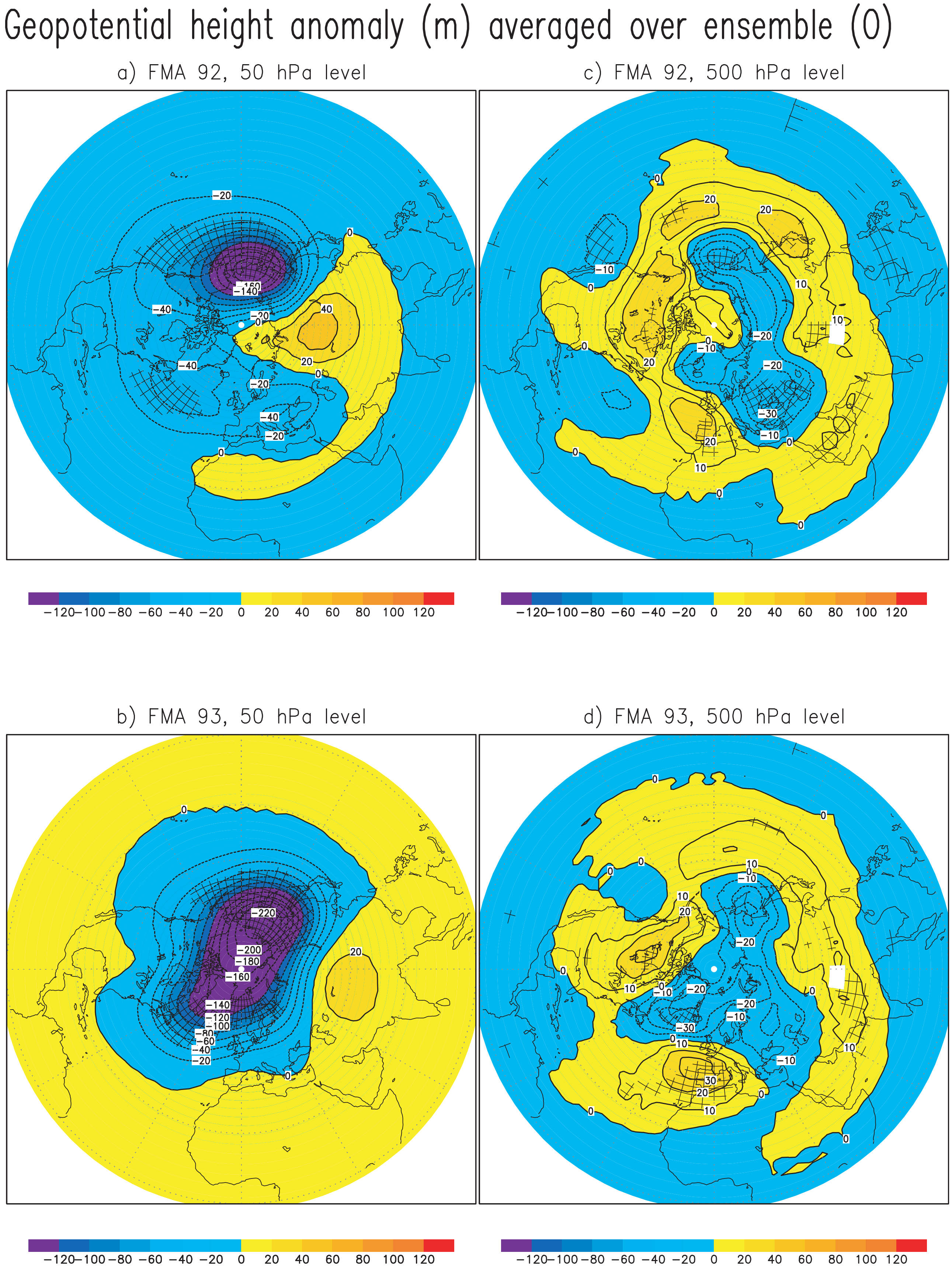

Figure 8. Anomalies of geopotential height (m) in the lower stratosphere at 50 hPa and in the middle troposphere at 500 hPa from ensemble (O) simulations for the winters of 1991/1992 and 1992/1993. Anomalies are calculated with respect to a 10-year mean from the control run with prescribed ozone. The hatching corresponds to a 90% confidence level obtained by a local Student's t test.

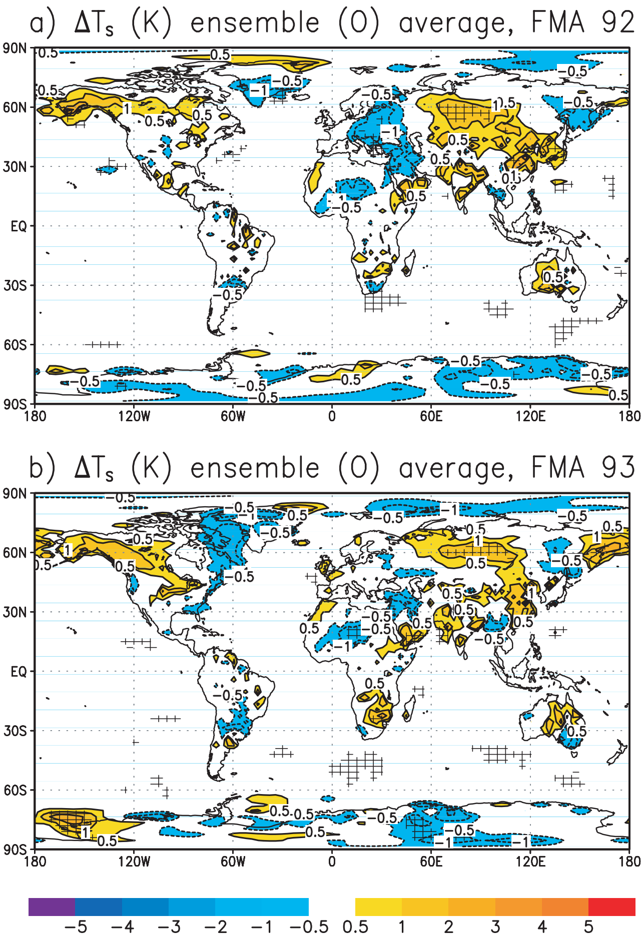

Figure 9. Anomalies of surface air temperature (K) caused by ozone changes from ensemble (O) simulations for the winters of 1991/1992 and 1992/1993. Anomalies are calculated with respect to a 10-year mean from the control run with prescribed ozone. The hatching corresponds to a 90% confidence level obtained by a local Student's *t* test.

out that the AO response in the (C) experiment was as strong as in the (A) experiment (Figure 10). Geopotential height anomalies from ensemble (C) in the lower stratosphere reached −80 m and −120 m in the winters of 1991/1992 and 1992/1993, respectively. In the middle troposphere, the geopotential height anomaly was −40 m in the polar region and +30 m in midlatitudes. Winter warming over Eurasia was 2 K and very similar to the pattern from ensemble (A). Warming over North America was stronger than in the (A) experiment, reaching 1 K, although it was not statistically significant (Figure 11). Cooling over North Africa and Middle East was about 1 K, as in series (A).

[29] Thus we see that the tropospheric portion of the aerosol forcing by itself shifts the AO to a positive phase. This suggests that the stratospheric aerosol heating is not a necessary, and may be not a major, component of aerosol radiative forcing that causes the AO response. Therefore in the (C) and (A) experiments, as well as in nature, there should be another mechanism involved that affects the AO and triggers the wave feedback.

[30] There were several previous model experiments that support this conclusion. *Graf* [1992], looking at the impact of a high-latitude volcanic eruption, forced the ECMWF GCM with only surface cooling and reported a winter warming effect. However, the model used in that study did not resolve the stratosphere very well. An idealized experiment conducted with the SKYHI GCM with a very strong equatorial heating of the stratosphere did not show significant amplification of the polar vortex.

[31] As a working hypothesis we assume that in the (C) ensemble an AO response is produced by changes primarily in the troposphere, which then affect the stratosphere, and trigger the wave feedback. Figure 12a shows the zonal mean temperature anomaly for both winters at 50 hPa for the individual realizations from ensemble (C) and the ensemble mean. The ensemble-mean temperature anomaly in the tropics was very small, as expected. The simulated cooling at the North Pole in some realizations was caused dynamically by a stronger polar vortex that depressed meridional wave heat flux [*Newman et al.*, 2001]. But the ensemble mean temperature rose at the pole by 1 K, which is well within the range of natural variability.

[32] The zonal mean surface temperature in both winters decreased in the tropics because of aerosol radiative cooling that was strong throughout all seasons in this region [*Stenchikov et al.*, 1998]. In the midlatitudes, aerosol surface forcing is weakest in the winter when insolation decreases. Therefore the winter-season meridional distribution of aerosol radiative cooling in the troposphere is most favorable to decreasing the meridional temperature gradient between the tropics and midlatitudes. The weakening of the meridional temperature gradient in the troposphere is amplified by a winter warming caused by changes in circulation. As a result of the contribution of all these processes, the temperature gradient near the surface decreases most significantly between 30°N and 60°N (Figure 12b). Decrease of the meridional temperature gradient in the troposphere leads to a decrease of mean zonal energy and amplitude of tropospheric waves. That should result in a decrease of a vertical wave activity flux, characterized by Eliassen-Palm (EP) flux, first in the troposphere, and then in the lower stratosphere.

[33] The 40-year mean vertical component of EP flux (calculated as defined by equation 2.1 of *Andrews et al.* [1983] but with reversed sign consistent with the current convention) from the control run from series (A) for the winter months at 400 hPa (Figure 12c) has a maximum value in the 30°–60°N latitude band. The EP flux for most of the eight winter individual realizations from ensemble (C) was weaker than the climatological mean in this latitude band. For two realizations, it was about 2/3 the control value and differed from it by more than one standard deviation (Figure 12c). The average vertical component of the EP flux at 400 hPa decreased in the 30°–60°N latitude band by about 5–6%. This diminished wave friction in the lower stratosphere, allowing the polar vortex to accelerate, and thus activated the wave feedback. This "tropospheric gradient" mechanism explains the results of the (C) experiment and suggests that aerosol tropospheric cooling (particularly in the low latitudes) decreases the meridional temperature gradient in the lower troposphere, damps generation of tropospheric waves, and affects the stratosphere. The tropospheric gradient mechanism feeds back and produces a further decrease of the tropospheric temperature gradient by strengthening the positive mode of the AO. At

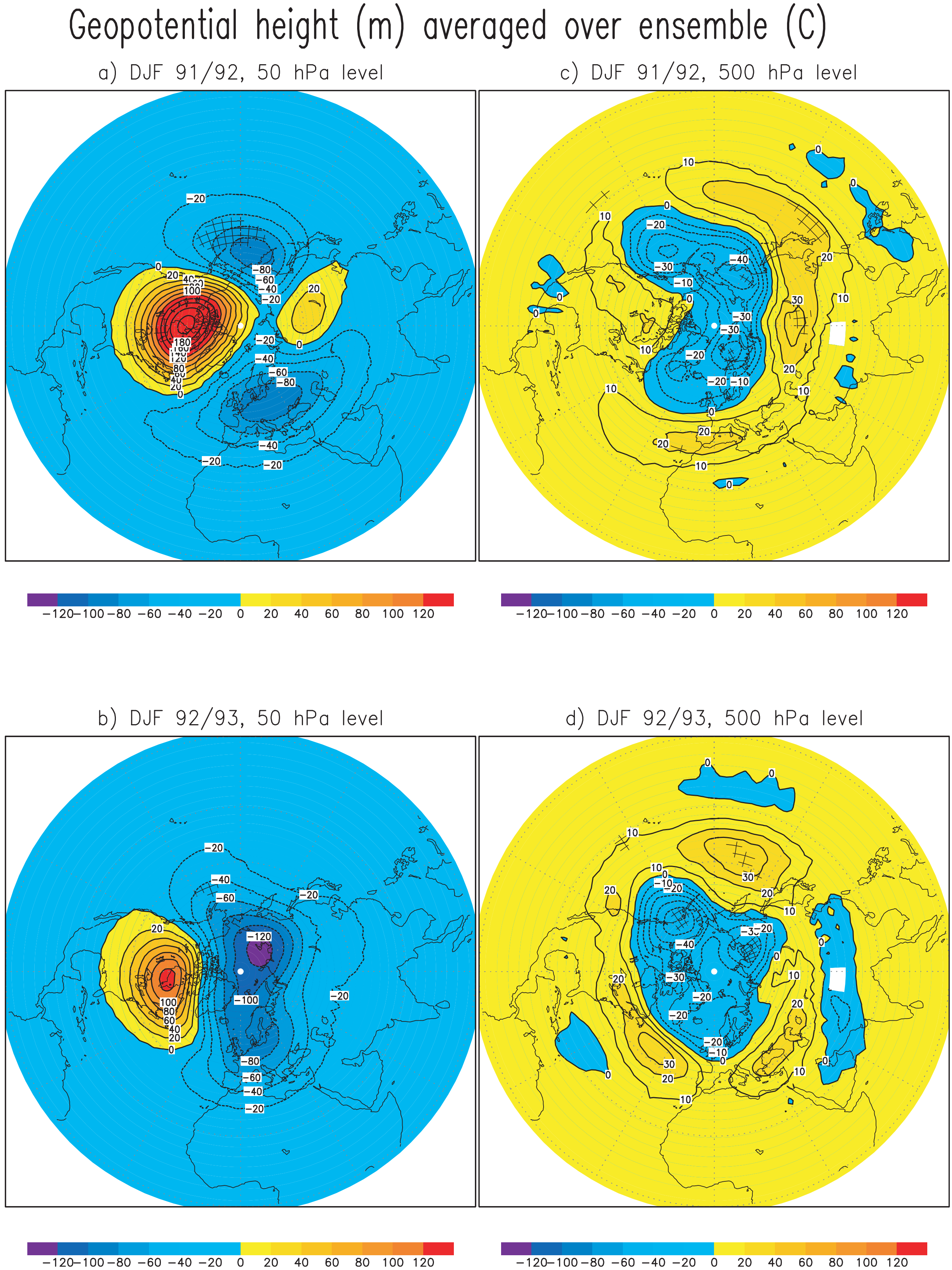

Figure 10. Anomalies of geopotential height (m) caused by Pinatubo aerosols without stratospheric aerosol heating, ensemble (C), for the winters of 1991/1992 and 1992/1993. Anomalies are calculated with respect to a 40-year mean from the control run for ensemble (A). The hatching corresponds to a 90% confidence level obtained by a local Student's *t* test.

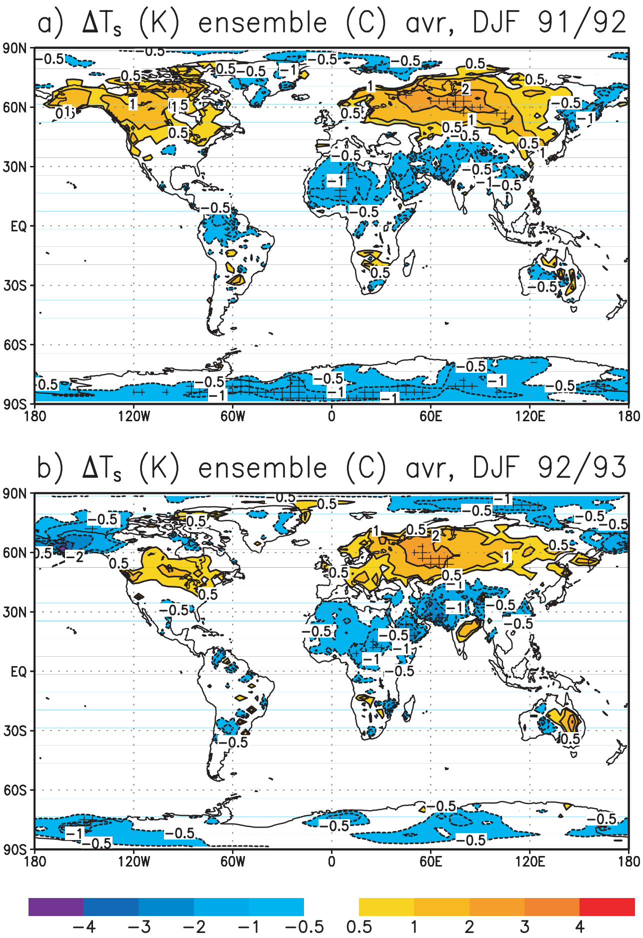

Figure 11. Anomalies of surface air temperature (K) caused by Pinatubo aerosols without aerosol stratospheric heating from ensemble (C) for the winters of 1991/1992 and 1992/1993. Anomalies are calculated with respect to a 40-year mean from the control run for ensemble (A). The hatching corresponds to a 90% confidence level obtained by a local Student's *t* test.

the same time, the meridional temperature gradient in the troposphere between midlatitudes and pole is strengthened, causing stronger zonal winds in high latitudes transporting heat from oceans to the continents.

[34] In response to radiative and dynamic forcing, surface air temperature in our experiments with prescribed SST changes primarily over land. Cooling of continents in the tropics and warming in midlatitudes in winter decreases the temperature contrast between land and ocean and contributes to weakening of planetary wave generation. Simulations with a GCM coupled to an ocean general circulation model would be required to quantify any ocean contribution. However, our recent study with a GCM coupled with the mixed layer ocean shows similar AO response to the same radiative forcing [*Soden et al.*, 2002].

4. Discussion and Conclusions

[35] The present experiments have shown that the tropospheric climate of the SKYHI GCM reacts systematically to the imposition of volcanic aerosols and ozone loss, and that these simulated changes are similar in many respects to those observed in the 2 years following the Mt. Pinatubo eruption. The tropospheric changes in winter are most plausibly explained by a significant dynamical coupling of the large-scale tropospheric circulation to the state of the stratospheric polar vortex. An anomalously strong stratospheric vortex seems to modulate the planetary wave field in such a way that an anomalously positive AO index is produced in the troposphere. The present work also suggests that volcanic effects activate this vortex-wave coupling mechanism (wave feedback) in two distinct ways (Figure 13). Volcanic ozone losses directly affect radiative forcing of the stratospheric vortex and lead to its strengthening. By contrast, in situ stratospheric radiative forcing from aerosols is largely confined to lower latitudes and so does not have a significantly direct effect on the vortex strength. However, aerosols have a significant effect on surface radiative heating and

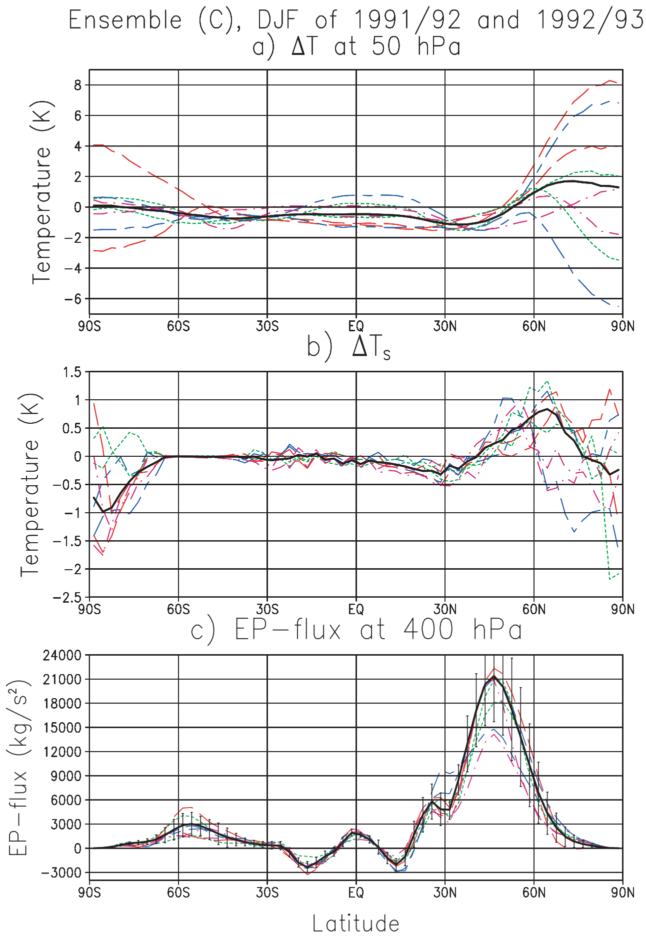

Figure 12. Zonal mean anomalies of (a) temperature at 50 hPa and (b) surface air temperature from ensemble (C) for the winters (DJF) of 1991/1992 and 1992/1993; each line is one member of the ensemble, and the solid line is the mean anomaly; anomalies are calculated with respect to a 40-year mean from the control run for ensemble (A); (c) vertical component of the EP flux (kg/s^2) at 400 hPa; each line is one member of the ensemble, and the solid line is the 40-year mean from the control run bars show one standard deviation calculated from 40-year control.

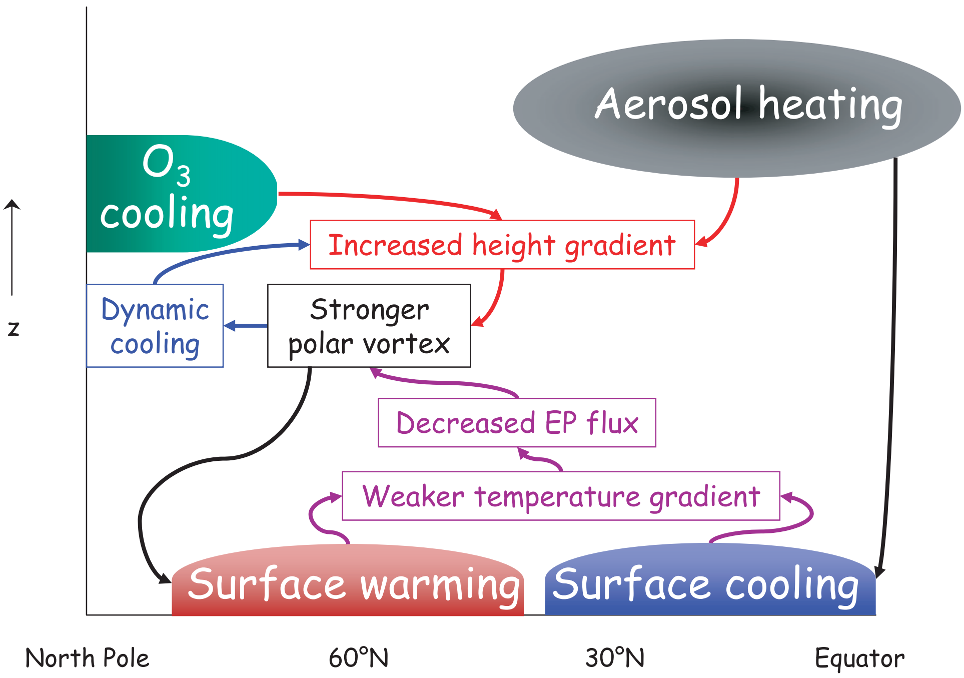

Figure 13. Schematic diagram of how the stratospheric gradient and tropospheric gradient mechanisms are triggered by volcanic aerosol clouds in the tropical stratosphere. The wave feedback mechanism amplifies the response.

produce changes in the meridional temperature gradient near the ground that in turn affect the large-scale wave field in the troposphere and associated EP flux into the stratosphere.

[36] The mechanisms of AO sensitivity discussed above are based on the effect of the lower stratospheric zonal wind on the vertical propagation of planetary waves because very strong westerlies are known to lead to enhanced vertical trapping of stationary planetary waves [*Charney and Drazin*, 1961]. Therefore these mechanisms could be activated in the period from late fall to early spring when the polar vortex is relatively strong.

[37] During the last 40 years the AO has gradually shifted to a predominantly positive phase, which could be due to increases in GHGs and ozone changes [*Graf et al.*, 1998; *Ramaswamy et al.*, 1996; *Volodin and Galin*, 1998]. *Shindell et al.* [1999a, 1999b, 2001] showed that in the GISS model, GHG increases produce a long-term positive trend of AO. They explained this effect by the stratospheric gradient mechanism, forced by a deep tropospheric heating in tropics. We suggest that changes in tropospheric planetary waves forced by the tropospheric gradient mechanism, because of polar amplification of warming caused by albedo-snow-sea ice effects, could also contribute to a long-term AO trend caused by GHGs.

[38] Radiatively induced effects of aerosols and ozone dominate over any chaotic dynamical variations in the tropics, but at higher latitudes the chaotic nature of the winter circulation plays a larger role. Therefore to be more certain about all aspects of the AO response we may need very large ensembles with 20–40 realizations to obtain statistically significant results for polar temperature anomalies in the lower stratosphere. As for comparison with observations, the real world only went through one realization, and we cannot expect the ensemble mean from the model runs to exactly reproduce one particular year. While our results suggest a plausible mechanism of AO response to large tropical eruptions, it also points out the limits of predicting the details of the pattern due to the chaotic nature of the atmospheric circulation.

[39] We can summarize our conclusions as follows:

1. In the stratosphere both the volcanic aerosol and ozone perturbations strengthen the polar vortex and force a positive phase of the AO.
2. In the troposphere both the aerosol and ozone perturbations produce winter warming consistent with the positive phase of the AO, however, the model tropospheric responses, caused separately by aerosols and ozone, are weaker than in observations, especially over North America.
3. The tropospheric cooling caused by volcanic aerosols affects the AO through the tropospheric gradient mechanism.
4. Ozone changes caused by volcanic eruptions produce a strong temperature response in the late winter and spring in the lower stratosphere in middle and high latitudes and have the potential to affect the AO through the stratospheric gradient mechanism. This mechanism may be also applicable to the effect of long-term ozone trends on the AO.
5. Based on the results reported in the present paper, it is suggested that changes in tropospheric temperature distribution and circulation caused by GHGs may affect the distribution of planetary waves in the troposphere and therefore impact the AO.
6. Statistical significance of simulated responses is an issue in the polar lower stratosphere and in the middle troposphere. Larger ensembles are necessary to get statistically stable results in model simulations.

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