Temperature Trend Patterns in Southern Hemisphere High Latitudes: Novel Indicators of Stratospheric Change

PU LIN AND QIANG FU
Department of Atmospheric Sciences, University of Washington, Seattle, Washington

SUSAN SOLOMON
Chemical Sciences Division, NOAA/Earth System Research Laboratory, Boulder, Colorado

JOHN M. WALLACE
Department of Atmospheric Sciences, University of Washington, Seattle, Washington

(Manuscript received 11 December 2008, in final form 16 June 2009)

ABSTRACT

Robust stratospheric temperature trend patterns are suggested in the winter and spring seasons in the Southern Hemisphere high latitudes from the satellite-borne Microwave Sounding Unit (MSU) measurement for 1979–2007. These patterns serve as indicators of key processes governing temperature and ozone changes in the Antarctic. The observed patterns are characterized by cooling and warming regions of comparable magnitudes, with the strongest local trends occurring in September and October. In September, ozone depletion induces radiative cooling, and strengthening of the Brewer–Dobson circulation (BDC) induces dynamical warming. Because the trends induced by these two processes are centered in different locations in September, they do not cancel each other, but rather produce a wavelike structure. In contrast, during October, the ozone-induced radiative cooling and the BDC-induced warming exhibit a more zonally symmetric structure than in September, and hence largely cancel each other. However, the October quasi-stationary planetary wavenumber 1 has shifted eastward from 1979 to 2007, producing a zonal wavenumber-1 trend structure, which dominates the observed temperature trend pattern.

Simulated temperature changes for 1979–2007 from coupled atmosphere–ocean general circulation model (AOGCM) experiments run for the Intergovernmental Panel on Climate Change Fourth Assessment Report (IPCC AR4) are compared with the observations. In general, the simulated temperature changes are dominated by zonally symmetric ozone-induced radiative cooling. The models fail to simulate the warming in the southern polar stratosphere, implying a lack of the BDC strengthening in these models. They also fail to simulate the quasi-stationary planetary wave changes observed in October and November.

1. Introduction

The high-latitude stratosphere is a critical component of the climate system. One of the most significant issues for global change is the development of the Antarctic ozone hole during the late twentieth century and the expected recovery during the twenty-first century. The local lower-stratospheric temperature affects the formation of ozone hole through both chemical and dynamical processes. It regulates the occurrence of polar stratospheric clouds (PSCs) that provide the surfaces on which heterogeneous chemical reactions occur (Solomon et al. 1986). It also reflects the strength of the polar vortex, which provides the isolated environment in which the chemical reactions that lead to the development of the ozone hole take place (e.g., McIntyre 1989; Schoeberl and Hartmann 1991).

It has also been realized recently that the stratosphere plays a much more active role in stratosphere–troposphere coupling than previously believed (Haynes 2005; Baldwin et al. 2007). Observational and modeling evidence shows that the change in the stratospheric circulation may lead to a response in both Northern and Southern Hemisphere
annular modes (NAM and SAM, respectively), and thereby influence surface weather and climate (e.g., Thompson and Wallace 2000; Thompson and Solomon 2002; Hartmann et al. 2000; Gillett and Thompson 2003). Such downward dynamical links are particularly important in high latitudes (Baldwin et al. 2007).

The spatial distribution of the long-term changes in stratospheric temperature in the Southern Hemisphere (SH) high latitudes has not been well documented. The presence of large natural variability in this region requires long and reliable observational records to identify a climate signal. Furthermore, the network of radiosonde observations is too sparse to reliably define the spatial patterns. Satellite measurements, which provide complete spatial coverage since 1979, are the primary data used in this study. Here we show the robust spatial trend patterns of stratospheric temperature in the SH high latitudes, which are consistent with previous studies of the zonal averages (e.g., Randel et al. 2009; Thompson and Solomon 2002), but reveal additional insights.

The state-of-the-art coupled atmosphere–ocean general circulation models (AOGCMs) that are the basis for the projections in the Intergovernmental Panel on Climate Change Fourth Assessment Report (IPCC AR4) tend to focus on the troposphere, and generally have poorly resolved stratospheres. Coupled chemistry–climate models simulate the stratosphere more realistically, but may still have problems accurately simulating dynamical processes such as wave propagation and gravity wave generation (Austin et al. 2003; Haynes 2005). Studies of Shine et al. (2003), Ramaswamy et al. (2006), Dameris et al. (2005), and others have attributed the observed global-mean lower-stratospheric cooling mainly to ozone depletion as well as to increasing greenhouse gas concentrations. Previous comparisons between simulations and observations in the SH high latitudes have been carried out using zonally averaged fields (Shine et al. 2003; Ramaswamy et al. 2006; Cordero and Forster 2006). Recently, based upon satellite-born Microwave Sounding Unit (MSU) channel 4 ($T_4$), observations, Johanson and Fu (2007) reported that about half of the lower stratosphere over Antarctica has warmed since 1979 during winter and spring seasons despite the expected cooling due to ozone loss. Hu and Fu (2009) further showed evidence of even stronger warming in trends for individual calendar months. They attributed such warming to the strengthening of the Brewer–Dobson circulation (BDC), and linked this dynamical heating to tropical sea surface temperature (SST) increases. We show here that the trend patterns can exhibit strong zonal asymmetries even though the climatology in this region is dominated by the zonally symmetric component of the flow. It is also of interest to compare model estimates to the observed patterns to gain insight into the processes driving stratospheric change and to assess the fidelity of their representation in models.

In this paper, we present an observational analysis of the stratospheric temperature trends in the SH, focusing on their spatial patterns and seasonal evolution. We compare the stratospheric temperature change with the ozone change on the decadal time scale, and partition it into the zonally symmetric and zonally asymmetric components. The contributions to the observed temperature trend pattern from ozone-depletion-induced radiative cooling and the BDC-strengthening-induced dynamic warming are evaluated by multiple regression. In addition to the radiative cooling and the dynamical warming, we find that changes in the quasi-stationary planetary waves also contribute to the temperature change on decadal time scales. We compare the simulated temperature trend patterns from IPCC AR4 models with observations. These simulations are found to substantially underestimate the dynamical responses to climate change.

This paper is organized as follows. Section 2 describes the data used in the study. Section 3 discusses the observed stratospheric temperature and ozone trend patterns in the winter and spring seasons. Section 4 discusses the processes that contribute to the trend patterns. Section 5 compares the simulated trend pattern with the observations. A summary and the conclusions are given in section 6.

2. Data

For the temperature analysis, we used the MSU Advanced Microwave Sounding Unit (AMSU) lower-stratospheric channel ($T_4$) monthly gridded (2.5° × 2.5°) brightness temperature data version 3.0 compiled by the Remote Sensing System (RSS) team for 1979–2007 (Mears et al. 2003). The $T_4$ weighting function extends from ~20 to ~120 hPa and peaks at around 60 hPa, thus well representing the lower stratosphere. Herein, we focus on the SH high latitudes from 45° to 82.5°S, the poleward limit of the MSU measurements.

To examine the ozone trend patterns, we used the version 8 monthly mean total column ozone data from the Total Ozone Mapping Spectrometer (TOMS). This version is formatted on a 1° × 1.25° grid, and is available since 1979, except for 1993–95. Because of the lack of solar radiation, there are no measurements over a relatively large area near the Pole during the winter season. A concern regarding the TOMS trend analysis is the uncorrectable scan-mirror problem that started in about 2000 [(World Meteorological Organization) WMO 2007, chapter 3]. The merged TOMS/Solar Backscatter Ultraviolet (SBUV) dataset constructed by the TOMS science
team (Stolarski et al. 2006) is regarded as a more reliable record that has the benefit of external calibration. The TOMS/SBUV data also partly fill the data gap between 1993 and 1995, but the resolution is much lower (5° × 10°). In our study, we found that the spatial patterns of the trends observed since 1979 do not differ significantly between the two datasets. Thus, we used the TOMS dataset to produce trend maps in order to take advantage of its high spatial resolution, whereas the TOMS/SBUV dataset was used for generating area-weighted spatially averaged ozone time series.

The National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis data (Kalnay et al. 1996) and the Japanese 25-year reanalysis (JRA-25) data (Onogi et al. 2007) are used to calculate the eddy heat flux and a zonal wind index, to represent the strength of the BDC as discussed in section 4. To compare with satellite measurements, we used monthly mean data from 1979 to 2007.

We also used the AOGCM simulations that were performed in support of the IPCC AR4 from the World Climate Research Programme (WCRP) Coupled Model Intercomparison Project phase 3 (CMIP3) multimodel dataset archive. Twenty-two AOGCM simulations contain 48 ensembles, which all considered the long-lived greenhouse gas increases. Thirteen models composing 28 ensembles also considered the stratospheric ozone depletion. A short summary of the model descriptions is listed in Table 1. Two sets of simulations are combined to get a full record from 1979 to 2007: one is the climate of the twentieth-century experiment (20C3M), the other is the set of SRES A1B simulations, which are initialized with conditions from the corresponding 20C3M ensembles at the end of the twentieth century. The ensembles in the 20C3M experiment that were not extended into SRES A1B were discarded.

The standard output for these models is at 17 vertical levels in the atmosphere (i.e., 1000, 925, 850, 700, 600, 500, 400, 300, 250, 200, 150, 100, 70, 50, 30, 20, and 10 hPa). Simulated $T_4$ is derived by vertically integrating the temperature profile with the MSU channel 4 weighting function. The horizontal resolution differs from model to model. To produce the model mean trend map, we interpolate the simulated data onto the grid format of MSU $T_4$ (2.5° × 2.5°) using a cubic-spline interpolation.

### 3. Monthly trend patterns in winter and spring seasons

#### a. Observed features in the monthly climatologies

Figure 1 shows the monthly climatological mean MSU $T_4$ and total ozone maps for 1979–2007 from July to December. During winter, the lower-stratospheric temperature field is dominated by a zonally symmetric cold pool (Figs. 1a,b), which implies a strong polar vortex above through thermal wind balance. Beginning in September, the zonal wind weakens as the polar stratosphere warms because of the absorption of the solar radiation. As the zonal wind velocity becomes smaller than the Rossby critical velocity (Holton 2004), quasi-stationary planetary waves (mainly the zonal wavenumber-1 component) are able to propagate from the troposphere into the stratosphere, and the polar vortex becomes displaced from the South Pole toward the Atlantic sector (e.g., Waugh and Randel 1999; Karpetchko et al. 2005). By December, the polar vortex is completely gone, and the zonal wind is too weak to allow the quasi-stationary waves to propagate upward from the troposphere, and thus the temperature field becomes zonally symmetric again (Fig. 1f). The ozone maps show that the low ozone area is generally confined to the shape of the polar vortex, except that its minimum occurs in September and October rather than in winter, since solar radiation is a necessary condition for ozone depletion. Nevertheless, the ozone field exhibits the same seasonal evolution as the temperature field, that is, from a more zonally symmetric state in July and August (Figs. 1g,h) to greater zonal asymmetry in September–November (Figs. 1i–k), and back to a more zonally symmetric state again in December (Fig. 1l).

The switch between the zonally symmetric and asymmetric states is more readily observed in the “eddy” fields, which are defined as

### Table 1. Model description.

<table>
<thead>
<tr>
<th>Model ID</th>
<th>Ensemble No.</th>
<th>Ozone forcing</th>
</tr>
</thead>
<tbody>
<tr>
<td>BCCR-BCM2.0</td>
<td>1</td>
<td>N</td>
</tr>
<tr>
<td>CCSM3.0</td>
<td>6</td>
<td>Y</td>
</tr>
<tr>
<td>CGCM3.1(T47)</td>
<td>5</td>
<td>N</td>
</tr>
<tr>
<td>CGCM3.1(T63)</td>
<td>1</td>
<td>N</td>
</tr>
<tr>
<td>CNRM-CM3</td>
<td>1</td>
<td>Y</td>
</tr>
<tr>
<td>CSIRO-MK3.0</td>
<td>1</td>
<td>Y</td>
</tr>
<tr>
<td>CSIRO-MK3.5</td>
<td>1</td>
<td>Y</td>
</tr>
<tr>
<td>ECHAM5(MPI-OM)</td>
<td>4</td>
<td>Y</td>
</tr>
<tr>
<td>FGOLS-g1.0</td>
<td>3</td>
<td>N</td>
</tr>
<tr>
<td>GFDL-CM2.0</td>
<td>1</td>
<td>Y</td>
</tr>
<tr>
<td>GFDL-CM2.1</td>
<td>1</td>
<td>Y</td>
</tr>
<tr>
<td>GISS-AOM</td>
<td>2</td>
<td>N</td>
</tr>
<tr>
<td>GISS-EH</td>
<td>3</td>
<td>Y</td>
</tr>
<tr>
<td>GISS-ER</td>
<td>4</td>
<td>Y</td>
</tr>
<tr>
<td>INGV-SXG</td>
<td>1</td>
<td>N</td>
</tr>
<tr>
<td>INM-CM3.0</td>
<td>1</td>
<td>N</td>
</tr>
<tr>
<td>IPSL-CM4</td>
<td>1</td>
<td>N</td>
</tr>
<tr>
<td>MIROC3.2(hires)</td>
<td>1</td>
<td>Y</td>
</tr>
<tr>
<td>MIROC3.2(medres)</td>
<td>3</td>
<td>Y</td>
</tr>
<tr>
<td>MRI-CGCM2.3.2</td>
<td>5</td>
<td>N</td>
</tr>
<tr>
<td>UKMO-HadCM3</td>
<td>1</td>
<td>Y</td>
</tr>
<tr>
<td>UKMO-HadGEM1</td>
<td>1</td>
<td>Y</td>
</tr>
</tbody>
</table>
where \( X \) is the original field, \([X]\) is the zonal-mean field, and \( X' \) is the eddy field. As shown in Fig. 2, for both stratospheric temperature and ozone, the eddy component is quite weak in winter. During September and October, when the zonal wind is favorable for the vertical propagation of the quasi-stationary wavenumber 1, the eddy component of the temperature and ozone fields exhibits a strong wavenumber-1-like structure. As the zonal wind continues to decrease, the eddy fields become weak again.

**b. Observed features in the trend patterns**

The monthly MSU \( T_4 \) trend patterns in the SH high latitudes in austral winter and spring seasons are shown in Figs. 3a–f. The monthly trend maps exhibit several interesting characteristics. First, the strong regional warming trends are largely confined to the three months from August to October, whereas the trends in July are fairly weak, and in November and December the trends are dominated by cooling. Strong cooling is also observed over part of the hemisphere in October. Second, the temperature trend patterns tend to be organized largely into zonal wavenumber-1-like patterns, with a clear phase shift of about 90° between September and October. The warming trends in the monthly data are about 2–3 times stronger than in the seasonal-mean data, with regional magnitudes of up to 3 K decade\(^{-1}\) occurring in September and October.

Figures 3g–l show the monthly trend patterns for total column ozone, which are dominated by negative trends in all winter and spring months, with strongest depletion during the “ozone hole” season, September–November. Ozone-induced radiative cooling is expected to be strongest during these months (Randel and Wu 1999). Ozone-induced radiative effects, on the other hand, cannot directly account for the strong warming, since the warming occurs in regions of observed ozone decrease.

Figure 4 shows the eddy component of the temperature and ozone trend patterns. The number at the upper-right corner indicates the area-weighted relative contribution of the eddy component to the total trends. This fraction is defined as

\[
f = \frac{\sum X'^2}{\sum X^2},
\]

where \( X' \) and \( X \) are the eddy component and the total trends, respectively, and \( \sum \) represents the area-weighted sum over the southern high latitudes (\( \geq 45^\circ S \)). The eddy components of both temperature and ozone trends are dominated by zonal wavenumber-1 structures in nearly all winter and spring months. Close agreement is seen in
the eddy component of the trend patterns for temperature and ozone in terms of both the relative magnitudes and the phase. The coherent ozone and temperature changes during the SH spring season have also been documented in previous studies (e.g., Newman and Randel 1988; Randel and Cobb 1994; Randel and Wu 1999). Such good agreement between these two independent observations confirms the reliability of the trend analysis.

The strongest warming trend occurs in September and October, which is also the season during which the climatological-mean zonal asymmetry (Figs. 2c,d) and quasi-stationary wave activity are the strongest (Randel 1988). The eddy components of the climatological mean fields and the trends are both dominated by a wave-number-1 structure. In August and September, the phase of wavenumber 1 in the climatology and in the trend is almost the same. However, in October and November, the phase in the trend is more than 90° to the east of that in the climatology. This difference suggests that the cause of the trend pattern observed in October and November may be different from that in September and August.

Figure 5 shows the monthly decadal mean $T_4$ climatologies averaged over 1979–88, 1989–98, and 1999–2007. Comparing the climatologies for these three decades, there is no obvious change in July and August. Gradual warming is seen in September over the subpolar region on the Australian side. In October and November, departures from the zonally symmetric state are clearly discernible. However, unlike the situation in September, the wave pattern in October and November does not remain in place, but gradually shifts eastward. The phase shift is most apparent in November and in the most recent decade.

Comparing the eddy component of the trends with the total trends, it is clear that the eddy contribution accounts for most of the local $T_4$ trend patterns except in November and December (see Figs. 4a–f versus Figs. 3a–f). The eddy component is especially strong in August, September, and October, the months when the warming is strong. It is then not surprising that little or no warming is reported in the zonal mean trends for these months (e.g., Thompson and Solomon 2005; Randel et al. 2009), because the warming is largely canceled by cooling of comparable magnitude in the zonal average.

The eddy contribution to the ozone trend is however much less than in $T_4$ in all winter and spring months. This difference indicates different behaviors of the zonally symmetric components of temperature and ozone on decadal time scales (i.e., little trend for temperature and a strong decreasing trend for ozone). Since a large zonally symmetric decrease in ozone would lead to a strong
radiative cooling of the stratosphere, a compensating zonally symmetric, dynamically induced warming is required to account for the observed absence of a zonal-mean stratospheric cooling trend (Thompson and Solomon 2009).

c. The robustness of the trend pattern

Considering the large variability in polar region during the late winter–spring season, and the unusual sudden stratospheric warming that occurred in Antarctica in 2002, one might question whether the observed trend patterns are statistical significant. Figure 6 shows the $T_4$ trend for August, September, October, and November from 1979 onward with different ending year. Consistent trend patterns are seen regardless of the ending years, indicating the observed trend patterns are not unduly influenced by the effect of unusual years such as 2002. The cooling trends in October and November are stronger and cover a larger area during early years, which is consistent with strong ozone depletion during the early part of the record (WMO 2007). Despite the stronger cooling over the polar cap region as a whole in these two months in the early years, the spatial structures of the trend patterns are nevertheless quite similar.

To further establish the statistical significance of the trend pattern, we carried out Monte Carlo tests by randomly switching the order of years, and calculating trends in these scrambled $T_4$ time series. By repeating such Monte Carlo simulations enough times, we can obtain the distribution of trends that can possibly be achieved by natural variability. In contrast to the Student’s $t$ test of trends at each grid point as done by Hu and Fu (2009), we focused on the robustness of the whole spatial trend pattern. As shown in Figs. 3a–f, the zonal wavenumber-1 component dominates the observed trend structure. We thus computed the wavenumber-1 component in those simulated Monte Carlo trends for each month by performing Fourier decomposition on each latitude circle and then averaging over 50°–70°S weighted by the cosine of the latitude. Figure 7 shows the distribution of the amplitudes of these mean wavenumber-1 trends among 10 000 Monte Carlo simulations. The possibility of achieving a wavenumber-1 trend structure as large as observed by chance is less than 1% in October. Thus, the observed temperature trend patterns are indeed strongly statistically significant. In other months, the observations are likely to be significant.

4. Interpretation of the trend patterns

In this section, we examine the $T_4$ trend patterns using multiple regression. We did not detrend the data before the regression analysis. Two physical processes are considered to be responsible for the high-latitude stratospheric temperature change. One is related to the absorption of solar radiation by the stratospheric ozone that warms the lower stratosphere. We defined an “ozone index” as the area-weighted spatial-mean total ozone over 45°S poleward to represent the ozone effect. The other process is associated with the BDC, which has a descending branch in high latitudes that warms that region adiabatically. The BDC is driven remotely by planetary and gravity wave breaking in the extratropical stratosphere, which acts like a “suction pump,” drawing air upward from the tropics (Haynes et al. 1991; Holton et al. 1995). The waves originate in the mid- to high-latitude troposphere, propagate upward, and break in the stratosphere, depositing negative momentum and decelerating the stratospheric jet. Thus, when the BDC becomes stronger, indicative of more wave activity propagating upward into the stratosphere, the stratospheric jet should become weaker, and vice versa.
A conventional index of the strength of the BDC is the eddy heat flux in the lower stratosphere (e.g., Hu and Tung 2002; Haklander et al. 2008), which is equivalent to the vertical component of the Eliassen–Palm (EP) flux, representing the amount of the wave activity propagating from the troposphere to the stratosphere (Andrews et al. 1987). We here defined an “eddy heat flux index” as the 3-month mean area-weighted average eddy heat flux at the 150-hPa level over 45°–90°S. Wave activity in the previous two months is also included because it can contribute to the dynamical heating in the current month (Hu and Tung 2002). Because of the limited resolution of the reanalysis data, this calculated eddy heat flux only takes into account the resolved waves, but misses the waves on the subgrid scale (e.g., gravity waves).

The stratospheric jet, feeling the total effects from wave breaking on all scales, also provides an indication of the strength of the BDC. Following Ueyama and Wallace (2010), we defined a “zonal wind index” as another index for the strength of the BDC, which is the

![Fig. 6. Monthly $T_4$ trend maps for August–November from 1979 to various ending years as indicated on the left.](image-url)
area-weighted spatial-mean zonal wind over the lower part of the stratospheric jet region (40°–70°S, 20–10 hPa). (No NCEP–NCAR reanalysis data are available above 10 hPa.) While having the advantage of being able to represent the impact of waves on all scales, this zonal wind index is subject to limitations of its own. It can be affected, not only by wave driving, but also by other processes, such as radiative cooling due to ozone depletion in the polar stratosphere, which can induce a strengthening of the zonal wind through the mean meridional circulation. We here show the results from both the eddy heat flux index and the zonal wind index, which agree quite well with one another during austral winter and spring seasons. This agreement lends confidence in these two indices for representing the strength of the BDC.

We perform multiple regression of gridded monthly-mean $T_4$ data upon the two indices in the following form:

$$T(x, y, t) = K_{O3}(x, y) \times X_{O3}(t) + K_{BDC}(x, y) \times X_{BDC}(t) + \text{residual},$$

where $X_{O3}$ is the standardized ozone index, and $X_{BDC}$ is the standardized eddy heat flux index or zonal wind index. The regression maps $K_{O3}$ and $K_{BDC}$ represent patterns of temperature anomalies that correspond to a one standard (positive) anomaly deviation in the indices assuming a linear relationship. The contributions to the $T_4$ trend attributable to changes in ozone concentration and the strength of the BDC are thus obtained by multiplying the regression maps by the linear trend in the corresponding standardized indices.
We calculate the eddy heat flux and the zonal wind indices using the NCEP–NCAR reanalysis data. Figure 8 shows the monthly $T_4$ trend derived from the NCEP–NCAR reanalysis, which agrees very well with the satellite observations (Figs. 3a–f). This close agreement lends confidence to the analysis of trends in wave activity based on the NCEP–NCAR reanalysis. We also derived the indices from the JRA-25 reanalysis data. The resulting regression maps and trend patterns (not shown) are very similar to those derived from the NCEP–NCAR reanalysis.

We focus on September and October because these are the months in which the trend patterns are most significant. Furthermore, August $T_4$ exhibits a behavior similar to September, and November is more like October. Similar results can be obtained from empirical orthogonal functions/principal component (EOF/PC) analysis, as shown in the appendix.

### a. September

Figures 9a,c show the time series of the eddy heat flux and ozone indices for September. The corresponding regression maps are shown in Figs. 9b,d expressed in units of kelvin per unit of standard deviation of the eddy heat flux or ozone.

The polar stratospheric temperature response to the eddy heat flux is dominated by negative values (Fig. 9b), where stronger poleward (more negative) eddy heat flux implies more upward propagation of wave activity, a stronger BDC, and stronger subsidence-induced warming. In contrast, the temperature response to ozone is dominated by positive values (Fig. 9d), because more ozone will absorb more solar radiation and will thus warm the stratosphere.

It is notable that the temperature responses to the variations in spatial-mean ozone and eddy heat flux are not zonally symmetric, and their primary centers are at different locations. The eddy heat flux response (Fig. 9b) is centered along the Antarctic coast poleward of Australia. The zonal asymmetry reflects the fact that the center of the polar vortex in September is generally displaced from the Pole toward South America (see Fig. 1c). Since sinking is inhibited near the center of the strong polar vortex (e.g., Manney et al. 1994, 1999), the descending

![Fig. 8. As in Figs. 3a–f, but for $T_4$ simulated from the NCEP–NCAR reanalysis.](image)

**Fig. 8.** As in Figs. 3a–f, but for $T_4$ simulated from the NCEP–NCAR reanalysis.

![Fig. 9. (a) Time series of the eddy heat flux index (solid line) and its linear trend (dashed line) in September. (b) Map of September $T_4$ regressed onto the eddy heat flux index in units of K per standard deviation. (c) Time series of the ozone index (solid line) and its linear trend (dashed line) in September. (d) Map of September $T_4$ regressed onto the ozone index in units of K per standard deviation. The trend uncertainty is given at the 95% confidence level ($2\sigma$). See text for the definitions for the eddy heat flux index and the ozone index.](image)
branch of the BDC tends to be concentrated on the other side of the Antarctica (i.e., toward Australia). Such zonal asymmetry in the distribution of the descending air is also evident in the ozone distribution, because the BDC advects ozone-rich air from the tropical region. The sinking region has been referred to in previous studies as the “ozone croissant” or the “ozone collar” (e.g., Grytsai et al. 2007; Mariotti et al. 2000).

The maximum amplitude of the ozone response (Fig. 9d) is located near (65°S, 15°E). This pattern is similar to the ozone trend pattern as shown in Fig. 3i, but is of opposing sign.

Figure 10a shows the observed September \( T_4 \) trend (same as Fig. 3c), and Figs. 10b,c show the contributions to the \( T_4 \) trend in the same month due to the changes in the eddy heat flux and ozone, respectively. The least squares best-fit linear combination of the two is shown in Fig. 10d. Since 1979, ozone concentration has decreased (Fig. 9c), resulting in a large area of radiative cooling (Fig. 10c). At the same time, the poleward eddy heat flux has strengthened in association with a strengthening of the BDC (Fig. 9a), which causes warming centered at the edge of the Antarctica toward the Australian side (Fig. 10b). A strengthening of the BDC has been predicted in response to global warming and ozone depletion in many climate models (e.g., Rind et al. 1998; Eichelberger and Hartmann 2005; Butchart et al. 2006; Li et al. 2008). Recent observational records also indicate an increased cooling in the tropical lower stratosphere associated with the strengthening of the BDC (Randel et al. 2006; Rosenlof and Reid 2008). The SH high-latitude stratospheric warming identified in this study, provides observational evidence that such a strengthening is, in fact, occurring.

The linear combination of the eddy heat flux- and ozone-driven trends captures the observations extremely well (Fig. 10d versus Fig. 10a), and the residual, defined as the observed trend minus the combined trend, is small over most of SH high latitudes (Fig. 10e). The close agreement suggests that the temperature change on the decadal time scale is indeed largely driven by the direct radiative impact of ozone depletion and the strengthening of the BDC. Also worth mentioning is the fact that the observed \( T_4 \) trend structure dominated by the wavenumber-1 component may not necessarily be caused by amplification of the planetary waves. Rather, it could be because the radiative response in temperature to ozone depletion and the dynamical response to the BDC have exhibited opposing tendencies during the past few decades, with different primary action centers. Thus, the cooling associated with decreasing ozone and the warming associated with the strengthening of the BDC would not cancel each other, but form a wavenumber-1-like structure instead.

There is a weak cooling in the September residual trend centered to the west of the Antarctic Peninsula, which is collocated with the secondary center of ozone depletion (see Fig. 3i), suggesting that the radiative cooling due to ozone depletion may be slightly underestimated by the multiple regression method. Note that the observed cooling could also be partially linked to the residual radiative cooling due to the ozone depletion in the previous year and the radiative effect of the increase of greenhouse gases, although these effects may be small at the level we considered.

Figure 11 shows the contributions to the \( T_4 \) trend due to changes in the BDC and ozone as represented by the zonal wind and ozone indices. They are similar to those based on eddy heat flux and ozone indices.

For August, multiple regression analysis yields patterns similar to those for September but with weaker trends. The seemingly wavenumber-1 trend structure in August can also be separated into two parts: radiative cooling due to ozone depletion and dynamic warming due to the strengthening of the BDC.

b. October

Unlike in September, the phase of wavenumber 1 in October has shifted eastward substantially during the last three decades as shown in Fig. 5. This shifting has certainly affected \( T_4 \) trend pattern. To separate the \( T_4 \) trend due to the phase shift from those due to the change in BDC and ozone-induced radiative effect, we added a third index in October to describe the phase shift of the
planetary wave. Here a “phase index” is defined as the longitude of the crest of the mean zonal wavenumber-1 component, which is averaged over 50°–70°S, weighted by the cosine of latitude.

Figures 12a,c,e show the time series of the eddy heat flux, ozone, and phase indices in October, and the corresponding regression maps are shown in Figs. 12b,d,f in units of kelvin per unit of standard deviation. The corresponding contributions to the trend in October $T_4$ are shown in Fig. 13.

The temperature response to changes in eddy heat flux in October (Fig. 12b) is negative as in September, but exhibits a primary action center located over the Ross Sea. Compared with September, this temperature response exhibits a more prominent zonally symmetric component with the primary center closer to the Pole.

The temperature response to ozone concentration variation in October (Fig. 12d) is positive as in September. It exhibits a quasi-zonal symmetric structure that is slightly displaced off the Pole toward western Atlantic sector, which is the same position as the climatological polar vortex (Figs. 1d,j).

The temperature response to the phase change of wavenumber 1 in October (Fig. 12f) exhibits zonal wavenumber-1 structure as expected. This wavenumber-1 structure is orthogonal to the climatological mean wavenumber 1. A positive phase index is associated with an eastward shift relative to the climatological mean state.

![Fig. 11. As in Fig. 10, but that the BDC is represented by the zonal wind index.](image)

![Fig. 12. (a)–(d) As in Fig. 9, but for October. (e) Time series of the phase index (solid line) and its linear trend (dashed line) in October. (f) Map of October $T_4$ regressed onto the phase index in units of K per standard deviation. The trend uncertainty is given at the 95% confidence level ($2\sigma$). See text for the definition for the phase index.](image)
During the past few decades, the strengthened BDC has induced strong warming (Fig. 13b), and ozone depletion has induced strong cooling (Fig. 13c) in October. These dynamical warming and radiative cooling patterns are more zonally symmetric than their counterparts in September, so they cancel each other to a large extent. In addition, phase of wavenumber 1 has shifted eastward by about $15^\circ$ decade$^{-1}$ (Fig. 12e). This shifting induces a temperature trend with a wavenumber-1 structure (Fig. 13d) that is similar to the observed total temperature trend (Fig. 13a). This trend pattern is consistent to that derived by rotating the climatological temperature field following the phase of the wavenumber-1 component in each year of last three decades (not shown). A similar phase change of wavenumber 1 is also observed in ozone field (Grytsai et al. 2007).

Previous studies (Quintanar and Mechoso 1995; Inatsu and Hoskins 2004) showed that the quasi-stationary zonal wavenumber 1 in the SH upper troposphere and lower stratosphere is primarily thermally forced by zonal asymmetries in the tropical SST distribution. While the tropospheric forcing has remained relatively constant, the phase of the planetary waves could have been affected by changes in the mean state of the atmosphere, especially in the zonal wind field, as the wave propagates upward (e.g., O’Neill and Youngblut 1982; Karoly and Hoskins 1982; Nishii and Nakamura 2004). In this case, the phase index in October is found to be significantly correlated with the ozone index ($r = -0.53$), but not with the eddy heat flux index or zonal wind index. The time series of phase index also exhibits strong peaks in the years 1982 and 1992, immediately after the major volcanic eruptions of El Chichón and Pinatubo. These results suggest that concentrations of ozone and volcanic aerosols may affect the phase of the planetary wavenumber 1 in the lower stratosphere, though further research is required to understand the mechanism responsible to this eastward phase shift.

The linear combination of the three responses (Fig. 13e) captures the observed $T_4$ trend very well, with a very small residual (Fig. 13f).

We also performed the multiple regression on $T_4$ with time series of the zonal wind, ozone, and phase indices as predictors. As shown in Fig. 14, the spatial structures of the corresponding regression maps were found to be similar to those derived from the eddy heat flux index. Multiple regression of $T_4$ upon the time series of zonal wind and ozone concentration only (i.e., without the phase index) can also capture the observation well. But in this case, the temperature trend associated with ozone change is a combination of radiative cooling and that due to wavenumber-1 rotation, which is consistent with the summation of Figs. 14c,d. This is not a surprising result, because the phase index does have a significant correlation with the ozone index.

The observed $T_4$ trend pattern in November can also be separated into three parts as in October: quasi-zonally symmetric patterns associated with changes in ozone concentration and the strength of the BDC, and a pattern associated with eastward phase shift of zonal wavenumber 1. Compared with October, the trend of the BDC is much weaker. The radiative effect induced by ozone depletion dominates in November.

5. Comparison between models and observations

In this section, we will compare the $T_4$ trend patterns deduced from IPCC AR4 AOGCMs with observations.
Figure 15a shows monthly trend maps from the ensemble model means, which are dominated by zonally symmetric cooling throughout winter and spring, without any regional patches of warming. The models seem to capture the temperature trend only from the direct radiative forcing. The maximum cooling is found in November, about 1 month later than the observed maximum ozone loss in the lower stratosphere.

We further separate the models into two groups based on how they treat the stratospheric ozone forcing. Figure 15b shows the results from the models with time-varying ozone including ozone depletion based on observations. The results shown in Fig. 15c are from models that only consider the climatological-mean seasonal cycle of stratospheric ozone but ignore any long-term change. In both cases, the monthly trend maps are dominated by cooling, though the cooling in models without ozone depletion is much weaker. Previous attribution experiments (e.g., Shine et al. 2003; Ramaswamy et al. 2006; Dameris et al. 2005) suggested that the cooling observed in the polar lower stratosphere can be mainly attributed to ozone depletion, although the increase of well-mixed greenhouse gas and the increase of stratospheric water vapor may also play a role. Thus, the weak cooling in the no-ozone-depletion group can be explained by greenhouse gas increases, whereas the strong cooling in the models considering ozone-depletion includes contributions from both ozone depletion and greenhouse gas increases.

Next we remove the zonally symmetric component of the simulated trend to check whether there is any wavenumber-1-like structure analogous to that in the observations. As shown in Fig. 16, the eddy component in models indeed tends to be organized into a wavenumber-1-like shape, though not as well defined as in the observations, but the amplitudes are generally about an order of magnitude smaller. The strongest and most well-defined wavenumber-1 structures are seen in November in the all-model ensemble mean as well as in the two subgroups, a month later than in the observations. The phase of the wavenumber-1 trend in the ozone-depletion group does not change much from month to month throughout August to November, with the crest remaining over the southern Pacific. For the no-ozone-depletion group, the phase does change from month to month. Both model groups capture the phase of the wavenumber-1 trend observed in November, but not in other months. In general, the ozone-depletion group does not exhibit any apparent improvement over the no-ozone-depletion group with respect to this behavior.

In contrast to the trend pattern, the climatological mean monthly temperature fields from the all model mean and the two groups are very similar to the observations (not shown). As for the eddy components, the simulated amplitudes are about half as large as in the observations, but the phases are the same. There is no obvious difference between the two model groups.

6. Summary and concluding remarks

Stratospheric temperature trends in the SH high latitudes over the past three decades exhibit well-defined, seasonally dependent spatial structures. Robust wavenumber-1-like structures with both warming and cooling are observed during the late winter and spring seasons, especially in September and October. A phase shift in the trend pattern is observed between September and October. Temperature and ozone exhibit similar patterns
in the eddy component of trend. However, for the zonally symmetric component, temperature and ozone behave quite differently on the decadal time scale.

We obtained the stratospheric temperature fingerprints in response to the changes in SH planetary wavenumber 1, total ozone (the direct radiative effect), and the strength of the BDC through multiple regression. In addition to the well-known direct radiative cooling induced by ozone depletion, our study indicates that the strengthening of the BDC produced an adiabatic warming of the polar stratosphere. The stationary planetary wavenumber 1 is also found to exhibit a significant change in phase during late spring. The temperature responses to the latter two processes (viz., the global and local dynamical processes) appear to be as strong as or even stronger than the ozone-induced radiative cooling in winter and spring months.

While the trends in $T_4$ in each calendar month are all influenced by radiation, BDC, and planetary waves, the lower-stratospheric temperature trends exhibit different properties in late winter and spring. In August and September, ozone-induced radiative cooling and BDC-induced dynamic warming both play important roles, whereas stationary wavenumber 1 does not exhibit a significant long-term change. The polar vortex is generally displaced from the Pole due to the presence of the planetary wavenumber 1. Thus, the strongest cooling inside the vortex associated with ozone depletion, and the strongest warming outside the vortex associated with the trend toward stronger descending motion, are centered in different locations. Our study has shown that the cooling and warming do not cancel one another, and the sum of the two contributions forms a wavenumber-1-like structure that dominates the trends. This finding has important implications for interpreting trend data and understanding climate change in Antarctic stratosphere.

In October and November, there is a systematic eastward shift in the quasi-stationary wavenumber 1 besides the ozone-depletion-induced radiative cooling and the BDC-strengthening-induced warming. This eastward shift alone leads to a wavenumber-1 temperature trend pattern that is about 90° to the east of the climatological wavenumber 1. In October, the phase shift accounts for most of the observed trend, while the asymmetric parts of the radiation and BDC further enhance the wave structure. In November, the radiative effect dominates, and the circulation effect and the phase-shift effect are relatively weak.

While seldom noticed by earlier studies, such an eastward shift in stationary wavenumber 1 certainly affects the spatial structure of regional climate change. Grytsai et al. (2007) related the eastward shift of wavenumber 1 over Antarctica to a horizontal displacement of the atmospheric centers of action under global climate change, and suggested that such change in the stratosphere may be related to changes in the tropospheric planetary waves. Our study indicates that the ozone and volcanic aerosol concentration may play roles in the phase shift of stationary wavenumber 1. While our results provide a first look, the mechanism for the eastward shift of stationary wavenumber 1 requires further investigation. It should be noted that in addition to the SH planetary wavenumber 1, the decadal change of the BDC can also be indirectly driven by the ozone depletion.

The IPCC AR4 ACGCMs do not capture the observed spatial trend pattern in the SH high-latitude stratosphere in the winter and spring seasons. They fail to simulate the response of the BDC to global warming in this region. The long-term changes in these waves are completely missed. Considering the prominence of the high-latitude stratosphere in stratosphere–troposphere coupling, our study suggests that in order to better understand climate change at the earth’s surface, it may be necessary to more realistically simulate the climate of the stratosphere.
Acknowledgments. We thank Drs. R. Ueyama, C. M. Johanson, and D. L. Hartmann for useful discussions. We acknowledge the modeling groups, the Program for Climate Model Diagnosis and Intercomparison (PCMDI), and the WCRP’s Working Group on Coupled Modeling (WGCM) for their roles in making available the WCRP CMIP3 multimodel dataset. Support of this dataset is provided by the Office of Science, U.S. Department of Energy. This work is supported by NASA Grant NNX08AG91G and NOAA Grant NA08OAR4310725. J. M. Wallace’s participation in this work is funded by NSF Grant 0318675.

APPENDIX

**EOF/PC Analysis of September/October Temperature**

EOF/PC analysis decomposes the data into orthogonal structures (called EOFs) and sorts them in accordance with the variance explained by each structure. The temporal evolution of each characteristic structure (EOF) is described by the corresponding PC. We performed EOF/PC analysis on the yearly time series of monthly mean September or October $T_4$ with the temporal mean over 1979–2007 removed at each grid point. Three significant EOFs are obtained in both months, which together can explain more than 90% of the total variance. We assigned the physical meaning of each EOF via the correlation between the PCs and the physical indices (viz., eddy heat flux, zonal wind, ozone, and phase indices as defined in section 4) as listed in Table A1.

Since we are more interested in the trend rather than in the interannual variability, we further calculated the contribution to the trend pattern from each EOF mode by multiplying each EOF by the linear trend in the corresponding PC. Given the orthogonality of the modes, we can define the “explained trend fraction” as

$$r_i = \frac{\sum X_i^2}{\sum X^2},$$

where $X_i$ is the trend field produced by the $i$th EOF mode, $X$ is the observed trend field, and $\sum$ represents the area-weighted sum over 45°–82.5°S. This statistic describes the fraction of the observed trend that projects onto the respective EOF modes.

Figure A1 shows the trend patterns projected onto the first three EOFs in September, which together explain more than 80% of the observed trend pattern. The first mode has spatial structure similar to the regression pattern from the eddy heat flux or zonal wind (see Fig. A1a versus Figs. 10b and 11b), and the PC-1 in September is highly correlated with the eddy heat flux and the zonal wind indices (see Table A1). Thus, the first mode in September can be interpreted as representing the warming effect from the strengthened BDC. The second mode (Fig. A1b) has a wavenumber-1 structure that is orthogonal to the climatological-mean wavenumber 1 (Fig. 2c), indicating a phase change in wavenumber 1 relative to the climatological mean. This mode is not strongly correlated with either eddy heat flux or ozone indices, and

![FIG. A1. Components of the September $T_4$ trend pattern for 1979–2007 that are linear congruent with the first three EOFs in units of K decade$^{-1}$. The numbers in the upper-right corner indicate the fraction of the observed trend pattern explained by each mode.](image)

<table>
<thead>
<tr>
<th>BDC</th>
<th>Eddy heat flux</th>
<th>Zonal wind</th>
<th>Ozone</th>
<th>Phase</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sep PC-1</td>
<td>0.78</td>
<td>0.92</td>
<td>-0.33</td>
<td></td>
</tr>
<tr>
<td>PC-2</td>
<td>-0.03</td>
<td>-0.05</td>
<td>0.24</td>
<td></td>
</tr>
<tr>
<td>PC-3</td>
<td>-0.03</td>
<td>-0.19</td>
<td>0.47</td>
<td></td>
</tr>
<tr>
<td>Oct PC-1</td>
<td>0.50</td>
<td>0.87</td>
<td>-0.81</td>
<td>0.31</td>
</tr>
<tr>
<td>PC-2</td>
<td>-0.21</td>
<td>-0.27</td>
<td>-0.28</td>
<td>0.93</td>
</tr>
<tr>
<td>PC-3</td>
<td>-0.35</td>
<td>-0.03</td>
<td>-0.18</td>
<td>-0.03</td>
</tr>
</tbody>
</table>
has a very weak trend since 1979. The trend pattern of the third mode in September agrees reasonably well with the ozone regression (Figs. 10c and 11c) and the ozone trend (Fig. 3i). PC-3 is also statistically significantly correlated with the ozone index. Thus, this mode can be largely interpreted as radiative cooling effect due to ozone depletion. In the third mode (Fig. A1c), there is also a subsidiary wavenumber-1 component that is in phase with the climatological-mean wavenumber 1 (see Fig. 2c). This is reflected in the weak warming trend in EOF-3 (Fig. A1c), indicating a slight increase in the amplitude of quasi-stationary wavenumber 1 in September.

Figure A2 shows the trend patterns projected onto the first three October EOF modes, which together explain about 95% of the observed trend. The first mode (Fig. A2a) exhibits strong zonal symmetry, and its PC is strongly correlated with both the ozone and the eddy heat flux indices. This is not a surprising result since the responses of temperature to the trends in zonal wind and ozone (radiative part) both exhibit zonally symmetric components. We hence interpreted the trend in the first mode as the zonally symmetric part of the radiative cooling that is largely cancelled by the adiabatic warming in the descending branch of the BDC.

The second and third October modes both involve wavenumber 1. The trend in the second mode exhibits a spatial structure similar to the phase change component of the $T_4$ trend. PC-2 is highly correlated with the phase index. Hence, we interpreted this mode as representing the phase shift effect. Since this mode accounts for ~84% of the observed temperature trend, it is suggested that the phase shift in October plays a dominant role in the temperature trend pattern on the decadal time scale. The third mode is in phase with the climatological-mean wavenumber 1, and hence it represents changes in the amplitude of stationary wavenumber 1. This mode does not have a significant correlation with the eddy heat flux index, the zonal wind index, or the ozone index. It has changed little on decadal time scales, and thus contributes a very small fraction of the observed trend pattern.

REFERENCES


CORRIGENDUM

PU LIN AND QIANG FU

Department of Atmospheric Sciences, University of Washington, Seattle, Washington

SUSAN SOLOMON

Chemical Sciences Division, NOAA/Earth System Research Laboratory, Boulder, Colorado

JOHN M. WALLACE

Department of Atmospheric Sciences, University of Washington, Seattle, Washington

Due to communication and production errors, Lin et al. (2009) was mistakenly published with black and white figures throughout, instead of color figures. To correct this the following pages contain the full article as it should have appeared, with the color figures. The staff of the Journal of Climate regrets any inconvenience these errors may have caused.

REFERENCE

Temperature Trend Patterns in Southern Hemisphere High Latitudes: Novel Indicators of Stratospheric Change

PU LIN AND QIANG FU
Department of Atmospheric Sciences, University of Washington, Seattle, Washington

SUSAN SOLOMON
Chemical Sciences Division, NOAA/Earth System Research Laboratory, Boulder, Colorado

JOHN M. WALLACE
Department of Atmospheric Sciences, University of Washington, Seattle, Washington

(Manuscript received 11 December 2008, in final form 16 June 2009)

ABSTRACT

Robust stratospheric temperature trend patterns are suggested in the winter and spring seasons in the Southern Hemisphere high latitudes from the satellite-borne Microwave Sounding Unit (MSU) measurement for 1979–2007. These patterns serve as indicators of key processes governing temperature and ozone changes in the Antarctic. The observed patterns are characterized by cooling and warming regions of comparable magnitudes, with the strongest local trends occurring in September and October. In September, ozone depletion induces radiative cooling, and strengthening of the Brewer–Dobson circulation (BDC) induces dynamical warming. Because the trends induced by these two processes are centered in different locations in September, they do not cancel each other, but rather produce a wavelike structure. In contrast, during October, the ozone-induced radiative cooling and the BDC-induced warming exhibit a more zonally symmetric structure than in September, and hence largely cancel each other. However, the October quasi-stationary planetary wavenumber 1 has shifted eastward from 1979 to 2007, producing a zonal wavenumber-1 trend structure, which dominates the observed temperature trend pattern.

Simulated temperature changes for 1979–2007 from coupled atmosphere–ocean general circulation model (AOGCM) experiments run for the Intergovernmental Panel on Climate Change Fourth Assessment Report (IPCC AR4) are compared with the observations. In general, the simulated temperature changes are dominated by zonally symmetric ozone-induced radiative cooling. The models fail to simulate the warming in the southern polar stratosphere, implying a lack of the BDC strengthening in these models. They also fail to simulate the quasi-stationary planetary wave changes observed in October and November.

1. Introduction

The high-latitude stratosphere is a critical component of the climate system. One of the most significant issues for global change is the development of the Antarctic ozone hole during the late twentieth century and the expected recovery during the twenty-first century. The local lower-stratospheric temperature affects the formation of ozone hole through both chemical and dynamical processes. It regulates the occurrence of polar stratospheric clouds (PSCs) that provide the surfaces on which heterogeneous chemical reactions occur (Solomon et al. 1986). It also reflects the strength of the polar vortex, which provides the isolated environment in which the chemical reactions that lead to the development of the ozone hole take place (e.g., McIntyre 1989; Schoeberl and Hartmann 1991).

It has also been realized recently that the stratosphere plays a much more active role in stratosphere–troposphere coupling than previously believed (Haynes 2005; Baldwin et al. 2007). Observational and modeling evidence shows that the change in the stratospheric circulation may lead to a response in both Northern and Southern Hemisphere
annular modes (NAM and SAM, respectively), and thereby influence surface weather and climate (e.g., Thompson and Wallace 2000; Thompson and Solomon 2002; Hartmann et al. 2000; Gillett and Thompson 2003). Such downward dynamical links are particularly important in high latitudes (Baldwin et al. 2007).

The spatial distribution of the long-term changes in stratospheric temperature in the Southern Hemisphere (SH) high latitudes has not been well documented. The presence of large natural variability in this region requires long and reliable observational records to identify a climate signal. Furthermore, the network of radiosonde observations is too sparse to reliably define the spatial patterns. Satellite measurements, which provide complete spatial coverage since 1979, are the primary data used in this study. Here we show the robust spatial trend patterns of stratospheric temperature in the SH high latitudes, which are consistent with previous studies of the zonal averages (e.g., Randel et al. 2009; Thompson and Solomon 2002), but reveal additional insights.

The state-of-the-art coupled atmosphere–ocean general circulation models (AOGCMs) that are the basis for the projections in the Intergovernmental Panel on Climate Change Fourth Assessment Report (IPCC AR4) tend to focus on the troposphere, and generally have poorly resolved stratospheres. Coupled chemistry–climate models simulate the stratosphere more realistically, but may still have problems accurately simulating dynamical processes such as wave propagation and gravity wave generation (Austin et al. 2003; Haynes 2005). Studies of Shine et al. (2003), Ramaswamy et al. (2006), Dameris et al. (2005), and others have attributed the observed global-mean lower-stratospheric cooling mainly to ozone depletion as well as to increasing greenhouse gas concentrations. Previous comparisons between simulations and observations in the SH high latitudes have been carried out using zonally averaged fields (Shine et al. 2003; Ramaswamy et al. 2006; Cordero and Forster 2006). Recently, based upon satellite-born Microwave Sounding Unit (MSU) channel 4 ($T_4$) observations, Johanson and Fu (2007) reported that about half of the lower stratosphere over Antarctica has warmed since 1979 during winter and spring seasons despite the expected cooling due to ozone loss. Hu and Fu (2009) further showed evidence of even stronger warming in trends for individual calendar months. They attributed such warming to the strengthening of the Brewer–Dobson circulation (BDC), and linked this dynamical heating to tropical sea surface temperature (SST) increases. We show here that the trend patterns can exhibit strong zonal asymmetries even though the climatology in this region is dominated by the zonally symmetric component of the flow. It is also of interest to compare model estimates to the observed patterns to gain insight into the processes driving stratospheric change and to assess the fidelity of their representation in models.

In this paper, we present an observational analysis of the stratospheric temperature trends in the SH, focusing on their spatial patterns and seasonal evolution. We compare the stratospheric temperature change with the ozone change on the decadal time scale, and partition it into the zonally symmetric and zonally asymmetric components. The contributions to the observed temperature trend pattern from ozone-depletion-induced radiative cooling and the BDC-strengthening-induced dynamic warming are evaluated by multiple regression. In addition to the radiative cooling and the dynamical warming, we find that changes in the quasi-stationary planetary waves also contribute to the temperature change on decadal time scales. We compare the simulated temperature trend patterns from IPCC AR4 models with observations. These simulations are found to substantially underestimate the dynamical responses to climate change.

This paper is organized as follows. Section 2 describes the data used in the study. Section 3 discusses the observed stratospheric temperature and ozone trend patterns in the winter and spring seasons. Section 4 discusses the processes that contribute to the trend patterns. Section 5 compares the simulated trend pattern with the observations. A summary and the conclusions are given in section 6.

2. Data

For the temperature analysis, we used the MSU/Advanced Microwave Sounding Unit (AMSU) lower-stratospheric channel ($T_4$) monthly gridded (2.5° × 2.5°) brightness temperature data version 3.0 compiled by the Remote Sensing System (RSS) team for 1979–2007 (Mears et al. 2003). The $T_4$ weighting function extends from −20 to −120 hPa and peaks at around 60 hPa, thus well representing the lower stratosphere. Herein, we focus on the SH high latitudes from 45° to 82.5°S, the poleward limit of the MSU measurements.

To examine the ozone trend patterns, we used the version 8 monthly mean total column ozone data from the Total Ozone Mapping Spectrometer (TOMS). This version is formatted on a 1° × 1.25° grid, and is available since 1979, except for 1993–95. Because of the lack of solar radiation, there are no measurements over a relatively large area near the Pole during the winter season. A concern regarding the TOMS trend analysis is the uncorrectable scan-mirror problem that started in about 2000 [(World Meteorological Organization) WMO 2007, chapter 3]. The merged TOMS/Solar Backscatter Ultraviolet (SBUV) dataset constructed by the TOMS science
team (Stolarski et al. 2006) is regarded as a more reliable record that has the benefit of external calibration. The TOMS/SBUV data also partly fill the data gap between 1993 and 1995, but the resolution is much lower ($5^\circ \times 10^\circ$). In our study, we found that the spatial patterns of the trends observed since 1979 do not differ significantly between the two datasets. Thus, we used the TOMS dataset to produce trend maps in order to take advantage of its high spatial resolution, whereas the TOMS/SBUV dataset was used for generating area-weighted spatially averaged ozone time series.

The National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis data (Kalnay et al. 1996) and the Japanese 25-year reanalysis (JRA-25) data (Onogi et al. 2007) are used to calculate the eddy heat flux and a zonal wind index, to represent the strength of the BDC as discussed in section 4. To compare with satellite measurements, we used monthly mean data from 1979 to 2007.

We also used the AOGCM simulations that were performed in support of the IPCC AR4 from the World Climate Research Programme (WCRP) Coupled Model Intercomparison Project phase 3 (CMIP3) multimodel dataset archive. Twenty-two AOGCM simulations contain 48 ensembles, which all considered the long-lived greenhouse gas increases. Thirteen models composing 28 ensembles also considered the stratospheric ozone depletion. A short summary of the model descriptions is listed in Table 1. Two sets of simulations are combined to get a full record from 1979 to 2007: one is the climate of the twentieth-century experiment (20C3M), the other is the set of SRES A1B simulations, which are initialized with conditions from the corresponding 20C3M ensembles at the end of the twentieth century. The ensembles in the 20C3M experiment that were not extended into SRES A1B were discarded.

The standard output for these models is at 17 vertical levels in the atmosphere (i.e., 1000, 925, 850, 700, 600, 500, 400, 300, 250, 200, 150, 100, 70, 50, 30, 20, and 10 hPa). Simulated $T_4$ is derived by vertically integrating the temperature profile with the MSU channel 4 weighting function. The horizontal resolution differs from model to model. To produce the model mean trend map, we interpolate the simulated data onto the grid format of MSU $T_4$ ($2.5^\circ \times 2.5^\circ$) using a cubic-spline interpolation.

### Table 1. Model description.

<table>
<thead>
<tr>
<th>Model ID</th>
<th>Ensemble No.</th>
<th>Ozone forcing</th>
</tr>
</thead>
<tbody>
<tr>
<td>BCCR-BCM2.0</td>
<td>1</td>
<td>N</td>
</tr>
<tr>
<td>CCSM3.0</td>
<td>6</td>
<td>Y</td>
</tr>
<tr>
<td>CGCM3.1(T47)</td>
<td>5</td>
<td>N</td>
</tr>
<tr>
<td>CGCM3.1(T63)</td>
<td>1</td>
<td>N</td>
</tr>
<tr>
<td>CNRM-CM3</td>
<td>1</td>
<td>Y</td>
</tr>
<tr>
<td>CSIRO-MK3.0</td>
<td>1</td>
<td>Y</td>
</tr>
<tr>
<td>CSIRO-MK3.5</td>
<td>1</td>
<td>Y</td>
</tr>
<tr>
<td>ECHAM5/MPI-OM</td>
<td>4</td>
<td>Y</td>
</tr>
<tr>
<td>FGOALS-g1.0</td>
<td>3</td>
<td>N</td>
</tr>
<tr>
<td>GFDL-CM2.0</td>
<td>1</td>
<td>Y</td>
</tr>
<tr>
<td>GFDL-CM2.1</td>
<td>1</td>
<td>Y</td>
</tr>
<tr>
<td>GISS-AOM</td>
<td>2</td>
<td>N</td>
</tr>
<tr>
<td>GISS-EH</td>
<td>3</td>
<td>Y</td>
</tr>
<tr>
<td>GISS-ER</td>
<td>4</td>
<td>Y</td>
</tr>
<tr>
<td>INGV-SXG</td>
<td>1</td>
<td>N</td>
</tr>
<tr>
<td>INM-CM3.0</td>
<td>1</td>
<td>N</td>
</tr>
<tr>
<td>IPSL-CM4</td>
<td>1</td>
<td>N</td>
</tr>
<tr>
<td>MIROC3.2(hires)</td>
<td>1</td>
<td>Y</td>
</tr>
<tr>
<td>MIROC3.2(medres)</td>
<td>3</td>
<td>Y</td>
</tr>
<tr>
<td>MRI-CGCM2.3.2</td>
<td>5</td>
<td>N</td>
</tr>
<tr>
<td>UKMO-HadCM3</td>
<td>1</td>
<td>Y</td>
</tr>
<tr>
<td>UKMO-HadGEM1</td>
<td>1</td>
<td>Y</td>
</tr>
</tbody>
</table>

### 3. Monthly trend patterns in winter and spring seasons

**a. Observed features in the monthly climatologies**

Figure 1 shows the monthly climatological mean MSU $T_4$ and total ozone maps for 1979–2007 from July to December. During winter, the lower-stratospheric temperature field is dominated by a zonally symmetric cold pool (Figs. 1a,b), which implies a strong polar vortex above through thermal wind balance. Beginning in September, the zonal wind weakens as the polar stratosphere warms because of the absorption of the solar radiation. As the zonal wind velocity becomes smaller than the Rossby critical velocity (Holton 2004), quasi-stationary planetary waves (mainly the zonal wavenumber-1 component) are able to propagate from the troposphere into the stratosphere, and the polar vortex becomes displaced from the South Pole toward the Atlantic sector (e.g., Waugh and Randel 1999; Karpetchko et al. 2005). By December, the polar vortex is completely gone, and the zonal wind is too weak to allow the quasi-stationary waves to propagate upward from the troposphere, and thus the temperature field becomes zonally symmetric again (Fig. 1f). The ozone maps show that the low ozone area is generally confined to the shape of the polar vortex, except that its minimum occurs in September and October rather than in winter, since solar radiation is a necessary condition for ozone depletion. Nevertheless, the ozone field exhibits the same seasonal evolution as the temperature field, that is, from a more zonally symmetric state in July and August (Figs. 1g,h) to greater zonal asymmetry in September–November (Figs. 1i–k), and back to a more zonally symmetric state again in December (Fig. 1l).

The switch between the zonally symmetric and asymmetric states is more readily observed in the “eddy” fields, which are defined as
where $X$ is the original field, $[X]$ is the zonal-mean field, and $X'$ is the eddy field. As shown in Fig. 2, for both stratospheric temperature and ozone, the eddy component is quite weak in winter. During September and October, when the zonal wind is favorable for the vertical propagation of the quasi-stationary wavenumber 1, the eddy component of the temperature and ozone fields exhibits a strong wavenumber-1-like structure. As the zonal wind continues to decrease, the eddy fields become weak again.

b. Observed features in the trend patterns

The monthly MSU $T_4$ trend patterns in the SH high latitudes in austral winter and spring seasons are shown in Figs. 3a–f. The monthly trend maps exhibit several interesting characteristics. First, the strong regional warming trends are largely confined to the three months from August to October, whereas the trends in July are fairly weak, and in November and December the trends are dominated by cooling. Strong cooling is also observed over part of the hemisphere in October. Second, the temperature trend patterns tend to be organized largely into zonal wavenumber-1-like patterns, with a clear phase shift of about 90° between September and October. The warming trends in the monthly data are about 2–3 times stronger than in the seasonal-mean data, with regional magnitudes of up to 3 K decade$^{-1}$ occurring in September and October.

Figures 3g–l show the monthly trend patterns for total column ozone, which are dominated by negative trends in all winter and spring months, with strongest depletion during the “ozone hole” season, September–November. Ozone-induced radiative cooling is expected to be strongest during these months (Randel and Wu 1999). Ozone-induced radiative effects, on the other hand, cannot directly account for the strong warming, since the warming occurs in regions of observed ozone decrease.

Figure 4 shows the eddy component of the temperature and ozone trend patterns. The number at the upper-right corner indicates the area-weighted relative contribution of the eddy component to the total trends. This fraction is defined as

$$ f = \frac{\sum X'^2}{\sum X^2}, $$

where $X'$ and $X$ are the eddy component and the total trends, respectively, and $\sum$ represents the area-weighted sum over the southern high latitudes ($\geq 45^\circ$S). The eddy components of both temperature and ozone trends are dominated by zonal wavenumber-1 structures in nearly all winter and spring months. Close agreement is seen in
The eddy component of the trend patterns for temperature and ozone in terms of both the relative magnitudes and the phase. The coherent ozone and temperature changes during the SH spring season have also been documented in previous studies (e.g., Newman and Randel 1988; Randel and Cobb 1994; Randel and Wu 1999). Such good agreement between these two independent observations confirms the reliability of the trend analysis.

The strongest warming trend occurs in September and October, which is also the season during which the climatological-mean zonal asymmetry (Figs. 2c,d) and quasi-stationary wave activity are the strongest (Randel 1988). The eddy components of the climatological mean fields and the trends are both dominated by a wave-number-1 structure. In August and September, the phase of wavenumber 1 in the climatology and in the trend is almost the same. However, in October and November, the phase in the trend is more than 90° to the east of that in the climatology. This difference suggests that the cause of the trend pattern observed in October and November may be different from that in September and August.

Figure 5 shows the monthly decadal mean $T_4$ climatologies averaged over 1979–88, 1989–98, and 1999–2007. Comparing the climatologies for these three decades, there is no obvious change in July and August. Gradual warming is seen in September over the subpolar region on the Australian side. In October and November, departures from the zonally symmetric state are clearly discernible. However, unlike the situation in September, the wave pattern in October and November does not remain in place, but gradually shifts eastward. The phase shift is most apparent in November and in the most recent decade.

Comparing the eddy component of the trends with the total trends, it is clear that the eddy contribution accounts for most of the local $T_4$ trend patterns except in November and December (see Figs. 4a–f versus Figs. 3a–f). The eddy component is especially strong in August, September, and October, the months when the warming is strong. It is then not surprising that little or no warming is reported in the zonal mean trends for these months (e.g., Thompson and Solomon 2005; Randel et al. 2009), because the warming is largely canceled by cooling of comparable magnitude in the zonal average.

The eddy contribution to the ozone trend is however much less than in $T_4$ in all winter and spring months. This difference indicates different behaviors of the zonally symmetric components of temperature and ozone on decadal time scales (i.e., little trend for temperature and a strong decreasing trend for ozone). Since a large zonally symmetric decrease in ozone would lead to a strong...
radiative cooling of the stratosphere, a compensating zonally symmetric, dynamically induced warming is required to account for the observed absence of a zonal-mean stratospheric cooling trend (Thompson and Solomon 2009).

c. The robustness of the trend pattern

Considering the large variability in polar region during the late winter–spring season, and the unusual sudden stratospheric warming that occurred in Antarctica in 2002, one might question whether the observed trend patterns are statistical significant. Figure 6 shows the $T_4$ trend for August, September, October, and November from 1979 onward with different ending year. Consistent trend patterns are seen regardless of the ending years, indicating the observed trend patterns are not unduly influenced by the effect of unusual years such as 2002. The cooling trends in October and November are stronger and cover a larger area during early years, which is consistent with strong ozone depletion during the early part of the record (WMO 2007). Despite the stronger cooling over the polar cap region as a whole in these two months in the early years, the spatial structures of the trend patterns are nevertheless quite similar.

To further establish the statistical significance of the trend pattern, we carried out Monte Carlo tests by randomly switching the order of years, and calculating trends in these scrambled $T_4$ time series. By repeating such Monte Carlo simulations enough times, we can obtain the distribution of trends that can possibly be achieved by natural variability. In contrast to the Student’s $t$ test of trends at each grid point as done by Hu and Fu (2009), we focused on the robustness of the whole spatial trend pattern. As shown in Figs. 3a–f, the zonal wavenumber-1 component dominates the observed trend structure. We thus computed the wavenumber-1 component in those simulated Monte Carlo trends for each month by performing Fourier decomposition on each latitude circle and then averaging over 50°–70°S weighted by the cosine of the latitude. Figure 7 shows the distribution of the amplitudes of these mean wavenumber-1 trends among 10 000 Monte Carlo simulations. The possibility of achieving a wavenumber-1 trend structure as large as observed by chance is less than 1% in October. Thus, the observed temperature trend patterns are indeed strongly statistically significant. In other months, the observations are likely to be significant.

4. Interpretation of the trend patterns

In this section, we examine the $T_4$ trend patterns using multiple regression. We did not detrend the data before the regression analysis. Two physical processes are considered to be responsible for the high-latitude stratospheric temperature change. One is related to the absorption of solar radiation by the stratospheric ozone that warms the lower stratosphere. We defined an “ozone index” as the area-weighted spatial-mean total ozone over 45°S poleward to represent the ozone effect. The other process is associated with the BDC, which has a descending branch in high latitudes that warms that region adiabatically. The BDC is driven remotely by planetary and gravity wave breaking in the extratropical stratosphere, which acts like a “suction pump,” drawing air upward from the tropics (Haynes et al. 1991; Holton et al. 1995). The waves originate in the mid- to high-latitude troposphere, propagate upward, and break in the stratosphere, depositing negative momentum and decelerating the stratospheric jet. Thus, when the BDC becomes stronger, indicative of more wave activity propagating upward into the stratosphere, the stratospheric jet should become weaker, and vice versa.
A conventional index of the strength of the BDC is the eddy heat flux in the lower stratosphere (e.g., Hu and Tung 2002; Haklander et al. 2008), which is equivalent to the vertical component of the Eliassen–Palm (EP) flux, representing the amount of the wave activity propagating from the troposphere to the stratosphere (Andrews et al. 1987). We here defined an “eddy heat flux index” as the 3-month mean area-weighted average eddy heat flux at the 150-hPa level over 45°–90°S. Wave activity in the previous two months is also included because it can contribute to the dynamical heating in the current month (Hu and Tung 2002). Because of the limited resolution of the reanalysis data, this calculated eddy heat flux only takes into account the resolved waves, but misses the waves on the subgrid scale (e.g., gravity waves).

The stratospheric jet, feeling the total effects from wave breaking on all scales, also provides an indication of the strength of the BDC. Following Ueyama and Wallace (2010), we defined a “zonal wind index” as another index for the strength of the BDC, which is the

![Figure 6: Monthly $T_4$ trend maps for August–November from 1979 to various ending years as indicated on the left.](image-url)
We perform multiple regression of gridded monthly-mean $T_4$ data upon the two indices in the following form:

\[ T(x, y, t) = K_{O_3}(x, y) \times X_{O_3}(t) + K_{BDC}(x, y) \times X_{BDC}(t) + \text{residual}, \]

where $X_{O_3}$ is the standardized ozone index, and $X_{BDC}$ is the standardized eddy heat flux index or zonal wind index. The regression maps $K_{O_3}$ and $K_{BDC}$ represent patterns of temperature anomalies that correspond to a one standard (positive) anomaly deviation in the indices assuming a linear relationship. The contributions to the $T_4$ trend attributable to changes in ozone concentration and the strength of the BDC are thus obtained by multiplying the regression maps by the linear trend in the corresponding standardized indices.
We calculate the eddy heat flux and the zonal wind indices using the NCEP–NCAR reanalysis data. Figure 8 shows the monthly $T_4$ trend derived from the NCEP–NCAR reanalysis, which agrees very well with the satellite observations (Figs. 3a–f). This close agreement lends confidence to the analysis of trends in wave activity based on the NCEP–NCAR reanalysis. We also derived the indices from the JRA-25 reanalysis data. The resulting regression maps and trend patterns (not shown) are very similar to those derived from the NCEP–NCAR reanalysis.

We focus on September and October because these are the months in which the trend patterns are most significant. Furthermore, August $T_4$ exhibits a behavior similar to September, and November is more like October. Similar results can be obtained from empirical orthogonal functions/principal component (EOF/PC) analysis, as shown in the appendix.

**a. September**

Figures 9a,c show the time series of the eddy heat flux and ozone indices for September. The corresponding regression maps are shown in Figs. 9b,d expressed in units of kelvin per unit of standard deviation of the eddy heat flux or ozone.

The polar stratospheric temperature response to the eddy heat flux is dominated by negative values (Fig. 9b), where stronger poleward (more negative) eddy heat flux implies more upward propagation of wave activity, a stronger BDC, and stronger subsidence-induced warming. In contrast, the temperature response to ozone is dominated by positive values (Fig. 9d), because more ozone will absorb more solar radiation and will thus warm the stratosphere.

It is notable that the temperature responses to the variations in spatial-mean ozone and eddy heat flux are not zonally symmetric, and their primary centers are at different locations. The eddy heat flux response (Fig. 9b) is centered along the Antarctic coast poleward of Australia. The zonal asymmetry reflects the fact that the center of the polar vortex in September is generally displaced from the Pole toward South America (see Fig. 1c). Since sinking is inhibited near the center of the strong polar vortex (e.g., Manney et al. 1994, 1999), the descending
branch of the BDC tends to be concentrated on the other side of the Antarctica (i.e., toward Australia). Such zonal asymmetry in the distribution of the descending air is also evident in the ozone distribution, because the BDC advects ozone-rich air from the tropical region. The sinking region has been referred to in previous studies as the “ozone croissant” or the “ozone collar” (e.g., Grytsai et al. 2007; Mariotti et al. 2000).

The maximum amplitude of the ozone response (Fig. 9d) is located near (65°S, 15°E). This pattern is similar to the ozone trend pattern as shown in Fig. 3i, but is of opposing sign.

Figure 10a shows the observed September $T_4$ trend (same as Fig. 3c), and Figs. 10b,c show the contributions to the $T_4$ trend in the same month due to the changes in the eddy heat flux and ozone, respectively. The least squares best-fit linear combination of the two is shown in Fig. 10d. Since 1979, ozone concentration has decreased (Fig. 9c), resulting in a large area of radiative cooling (Fig. 10c). At the same time, the poleward eddy heat flux has strengthened in association with a strengthening of the BDC (Fig. 9a), which causes warming centered at the edge of the Antarctica toward the Australian side (Fig. 10b). A strengthening of the BDC has been predicted in response to global warming and ozone depletion in many climate models (e.g., Rind et al. 1998; Eichelberger and Hartmann 2005; Butchart et al. 2006; Li et al. 2008). Recent observational records also indicate an increased cooling in the tropical lower stratosphere associated with the strengthening of the BDC (Randel et al. 2006; Rosenlof and Reid 2008). The SH high-latitude stratospheric warming identified in this study, provides observational evidence that such a strengthening is, in fact, occurring.

The linear combination of the eddy heat flux- and ozone-driven trends captures the observations extremely well (Fig. 10d versus Fig. 10a), and the residual, defined as the observed trend minus the combined trend, is small over most of SH high latitudes (Fig. 10e). The close agreement suggests that the temperature change on the decadal time scale is indeed largely driven by the direct radiative impact of ozone depletion and the strengthening of the BDC. Also worth mentioning is the fact that the observed $T_4$ trend structure dominated by the wavenumber-1 component may not necessarily be caused by amplification of the planetary waves. Rather, it could be because the radiative response in temperature to ozone depletion and the dynamical response to the BDC have exhibited opposing tendencies during the past few decades, with different primary action centers. Thus, the cooling associated with decreasing ozone and the warming associated with the strengthening of the BDC would not cancel each other, but form a wavenumber-1-like structure instead.

There is a weak cooling in the September residual trend centered to the west of the Antarctic Peninsula, which is collocated with the secondary center of ozone depletion (see Fig. 3i), suggesting that the radiative cooling due to ozone depletion may be slightly underestimated by the multiple regression method. Note that the observed cooling could also be partially linked to the residual radiative cooling due to the ozone depletion in the previous year and the radiative effect of the increase of greenhouse gases, although these effects may be small at the level we considered.

Figure 11 shows the contributions to the $T_4$ trend due to changes in the BDC and ozone as represented by the zonal wind and ozone indices. They are similar to those based on eddy heat flux and ozone indices.

For August, multiple regression analysis yields patterns similar to those for September but with weaker trends. The seemingly wavenumber-1 trend structure in August can also be separated into two parts: radiative cooling due to ozone depletion and dynamic warming due to the strengthening of the BDC.

b. October

Unlike in September, the phase of wavenumber 1 in October has shifted eastward substantially during the last three decades as shown in Fig. 5. This shifting has certainly affected $T_4$ trend pattern. To separate the $T_4$ trend due to the phase shift from those due to the change in BDC and ozone-induced radiative effect, we added a third index in October to describe the phase shift of the
planetary wave. Here a “phase index” is defined as the longitude of the crest of the mean zonal wavenumber-1 component, which is averaged over 50°–70°S, weighted by the cosine of latitude.

Figures 12a,c,e show the time series of the eddy heat flux, ozone, and phase indices in October, and the corresponding regression maps are shown in Figs. 12b,d,f in units of kelvin per unit of standard deviation. The corresponding contributions to the trend in October $T_4$ are shown in Fig. 13.

The temperature response to changes in eddy heat flux in October (Fig. 12b) is negative as in September, but exhibits a primary action center located over the Ross Sea. Compared with September, this temperature response exhibits a more prominent zonally symmetric component with the primary center closer to the Pole.

The temperature response to ozone concentration variation in October (Fig. 12d) is positive as in September. It exhibits a quasi-zonal symmetric structure that is slightly displaced off the Pole toward western Atlantic sector, which is the same position as the climatological polar vortex (Figs. 1d,j).

The temperature response to the phase change of wavenumber 1 in October (Fig. 12f) exhibits zonal wavenumber-1 structure as expected. This wavenumber-1 structure is orthogonal to the climatological mean wavenumber 1. A positive phase index is associated with an eastward shift relative to the climatological mean state.

Fig. 12. (a)–(d) As in Fig. 9, but for October. (e) Time series of the phase index (solid line) and its linear trend (dashed line) in October. (f) Map of October $T_4$ regressed onto the phase index in units of K per standard deviation. The trend uncertainty is given at the 95% confidence level ($2\sigma$). See text for the definition for the phase index.
During the past few decades, the strengthened BDC has induced strong warming (Fig. 13b), and ozone depletion has induced strong cooling (Fig. 13c) in October. These dynamical warming and radiative cooling patterns are more zonally symmetric than their counterparts in September, so they cancel each other to a large extent. In addition, phase of wavenumber 1 has shifted eastward by about 15° decade⁻¹ (Fig. 12e). This shifting induces a temperature trend with a wavenumber-1 structure (Fig. 13d) that is similar to the observed total temperature trend (Fig. 13a). This trend pattern is consistent to that derived by rotating the climatological temperature field following the phase of the wavenumber-1 component in each year of last three decades (not shown). A similar phase change of wavenumber 1 is also observed in ozone field (Grytsai et al. 2007).

Previous studies (Quintanar and Mechoso 1995; Inatsu and Hoskins 2004) showed that the quasi-stationary zonal wavenumber 1 in the SH upper troposphere and lower stratosphere is primarily thermally forced by zonal asymmetries in the tropical SST distribution. While the tropospheric forcing has remained relatively constant, the phase of the planetary waves could have been affected by changes in the mean state of the atmosphere, especially in the zonal wind field, as the wave propagates upward (e.g., O’Neill and Youngblut 1982; Karoly and Hoskins 1982; Nishii and Nakamura 2004). In this case, the phase index in October is found to be significantly correlated with the ozone index ($r = -0.53$), but not with the eddy heat flux index or zonal wind index. The time series of phase index also exhibits strong peaks in the years 1982 and 1992, immediately after the major volcanic eruptions of El Chichón and Pinatubo. These results suggest that concentrations of ozone and volcanic aerosols may affect the phase of the planetary wavenumber 1 in the lower stratosphere, though further research is required to understand the mechanism responsible to this eastward phase shift.

The linear combination of the three responses (Fig. 13e) captures the observed $T_4$ trend very well, with a very small residual (Fig. 13f).

We also performed the multiple regression on $T_4$ with time series of the zonal wind, ozone, and phase indices as predictors. As shown in Fig. 14, the spatial structures of the corresponding regression maps were found to be similar to those derived from the eddy heat flux index. Multiple regression of $T_4$ upon the time series of zonal wind and ozone concentration only (i.e., without the phase index) can also capture the observation well. But in this case, the temperature trend associated with ozone change is a combination of radiative cooling and that due to wavenumber-1 rotation, which is consistent with the summation of Figs. 14c,d. This is not a surprising result, because the phase index does have a significant correlation with the ozone index.

The observed $T_4$ trend pattern in November can also be separated into three parts as in October: quasi-zonally symmetric patterns associated with changes in ozone concentration and the strength of the BDC, and a pattern associated with eastward phase shift of zonal wavenumber 1. Compared with October, the trend of the BDC is much weaker. The radiative effect induced by ozone depletion dominates in November.

5. Comparison between models and observations

In this section, we will compare the $T_4$ trend patterns deduced from IPCC AR4 AOGCMs with observations.
Figure 15a shows monthly trend maps from the ensemble model means, which are dominated by zonally symmetric cooling throughout winter and spring, without any regional patches of warming. The models seem to capture the temperature trend only from the direct radiative forcing. The maximum cooling is found in November, about 1 month later than the observed maximum ozone loss in the lower stratosphere.

We further separate the models into two groups based on how they treat the stratospheric ozone forcing. Figure 15b shows the results from the models with time-varying ozone including ozone depletion based on observations. The results shown in Fig. 15c are from models that only consider the climatological-mean seasonal cycle of stratospheric ozone but ignore any long-term change. In both cases, the monthly trend maps are dominated by cooling, though the cooling in models without ozone depletion is much weaker. Previous attribution experiments (e.g., Shine et al. 2003; Ramaswamy et al. 2006; Dameris et al. 2005) suggested that the cooling observed in the polar lower stratosphere can be mainly attributed to ozone depletion, although the increase of well-mixed greenhouse gas and the increase of stratospheric water vapor may also play a role. Thus, the weak cooling in the no-ozone-depletion group can be explained by greenhouse gas increases, whereas the strong cooling in the models considering ozone-depletion includes contributions from both ozone depletion and greenhouse gas increases.

Next we remove the zonally symmetric component of the simulated trend to check whether there is any wavenumber-1-like structure analogous to that in the observations. As shown in Fig. 16, the eddy component in models indeed tends to be organized into a wave-number-1-like shape, though not as well defined as in the observations, but the amplitudes are generally about an order of magnitude smaller. The strongest and most well-defined wavenumber-1 structures are seen in November in the all-model ensemble mean as well as in the two subgroups, a month later than in the observations. The phase of the wavenumber-1 trend in the ozone-depletion group does not change much from month to month throughout August to November, with the crest remaining over the southern Pacific. For the no-ozone-depletion group, the phase does change from month to month. Both model groups capture the phase of the wavenumber-1 trend observed in November, but not in other months. In general, the ozone-depletion group does not exhibit any apparent improvement over the no-ozone-depletion group with respect to this behavior.

In contrast to the trend pattern, the climatological mean monthly temperature fields from the all model mean and the two groups are very similar to the observations (not shown). As for the eddy components, the simulated amplitudes are about half as large as in the observations, but the phases are the same. There is no obvious difference between the two model groups.

6. Summary and concluding remarks

Stratospheric temperature trends in the SH high latitudes over the past three decades exhibit well-defined, seasonally dependent spatial structures. Robust wavenumber-1-like structures with both warming and cooling are observed during the late winter and spring seasons, especially in September and October. A phase shift in the trend pattern is observed between September and October. Temperature and ozone exhibit similar patterns.
in the eddy component of trend. However, for the zonally symmetric component, temperature and ozone behave quite differently on the decadal time scale.

We obtained the stratospheric temperature fingerprints in response to the changes in SH planetary wavenumber 1, total ozone (the direct radiative effect), and the strength of the BDC through multiple regression. In addition to the well-known direct radiative cooling induced by ozone depletion, our study indicates that the strengthening of the BDC produced an adiabatic warming of the polar stratosphere. The stationary planetary wavenumber 1 is also found to exhibit a significant change in phase during late spring. The temperature responses to the latter two processes (viz., the global and local dynamical processes) appear to be as strong as or even stronger than the ozone-induced radiative cooling in winter and spring months.

While the trends in $T_4$ in each calendar month are all influenced by radiation, BDC, and planetary waves, the lower-stratospheric temperature trends exhibit different properties in late winter and spring. In August and September, ozone-induced radiative cooling and BDC-induced dynamic warming both play important roles, whereas stationary wavenumber 1 does not exhibit a significant long-term change. The polar vortex is generally displaced from the Pole due to the presence of the planetary wavenumber 1. Thus, the strongest cooling inside the vortex associated with ozone depletion, and the strongest warming outside the vortex associated with the trend toward stronger descending motion, are centered in different locations. Our study has shown that the cooling and warming do not cancel one another, and the sum of the two contributions forms a wavenumber-1-like structure that dominates the trends. This finding has important implications for interpreting trend data and understanding climate change in Antarctic stratosphere.

In October and November, there is a systematic eastward shift in the quasi-stationary wavenumber 1 besides the ozone-depletion-induced radiative cooling and the BDC-strengthening-induced warming. This eastward shift alone leads to a wavenumber-1 temperature trend pattern that is about 90° to the east of the climatological wave-number 1. In October, the phase shift accounts for most of the observed trend, while the asymmetric parts of the radiation and BDC further enhance the wave structure. In November, the radiative effect dominates, and the circulation effect and the phase-shift effect are relatively weak.

While seldom noticed by earlier studies, such an eastward shift in stationary wavenumber 1 certainly affects the spatial structure of regional climate change. Grytsai et al. (2007) related the eastward shift of wavenumber 1 over Antarctica to a horizontal displacement of the atmospheric centers of action under global climate change, and suggested that such change in the stratosphere may be related to changes in the tropospheric planetary waves. Our study indicates that the ozone and volcanic aerosol concentration may play roles in the phase shift of stationary wavenumber 1. While our results provide a first look, the mechanism for the eastward shift of stationary wavenumber 1 requires further investigation. It should be noted that in addition to the SH planetary wavenumber 1, the decadal change of the BDC can also be indirectly driven by the ozone depletion.

The IPCC AR4 ACGCMs do not capture the observed spatial trend pattern in the SH high-latitude stratosphere in the winter and spring seasons. They fail to simulate the response of the BDC to global warming in this region. The long-term changes in these waves are completely missed. Considering the prominence of the high-latitude stratosphere in stratosphere–troposphere coupling, our study suggests that in order to better understand climate change at the earth’s surface, it may be necessary to more realistically simulate the climate of the stratosphere.
Acknowledgments. We thank Drs. R. Ueyama, C. M. Johanson, and D. L. Hartmann for useful discussions. We acknowledge the modeling groups, the Program for Climate Model Diagnosis and Intercomparison (PCMDI), and the WCRP’s Working Group on Coupled Modeling (WGCM) for their roles in making available the WCRP CMIP3 multimodel dataset. Support of this dataset is provided by the Office of Science, U.S. Department of Energy. This work is supported by NASA Grant NNX08AG91G and NOAA Grant NA08OAR4310725. J. M. Wallace’s participation in this work is funded by NSF Grant 0318675.

APPENDIX

EOF/PC Analysis of September/October Temperature

EOF/PC analysis decomposes the data into orthogonal structures (called EOFs) and sorts them in accordance with the variance explained by each structure. The temporal evolution of each characteristic structure (EOF) is described by the corresponding PC. We performed EOF/PC analysis on the yearly time series of monthly mean September or October $T_4$ with the temporal mean over 1979–2007 removed at each grid point. Three significant EOFs are obtained in both months, which together can explain more than 90% of the total variance. We assigned the physical meaning of each EOF via the correlation between the PCs and the physical indices (viz., eddy heat flux, zonal wind, ozone, and phase indices as defined in section 4) as listed in Table A1.

Since we are more interested in the trend rather than in the interannual variability, we further calculated the contribution to the trend pattern from each EOF mode by multiplying each EOF by the linear trend in the corresponding PC. Given the orthogonality of the modes, we can define the “explained trend fraction” as

$$r_i = \frac{\sum X_i^2}{\sum X^2},$$

where $X_i$ is the trend field produced by the $i$th EOF mode, $X$ is the observed trend field, and $\sum$ represents the area-weighted sum over $45^\circ$–$82.5^\circ$S. This statistic describes the fraction of the observed trend that projects onto the respective EOF modes.

Figure A1 shows the trend patterns projected onto the first three EOFs in September, which together explain more than 80% of the observed trend pattern. The first mode has spatial structure similar to the regression pattern from the eddy heat flux or zonal wind (see Fig. A1a versus Figs. 10b and 11b), and the PC-1 in September is highly correlated with the eddy heat flux and the zonal wind indices (see Table A1). Thus, the first mode in September can be interpreted as representing the warming effect from the strengthened BDC. The second mode (Fig. A1b) has a wavenumber-1 structure that is orthogonal to the climatological-mean wavenumber 1 (Fig. 2c), indicating a phase change in wavenumber 1 relative to the climatological mean. This mode is not strongly correlated with either eddy heat flux or ozone indices, and

<table>
<thead>
<tr>
<th>BDC</th>
<th>Eddy heat flux</th>
<th>Zonal wind</th>
<th>Ozone</th>
<th>Phase</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sep</td>
<td>PC-1</td>
<td>0.78</td>
<td>0.92</td>
<td>−0.33</td>
</tr>
<tr>
<td></td>
<td>PC-2</td>
<td>−0.03</td>
<td>−0.05</td>
<td>0.24</td>
</tr>
<tr>
<td></td>
<td>PC-3</td>
<td>−0.03</td>
<td>−0.19</td>
<td>0.47</td>
</tr>
<tr>
<td>Oct</td>
<td>PC-1</td>
<td>0.50</td>
<td>0.87</td>
<td>−0.81</td>
</tr>
<tr>
<td></td>
<td>PC-2</td>
<td>−0.21</td>
<td>−0.27</td>
<td>−0.28</td>
</tr>
<tr>
<td></td>
<td>PC-3</td>
<td>−0.35</td>
<td>−0.03</td>
<td>−0.18</td>
</tr>
</tbody>
</table>

Fig. A1. Components of the September $T_4$ trend pattern for 1979–2007 that are linear congruent with the first three EOFs in units of K decade$^{-1}$. The numbers in the upper-right corner indicate the fraction of the observed trend pattern explained by each mode.
has a very weak trend since 1979. The trend pattern of the third mode in September agrees reasonably well with the ozone regression (Figs. 10c and 11c) and the ozone trend (Fig. 3i). PC-3 is also statistically significantly correlated with the ozone index. Thus, this mode can be largely interpreted as radiative cooling effect due to ozone depletion. In the third mode (Fig. A1c), there is also a subsidiary wavenumber-1 component that is in phase with the climatological-mean wavenumber 1 (see Fig. 2c). This is reflected in the weak warming trend in EOF-3 (Fig. A1c), indicating a slight increase in the amplitude of quasi-stationary wavenumber 1 in September.

Figure A2 shows the trend patterns projected onto the first three October EOF modes, which together explain about 95% of the observed trend. The first mode (Fig. A2a) exhibits strong zonal symmetry, and its PC is strongly correlated with both the ozone and the eddy heat flux indices. This is not a surprising result since the responses of temperature to the trends in zonal wind and ozone (radiative part) both exhibit significant zonally symmetric components. We hence interpreted the trend in the first mode as the zonally symmetric part of the radiative cooling that is largely cancelled by the adiabatic warming in the descending branch of the BDC.

The second and third October modes both involve wavenumber 1. The trend in the second mode exhibits a spatial structure similar to the phase change component of the $T_4$ trend. PC-2 is highly correlated with the phase index. Hence, we interpreted this mode as representing the phase shift effect. Since this mode accounts for ~84% of the observed temperature trend, it is suggested that the phase shift in October plays a dominant role in the temperature trend pattern on the decadal time scale. The third mode is in phase with the climatological-mean wavenumber 1, and hence it represents changes in the amplitude of stationary wavenumber 1. This mode does not have a significant correlation with the eddy heat flux index, the zonal wind index, or the ozone index. It has changed little on decadal time scales, and thus contributes a very small fraction of the observed trend pattern.

**REFERENCES**


——, and F. Wu, 1999: Cooling of the Arctic and Antarctic polar stratospheres due to ozone depletion. J. Climate, 12, 1467–1479.
——, and ——, 2005: Recent stratospheric climate trends as evidenced in radiosonde data: Global structure and tropospheric linkages. J. Climate, 18, 4785–4795.