What controls the temperature of the Arctic stratosphere during the spring?

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Abstract.

The temperature of the polar lower stratosphere during spring is the key factor in changing the magnitude of ozone loss in the Arctic polar vortex. In this paper, we quantitatively demonstrate that the polar lower-stratospheric temperature is primarily controlled by planetary-scale waves. We use NCEP/NCAR reanalysis data covering the last two decades to investigate how these planetary waves are connected to polar lower-stratospheric temperatures. In particular, we show that the tropospheric eddy heat flux in middle to late winter (January–February) is highly correlated with the mean polar stratospheric temperature during March. These planetary waves are forced by both thermal and orographic processes in the troposphere, and propagate into the stratosphere in the middle and high latitudes. Strong mid-winter planetary wave forcing leads to a warmer spring Arctic lower stratosphere in early spring, while weak mid-winter forcing leads to cooler spring Arctic temperatures. In addition, this planetary wave driving also has a strong impact on the strength of the polar vortex.

1. Introduction

The temperature of the polar lower stratosphere is mainly controlled by the variation of the solar declination from the winter to spring solstices, and by the distribution of radiatively active trace gases such as ozone, water, and carbon dioxide [*Shine,* 1987]. However, during the winter, the Arctic lower stratosphere is considerably warmer than would be expected from a pure radiative calculation [*Fels,* 1982]. This temperature difference is due to waves which propagate up from the troposphere into the stratosphere, where they dissipate. This wave dissipation causes a poleward and downward circulation which acts to warm the polar region and drive it away from the radiative equilibrium [*Andrews et al.,* 1987]. In addition, these waves critically impact concentrations of stratospheric ozone.

Brewer [1949] and *Dobson* [1956] first illustrated the long-term slow-rising circulation in the tropics and the slow descent in the extratropics. This polar descent via the Brewer-Dobson circulation acts to warm the polar regions. Two- and three-dimensional modeling studies have shown how this circulation carries ozone from the photochemical production region in the tropical upper stratosphere to the polar lower stratosphere, resulting in the steady accumulation of ozone in the Northern Hemisphere polar lower stratosphere [e.g., *Chipperfield and Jones,* 1999].

Early investigators saw large increases of total ozone and temperature following major stratospheric warmings [e.g., *London,* 1963]. Our theoretical understanding of the relationship of waves to the stratospheric mass circulation and temperature caught up with observations when it was recognized that the transport circulation [*Plumb,* 1979] was approximated by the residual circulation [*Dunkerton,* 1978]. The relationship of the stratospheric circulations to the wave driving from the troposphere was solidified by *Haynes et al.* [1991] with the development of the "downward control" principle, which simply states that the circulation across an isentropic surface is controlled by upward Rossby and gravity wave propagation which break at higher levels. *Fusco and Salby* [1999] have used this concept to show that the interannual variation of northern hemisphere midlatitude total ozone is coherent with upwelling planetary wave activity.

The discovery of the Antarctic ozone hole by *Farman et al.* [1985] and the recognition of heterogeneous chemical processes as a principal cause of this ozone loss highlighted the important role of polar temperatures.

Low temperatures (< 195 K) result in the formation of polar stratospheric clouds (PSC), which lead to a release of chlorine from reservoir species into reactive forms. The rising of the sun in the polar spring period combined with these low temperatures results in large ozone losses via the ClO-ClO and BrO-ClO catalytic cycles [*Anderson et al.,* 1989].

Since the Arctic winter stratosphere is much warmer than the Antarctic winter stratosphere, PSCs are much less prevalent in the Arctic [*Poole and Pitts,* 1994]. This results in smaller Arctic ozone losses than over Antarctica. However, during the mid-1990s, spring Arctic ozone levels have been extremely low [*Newman et al.*, 1997; *Fioletov et al.,* 1997]. These low levels are due to both transport and chemical effects [*Chipperfield and Jones,* 1999]. *Coy et al.* [1997] showed that the extremely low ozone values of 1997 were accompanied by very low temperatures and extremely weak planetary wave driving of the stratosphere. *Zurek et al.* [1996] have shown that the mid-1990s (Upper Atmosphere Research Satellite period) is characterized by reductions in the frequency of stratospheric warmings. *Pawson and Naujokat* [1997, 1999] have shown that there has been an increase in the areal extent of these low temperatures, especially in late winter. *Waugh et al.* [1999] showed that the Arctic vortex had much later breakup dates in the 1990s as compared with earlier decades. These studies suggest that the 1990s were characterized by a climatic period of weaker planetary wave driving and lower temperatures.

In this paper we will quantitatively demonstrate the connection between the polar lower-stratospheric spring temperature in early March and the winter planetary wave driving. We will derive a simple theoretical relationship between the wave driving and the temperature, discuss the data used, and will demonstrate the quantitative relationship between the wave driving and the temperature. We will present the correlation of a variety of parameters to show the full impact of planetary wave driving of the stratosphere.

2. Theoretical formulation

The wave driving of the stratosphere is necessary for understanding the evolution of the stratospheric temperatures, circulation, and the transport of trace gases such as ozone. The transformed Eulerian mean (TEM) circulation provides a wave-mean flow theoretical framework for the overall dynamics of the stratosphere [see *Andrews et al.,* 1987, for a basic description of the TEM]. This TEM formulation can be used to directly connect wave effects with the circulation and the temperature of the polar lower stratosphere.

The zonal-mean thermodynamic equation can be simply expressed in the TEM formulation as

$$\frac{\partial \overline{T}}{\partial t} + \overline{w}^* S = \overline{Q}, \tag{1}$$

where $\overline{w}^*$ is the vertical component of the residual circulation, T is the temperature, Q is the heating, t is time, the overbar refers to a zonal-mean quantity, and S is the static stability ($= \partial T/\partial z + g/c_p$, where z is the pressure altitude, g is the gravitational constant, and c_p is the specific heat at constant pressure).

The vertical residual circulation component is expressed in Eulerian terms as

$$\overline{w}^* = \overline{w} + \frac{1}{a \cos\phi} \frac{\partial}{\partial \phi} \left(\frac{\cos\phi \, \overline{v'T'}}{S} \right), \tag{2}$$

where w is the vertical component of the wind, a is the Earth's radius, ϕ is latitude, and the prime indicates departures from the zonal mean. The meridional eddy heat flux is given by $\overline{v'T'}$. In this formulation of $\overline{w}^*$, we have neglected the vertical heat flux ($\overline{w'T'}$), since it is typically quite small in the stratosphere, although it can be quite important for the mesosphere via the heat flux associated with gravity waves.

We approximate the heating as $\overline{Q} = -\alpha(\overline{T} - \overline{T}_{re})$, a standard Newtonian cooling approximation [*Fels,* 1982], where $\overline{T}_{re}$ is the radiative equilibrium temperature and α is the damping rate. *Newman and Rosenfield* [1997] used temperature observations and a diabatic heating model to show that this linear relationship is extremely robust, and that α^{-1} is approximately 30 days in the lower stratosphere. Because the radiative equilibrium varies slowly with time, we determine the steady-state ($\partial \overline{T}/\partial t = 0$) solution of (1) yielding

$$\overline{T}_s = \overline{T}_{re} - \frac{\overline{w}^*_s S}{\alpha}. \tag{3}$$

where the subscript s denotes the steady-state solution value. This steady-state solution shows that the temperature of the lower stratosphere is controlled by the radiative equilibrium temperature and the strength of the residual circulation. A stronger downward circulation produces warmer temperatures, while a weaker circulation leads to colder temperatures closer to the radiative equilibrium. Further, weak thermal damping amplifies the effect of the circulation. While this steady-state solution allows us to examine the year-to-year variation of winter temperatures, it does not shed much light on those factors that impact the polar lower stratospheric temperature during the chemically critical period of late winter.

We can temporally integrate (1) to gain insight into the time-dependent solution. We assume that both the radiative equilibrium temperature and the radiative damping rate are both constant over the periods that we are considering (mid-winter), as will be shown in section 5. We integrate (1) from a time prior to our period of interest ($-t$, early winter, for example) to the day of interest ($t = 0$, approximately early March).

$$\overline{T}(0) = \overline{T}_{re} + e^{-\alpha t}[\overline{T}(-t) - \overline{T}_{re}] - \int_{-t}^{0} e^{\alpha \hat{t}} \overline{w}^* S d\hat{t}. \tag{4}$$

The first term on the right hand side is the radiative equilibrium temperature. The second term on the right hand side represents the initial deviation from the radiative temperature. The third term is the weighted integral of the vertical residual circulation. For a typical damping time scale of 30 days, a temperature deviation or a strong $\overline{w}^*$ occuring in the November–December period will have limited impact on the March polar temperature.

The steady-state solution to (1) is retrieved from (4) for constant $\overline{w}^*$ and $\alpha t >> 1$. However, the interannual variability of both March polar temperatures and $\overline{w}^*$, combined with the relatively short damping times ($\alpha^{-1} \approx 30$ days) with respect to the annual cycle, suggests that the steady-state assumption (3) is not valid. Hence, we must use the time-dependent solution (4) for interpreting these temperatures.

Understanding how $\overline{w}^*$ is controlled by planetary waves is the key to connecting the wave driving to polar temperatures. *Haynes et al.* [1991] showed how the extratropical diabatic mass flow across a given isentropic surface may be regarded as being controlled exclusively by the Eliassen-Palm (E-P) flux divergence distribution above that surface under the assumption that $\partial\overline{u}/\partial t \approx 0$. In a practical sense, the assumption that $\partial\overline{u}/\partial t$ is small is reasonable in the stratosphere for monthly averages. Further, *Haynes et al.* [1991] also show that the downward control principle is generally applicable for longer-period, larger-scale waves in a time-dependent solution. Hence, the dissipation of vertically-propagating planetary waves is a principal forcing of the residual circulation in the polar lower stratosphere.

This control of the residual circulation is mathematically shown via the steady-state zonal-mean momentum equation in the TEM formulation. Following *Haynes et al.* [1991],

$$\overline{w}^* = -\frac{1}{a\rho_0 \cos\phi} \frac{\partial}{\partial \phi} \left[\int_{z}^{\infty} \left(\frac{\nabla \cdot \mathbf{F}}{af} \right) d\hat{z} \right], \tag{5}$$

where $\nabla \cdot \mathbf{F}$ is the E-P flux divergence, f is the Coriolis parameter, and ρ_0 is the density ($\equiv \rho_s e^{-z/H}$, where H is a mean scale height and ρ_s is the density at 1000 hPa). The E-P flux vector ($\mathbf{F}$) has components $a\rho_0 \cos\phi\,(-\overline{u'v'}$, $f\,\overline{v'T'}/S)$, and

$$\begin{aligned}\nabla \cdot \mathbf{F} &= -\frac{1}{\cos\phi}\frac{\partial}{\partial\phi}(\rho_0 \cos^2\phi\,\overline{u'v'}) \\ &\quad + af\cos\phi\,\frac{\partial}{\partial z}\left(\frac{\rho_0\overline{v'T'}}{S}\right). \qquad (6)\end{aligned}$$

Integration of (5) from a midlatitude reference latitude ϕ_r to the pole yields the polar-cap area-weighted average of $\overline{w}^*$ as

$$\langle\overline{w}^*\rangle_{\phi_r} = \frac{1}{a\rho_0(1-\sin\phi_r)}\int_z^\infty \left\{\frac{\nabla\cdot\mathbf{F}}{af}\right\}_{\phi_r} d\hat{z}, \qquad (7)$$

where $\langle\rangle_{\phi_r}$ is the average of the quantity in angle brackets between ϕ_r and the pole (a polar-cap area-weighted average, not the global average) and $\{\}_{\phi_r}$ is the quantity in brackets evaluated at ϕ_r. Substituting (6) into (7), choosing a value of ϕ_r where the first term is small, and integrating we get

$$\langle\overline{w}^*\rangle_{\phi_r} = -\frac{\cos\phi_r}{a(1-\sin\phi_r)}\left\{\frac{\overline{v'T'}}{S}\right\}_{\phi_r}. \qquad (8)$$

This equation shows that $\overline{w}^*$ should be anticorrelated with the eddy heat flux. In this formulation, the assumptions that the first term on the RHS of (6) and $\partial\overline{u}/\partial t$ are both small will be justified in section 3.

Because we assume a steady-state condition, (8) should not be strictly applied in (4). However, we practically find that $\overline{w}^*$is well correlated with the eddy heat flux, as *Haynes et al.* [1991] found and as will be justified in section 3. Hence, we substitute (8) into (4) to get

$$\begin{aligned}\langle\overline{T}(0)\rangle_{\phi_r} &= \langle\overline{T}_{re}\rangle_{\phi_r} + e^{-\alpha t}[\langle\overline{T}(-t)\rangle_{\phi_r} - \langle\overline{T}_{re}\rangle_{\phi_r}] \\ &\quad + \frac{\cos\phi_r}{a(1-\sin\phi_r)}\int_{-t}^0 e^{\alpha\hat{t}}\,\{\overline{v'T'}\}_{\phi_r} d\hat{t}. \qquad (9)\end{aligned}$$

Therefore, the polar mean temperature is directly correlated with the exponentially-weighted integral of the preceding eddy heat flux.

Another important aspect of the eddy heat flux is that it is proportional to the vertical group velocity of a planetary wave [*Edmon et al.,* 1980]. The eddy heat flux for a particular wave is proportional to the square of the wave amplitude, and the vertical and zonal wavenumbers. Since the eddy heat flux is almost always positive [*Newman and Nash,* 2000], the wave energy is always propagating vertically from the troposphere to the stratosphere. Hence, we interchangeably refer to the eddy heat flux as the wave driving of the stratosphere.

The TEM formulation of the thermodynamic equation has been used to show the connection between the circulation and the temperature of the lower stratosphere. We find that the steady-state assumption for the lower-stratospheric temperature is generally not valid for the early March period because of the dependence of these March polar temperatures on the highly variable January–February residual circulation. We also find that it is both the direct effect of the diabatic heating, and the weighted impact of the residual circulation that controls the polar lower-stratospheric temperature. Further, the temperature anomaly is related to the timing and strength of the residual circulation, and the wave driving of the stratosphere controls the strength of that residual circulation. We theoretically find from (8) that the eddy heat flux (vertical component of the wave driving) is anticorrelated with the polar vertical residual circulation.

3. Data

We use two meteorological analyses in this paper. These are the National Centers for Environmental Prediction's (NCEP) Climate Prediction Center (CPC) stratospheric analysis and the jointly produced NCEP/NCAR reanalysis. The NCEP/CPC analysis is an operational product which is subject to data changes. In contrast, the NCEP/NCAR reanalysis is a continuous, consistent assimilation system with analyses extending back to 1958. Herein, we use data beginning in 1979 because of the problems associated with analyses prior to incorporation of satellite data [*Pawson and Fiorino,* 1998; *Santer et al.,* 1999].

The NCEP/NCAR reanalyses use the NCEP spectral model with a T62 triangular truncation and satellite temperature retrievals. This analysis system is consistently run for all of the years, such that changes in the archived meteorological fields are either due to changes in observational systems or due to real geophysical processes [*Kalnay et al.,* 1996]. The data are available on pressure levels from 1000 to 10 hPa, and a 2.5° longitude by 2.5° latitude horizontal grid.

The NCEP/CPC produces analyses based on the NCEP T126 GDAS model from 1000 to 100 hPa, and an objective analysis of satellite and radiosonde data for pressure levels from 70 to 0.4 hPa [*Gelman et* *al.,* 1986; *Nagatani et al.,* 1988; *Finger et al.,* 1993]. The horizontal grid is a 65 $\times$ 65 point stereographic projection which we interpolate to a grid of 5° longitude by 2° latitude. At 100 hPa and below, these data are solely based on the T126 GDAS product. Temperatures and geopotential heights are basic products of this tropospheric-stratospheric hybrid system, while we produce consistent balanced winds at all levels using the method described in *Randel* [1987] and *Newman et al.* [1988].

Newman and Nash [2000] showed that there are 30% peak-to-peak differences between eddy heat fluxes derived from the variety of analyses produced by the various meteorological centers. The large differences result from the differences in stationary planetary waves in the temperature and meridional wind fields. In contrast, planetary-scale transient waves showed excellent agreement amongst the five analyzed data sets used in *Newman and Nash* [2000].

Good NCEP/NCAR data quality is necessary to test our three assumptions from the previous section: $\partial\overline{u}/\partial t \approx 0$, $\overline{u'v'} \approx 0$ at 60°N, and (8) is valid on a day-to-day basis. First, we've tested the zonal-mean zonal wind momentum balance which is given by

$$\frac{\partial\overline{u}}{\partial t} = f\overline{v}^{*} + \frac{\nabla\cdot\mathbf{F}}{a\rho_0\cos\phi}. \tag{10}$$

For example, a 22-year average of the 30 hPa 50–70°N region for the January 15–February 28 period yields values of 0.2, -2.1, and 2.0 m s^{-1} day^{-1}, respectively. Individual years show that $\partial\overline{u}/\partial t$ is small relative to the two forcing terms. While the 22-year average of the two terms on the RHS of (10) shows a nice balance, the individual years rarely balance. This poor balance is a reflection of the large uncertainties associated with calculating these highly derived quantities. These results are seen throughout the stratosphere. Second, we assume that the mass-weighted eddy momentum flux is zero, which is true on average at a latitude of 67°N. Third, we assert in section 2 that $\overline{w}^{*}$is proportional to $-\overline{v'T'}$ on a day-to-day basis. Since the vertical wind is only provided in terms of the time derivative of the pressure ($\overline{\omega} \equiv \partial\overline{p}/\partial t$, where p is the pressure) and is only available to 100 hPa, we have used the smoothed zonal-mean meridional wind with the continuity equation to calculate the zonal-mean vertical wind, and then calculated the vertical residual circulation from this mean vertical wind. Using 22 years of daily January values at 50 hPa, we find that the polar-cap average of the vertical residual circulation has a correlation of 0.92 with the 60°N heat flux.

4. The Eddy Heat Flux-Polar Temperature Connection

Using the theoretical considerations of section 2, the polar lower-stratospheric temperature is dependent on the time-integrated effects of the eddy heat flux. We test this by averaging the NCEP/NCAR reanalysis total eddy heat flux at 100 hPa between 45°N and 75°N. The time scale of the integration in this term is near 30 days, as determined by the thermal damping rate (α^{-1}) of *Newman and Rosenfield* [1997]. As a simple test, Figure 1a shows this total eddy heat flux over the period January 15 to February 28 versus the 50 hPa March 1–15 polar-averaged temperature (60°–90°N) for each year from 1979 to 2000. The very strong correlation ($r = 0.82$) between March temperatures and these late winter eddy heat fluxes, means that we can account for 15.3 K of the 18.6 K peak-to-peak interannual variation of the polar lower-stratospheric temperature. Using NCEP/CPC data yields a correlation of 0.85, showing that that magnitude of the correlation is not dependent on the data set. The correlation between temperature and eddy heat flux for the simultaneous period (Figure 1b) is much poorer ($r = 0.25$). Figure 1

Assuming that the 22 years are independent and that the residual errors are normally distributed, the 0.82 correlation is significant at a confidence level greater than 99.9%. If we assume that there are possible errors in the calculated eddy heat flux of up to 30% peak-to-peak [*Newman and Nash,* 2000] we can determine the impact on the correlation with a Monte Carlo simulation. We added normally distributed "noise" to the calculated eddy heat flux with a scaling of 15% of the standard deviation of this eddy heat flux and performed 10,000 trials. We found a mean correlation value of 0.81 with 90% confidence limits of 0.78 and 0.84 and 99% confidence limits of 0.76 and 0.86. Since our 0.82 correlation is not statistically different from the Monte Carlo mean, we conclude that eddy heat flux errors do not affect the eddy heat flux-temperature correlation.

From (9), we should correlate the polar-cap temperature with an exponentially-weighted eddy heat flux. However, we use simple cosine-weighted and time averages of the eddy heat flux. A simple time average approximates an exponential weighting, if the averaging period is equal to the damping time. The 100 hPa eddy heat flux is used since it is highly correlated with the 50 hPa values ($r = 0.95$), and represents the wave energy propagating into the stratosphere. Finally, the 60°N eddy heat flux is highly correlated with the 45°–75°N cosine-weighted average, but averaging reduces noise estimated from the uncertainty in the eddy heat flux.

The exponentially-weighted total eddy heat flux at 60°N and 50 hPa has a correlation of 0.85 with the 50 hPa polar-cap temperature for March 1.

The relationship of the midlatitude lower-stratospheric eddy heat flux to the entire temperature field has been determined using a 1-point correlation map. Instead of an average polar temperature, the same 22-year eddy heat flux time series used in Figure 1 is correlated with each latitude and pressure in the March 1–15 average zonal-mean temperature fields for the 22 years of 1979 to 2000. This eddy heat flux correlation with the temperature field is shown in Plate 1a. Large positive correlations cover the entire region from about 60°N to the pole, and from the tropopause up to 20 hPa. The correlation has a maximum value greater than 0.8. We also note the strong negative correlation near the tropical tropopause ($r < -0.5$). Correlations of 0.6 and 0.5 are significant at the 99% and 98.3% confidence levels, respectively. Again, the CPC data show a similar pattern to the NCEP/NCAR reanalyses. Plate 1

The strong correlations exhibited in Figure 1a and Plate 1a are in good agreement with our theoretical considerations. We expect stronger wave driving of the stratosphere to enhance the residual (i.e., Brewer-Dobson) circulation. The enhanced sinking motion in the polar region will increase the temperature, while the concomitant rising motion in the tropics will lead to lower tropical temperatures.

The eddy heat flux time series and its interannual variation is dominated by planetary-scale waves. We can partition the total eddy heat flux variations shown in Figure 1 and used in Plate 1a into the flux due to zonal harmonic waves 1–3 and the flux due to waves 4 and higher. Plate 1b displays the correlation of the eddy heat flux for waves 1–3 with the March 1–15 temperature field. A comparison of Plates 1a and 1b shows that restricting the eddy heat flux to only the planetary-scale waves leads to a slight improvement of the correlation. A weak negative correlation is found with waves 4 and higher (not shown).

We can reverse our correlations and examine the relationship between the lower-stratospheric polar temperature and the eddy heat flux patterns. We use the 22-year time series of 50 hPa temperature averaged over 60°–90°N and March 1–15 (y-axis of Figure 1) with the 22 years of zonal-mean eddy heat flux fields. This polar temperature correlation with the eddy heat flux (averaged over January 15 to February 28) is shown for the total eddy heat flux in Plate 2a and for waves 1–3 in Plate 2b. The maximum correlation of the time series is found Plate 2 at about 65°N and 100 hPa, with significant correlation ($r > 0.6$) extending down to 400 hPa. The correlations are rather narrow, but coherent over deep layers. As with the correlation of the waves 1–3 eddy heat flux with the zonal-mean temperature field in Plate 1b, the correlation of the 60°–90°N temperature with the zonal-mean waves 1–3 eddy heat flux is also slightly improved over the total eddy heat flux.

The maximum correlation between the eddy heat flux and the 50 hPa temperature time series averaged over March 1–15 and between 60°–90°N is found close to the peak of the eddy heat flux. The white lines superimposed on Plate 2b show the magnitude of the mean eddy heat flux. At 100 hPa, the eddy heat flux maximum is found at about 60°N, while the temperature-eddy heat flux correlation has it's highest value at about 65°N. Generally, the peak of the correlation is associated with the peak of the eddy heat flux, however the correlation peak leans more to the south with height. The slight negative correlation at 100 hPa and 35°N occurs in a region where there are weak mean eddy heat flux values.

The 45-day period, January 15 to February 28, is arbitrarily chosen using a radiative damping time scale near 30 days. However, the large correlations are fairly insensitive to both the averaging period and latitude range of the eddy heat flux. For the 100 hPa waves 1–3 eddy heat flux averaged over 45°–75°N, the correlation with the 50 hPa temperature averaged over March 1–15 and 60°–90°N is largest for the averaging period of January 22–March 7 ($r = 0.92$). The region of maximum correlation can be slightly improved by also including a wider latitudinal averaging band. Averaging from 5°–75°N for the January 22–March 7 period further improves the correlation to $r = 0.94$. All of the correlations are extremely robust if the late January and February periods are included with latitudinal bounds that at least cover the region of the peak eddy heat flux (50°–75°N). In addition, excellent correlations are also found using exponential damping rates consistent with (9) ranging from 30-100 days.

As was shown in (9), the correlation between the eddy heat flux and the polar-cap temperature is a function of the damping rate. In the middle and upper stratosphere, damping rates are quite small [*Kiehl and Solomon*, 1986; *Shine,* 1987; *Newman and Rosenfield,* 1997; *Mlynczak et al.,* 1999], hence, the phase lead of the eddy heat flux at 10 hPa that produces a good correlation with the 10 hPa polar-cap temperature ought to be quite short. Using a 15-day damping time scale with the 10 hPa eddy heat flux yields a correlation of 0.95 with the 10 hPa polar-cap temperature. Similarly, at lower altitudes, the phase lead of the eddy heat flux is longer. For example, the 100 hPa eddy heat flux is correlated with the 100 hPa polar-cap temperature at 0.75 for a damping time scale of 90 days.

5. Annual cycle of the eddy heat flux-temperature relationship and a comparison to the radiatively estimated temperature

The eddy heat flux relationship to temperature is very robust over the entire course of the winter. We again use the 100 hPa waves 1–3 eddy heat flux averaged for 45°–75° N and for the 45-day time period prior to the 50 hPa temperature field time period. Plate 3 shows the correlations for early December and late January. The patterns are very similar to the early March pattern. We have also used a 55-day exponential weighting consistent with the integration shown in the third term of (9) to calculate the linear relation between the eddy heat flux and the temperature at 80°–85°N for each day over the entire year. In this case, the correlation is greater than 0.7 for the entire period from early December to early April. Plate 3

The excellent correlation between the 100 hPa eddy heat flux and the 50 hPa polar temperature allows us to estimate the polar temperature in the absence of dynamics (i.e., when $\overline{v'T'} = 0$). We make this estimate by least-squares fitting a linear relationship of the 100 hPa eddy heat flux to the 50 hPa polar temperature, and then calculating the intercept of this line. We can fit a straight line to Figure 1a that has a intercept of 199.8 K and a slope of 0.92 K m^{-1} s. The temperature of 199.8 K would be the March 1–15 temperature of the polar region in the absence of any dynamics (i.e., if the January 15 to February 28 eddy heat flux at 100 hPa was equal to zero).

We can use this concept to calculate the polar temperature in the absence of dynamics for each day of the entire year. Figure 2 shows the mean annual cycle of the temperature (thick line) averaged over the 22-year period 1979–2000 in the latitude band 80°–85°N. The white line is the temperature in the absence of dynamics, calculated by linearly fitting the eddy heat flux to temperature relationship for each day, and then plotting the intercept of this linear fit. The gray shading shows the 95% confidence interval as determined from a two-sided Student's t-distribution. While the eddy heat flux relationship to temperature is quite poor in summer, the small values of both the eddy heat flux and temperature lead to small uncertainties in the estimated temperature Figure 2 intercept. During mid-winter the “dynamics free” temperature is much colder than the mean temperature.

The “dynamics free” temperature shown in Figure 2 can be compared to a radiation model estimate of the temperature in the absence of dynamics. The radiative temperatures were calculated by time-integrating the radiative term forward with a seasonal cycle and with $\overline{w}^*$ set to zero in (1), using the GSFC radiative model [*Rosenfield et al.,* 1994]. Because radiatively-determined lower-stratospheric temperatures are sensitive to temperatures below the tropopause, tropospheric temperatures were specified using a time-varying climatology from the surface to a specified level. Two cases were considered using specified tropopause pressures of 200 and 100 hPa. The radiatively determined temperatures for the two cases at 80°N are also shown in Figure 2 as the upper (100 hPa) and lower (200 hPa) dashed lines. The “radiative equilibrium” is never reached in this model integration because of the thermal inertia and the radiative time scale. Agreement between the radiatively-determined temperature (dashed lines) and the “dynamics free” temperature estimate is reasonably good in mid-winter. Large temperature discrepancies are apparent in the spring, summer, and fall. Based upon a 30- to 60-day relaxation time scale, a 10 K temperature discrepancy would be accounted for by an additional 0.15–0.3 K day^{-1} heating. Such a small error is generally within the uncertainty of the heating calculation. For example, if nitrogen dioxide (NO_2) shortwave heating was included, it might account for part of this discrepancy because of the high NO_2 concentrations and continuous sunlight of polar summer [*Kiehl and Solomon*, 1986]. However, the inclusion of longwave cooling by nitrous oxide (N_2O) and methane (CH_4) would lead to colder radiative temperatures.

6. Connection of the eddy heat flux to the eddy momentum and the E-P flux divergence

An increase of the wave driving of the stratosphere is correlated with an increase in the stratospheric eddy momentum flux and a consequent increase in the E-P flux divergence. We test this by correlating our 22-year eddy heat flux time series with both the momentum and E-P flux divergence fields. We use the same periods for the correlation, since the wave propagation time scale of a few days is short compared to our radiative damping time scale of 1–2 months. Plate 4 displays the correlation of the eddy heat flux with the eddy momentum flux (Plate 4a) and the E-P flux divergence (Plate 4b). Planetary waves tend to propagate upward and equatorward Plate 4 towards the higher index of refraction region in the tropics [*Edmon et al.,* 1980]. The 100 hPa eddy heat flux correlation with the eddy momentum flux is consistent with the maximum correlation at slightly higher altitude (30 hPa) and slightly equatorward (45°N) as a result of this wave propagation.

As the waves propagate upward, they deposit momentum and decelerate the mean flow [*Edmon et al.,* 1980]. The E-P flux divergence is the wave forcing term in the zonal-mean momentum equation, where a strong negative E-P flux divergence acts to decelerate the zonal-mean flow. This wave momentum deposition results in the high degree of correlation of the eddy heat flux with the E-P flux divergence (Plate 4b). The E-P flux divergence is well correlated over an extensive region centered on about the 30 hPa layer.

The E-P flux divergence decelerates the polar night jet stream. From the momentum equation, the strong deceleration via the E-P flux divergence will produce both a weaker jet and warmer polar temperatures. Again, we expect that the E-P flux divergence will have a time-lagged impact on the jet. We test this by correlating the 22-year E-P flux divergence time series for the January 15–February 28 period with the zonal-mean wind and temperature fields during March 1–15. The correlation with the wind field (Plate 5a) is excellent, with the correlation extending over a tremendous vertical depth. Strong wave-forced decelerations at the 30 hPa level via the E-P flux divergence (i.e., large negative values of the E-P flux divergence) lead to a weakened jet stream. This weakening of the flow field is coherent from the middle stratosphere into the troposphere in the 50°–60°N region.

Plate 5

The correlation of the E-P flux divergence time series with the temperature fields (Plate 5b) is also excellent, with a strong negative correlation poleward and below the E-P flux divergence average region. This correlation pattern results from the residual circulation. From the TEM momentum equation, a strong E-P flux divergence-induced deceleration is balanced by a poleward meridional residual circulation flow. From our continuity equation, this poleward flow produces a downward flow near the pole, which will act to raise the polar temperature slightly below the E-P flux divergence level. Hence, the 50 hPa polar temperature is strongly anti-correlated with the mid-latitude 30 hPa E-P flux divergence.

7. Summary

The interannual variability of March polar stratospheric temperatures is principally related to the tropospheric to lower-stratospheric eddy heat flux observed in January–February. This correlation is caused by the planetary-scale waves. A strong planetary-wave eddy heat flux in the 100–400 hPa and 45°–75°N region during January–February results in a warm March polar lower stratosphere, while a weak planetary-wave eddy heat flux results in a cold March polar lower stratosphere. The eddy heat flux is also consistently correlated with the stratospheric momentum flux, E-P flux divergence, and zonal-mean wind. These results are both statistically significant, and are extremely robust to varying averaging periods and latitude ranges. Given the uncertainties in the eddy heat flux estimates, the winter polar lower-stratospheric temperature is almost completely determined by the January–February 100 hPa eddy heat flux.

This January–February eddy heat flux correlation with the March lower stratospheric temperature is consistent with a theoretical framework based on linear wave propagation. This process is schematically shown in Figure 3. The eddy momentum and heat fluxes are directly proportional to the wave activity times the horizontal and vertical group velocities, respectively [*Edmon et al.,* 1980]. Large eddy heat fluxes represent an upward propagation of planetary wave activity into the stratosphere (Figure 3, point 1). In the stratosphere, these waves are refracted towards the equator and deposit easterly momentum (Figure 3, point 2) [*Karoly et al.,* 1982]. The equatorward wave refraction is shown in Plate 4a by the strong simultaneous correlation of the eddy heat flux with the eddy momentum flux above and equatorward of the eddy heat flux region. The deposition of easterly momentum is shown in Plate 4b by the strong negative correlation over a fairly broad region. This momentum deposition is balanced by a northward residual circulation (Figure 3, point 3) which acts to decelerate the jet stream. Plate 5a illustrates this deceleration with the correlation of the E-P flux divergence with the wind. A strong wave-induced deceleration results in a weak March jet. This wave-induced meridional residual circulation causes rising motion in the tropics, and sinking motion in the polar region (Figure 3, point 4). The polar sinking motion warms the stratosphere. Plate 5b illustrates this warming with the correlation of the E-P flux divergence with the temperature. The warm temperatures are a result of the wave events, and are not observed simultaneously with the wave event, but *after* the wave event. In the *absence* of mid-winter waves propagating

Figure 3

into the stratosphere, the wave forced jet deceleration is weak, resulting in a stronger March polar night jet and a colder March polar lower stratosphere.

8. Conclusions

The temperature of the polar lower stratosphere during March is the key factor in understanding polar ozone losses. Very cold years such as 1996 and 1997 have led to large chlorine-catalyzed ozone losses, while warmer years have very little ozone loss. The temperature of the early March polar lower stratosphere is principally driven by the mid-winter strength and duration of planetary waves propagating into the stratosphere. The cold polar stratospheric March periods during 1997 and 2000 were directly a result of the weak January–February wave driving of the stratosphere, while the warm polar stratospheric March periods during 1998 and 1999 were driven by the strong January–February wave driving.

The fundamental driver for the polar lower-stratospheric temperature is the basic radiative state. This involves the solar angle via polar night and continuous summer daylight, and the concentrations of basic radiative gases such as carbon dioxide, water, and ozone [*Shine,* 1987]. However, the winter temperature of the polar lower stratosphere is much warmer than would be expected from purely radiative considerations [*Fels,* 1982]. This observational-based study quantitatively shows that the difference between the calculated pure radiatively-driven polar temperature and actual observations is due to planetary-scale waves driving the polar temperature away from radiative equilibrium.

Reductions of planetary-scale waves propagating into the stratosphere will have a triple impact on ozone levels. First, reduced wave driving will lead to cooler polar temperatures. Second, a reduction will weaken the transport of ozone into the polar region. Finally, a reduction may lead to a warmer tropical tropopause, thereby enhancing stratospheric water levels.

Relatively modest changes of the tropospheric planetary waves can seriously impact the polar lower-stratospheric temperatures, and hence ozone loss. A 10% reduction of the 100 hPa eddy heat flux would result in a 1.7 K temperature decrease. Such sensitivity is a serious test for general circulation models (GCM) that are coupled to stratospheric chemistry models. In fact, these results provide a key test for GCMs which have predicted weakened wave driving in future climates (e.g., *Shindell et al.,* 1998). These GCMs must accurately capture both the timing between the tropospheric wave driving and the lower-stratospheric temperature, and the quantitative sensitivity of this wave driving temperature relationship.

In addition to the impact of the wave driving on temperatures, the wave driving also impacts the advection of ozone via the residual circulation. Years with weak wave driving have weaker residual circulations, and therefore less poleward and downward transport of ozone. A 10% reduction of the 100 hPa eddy heat flux would result in a 10% weakening of the circulation. The effect of the residual circulation interannual variability on polar ozone levels has been shown by *Chipperfield and Jones* [1999]. This interannual variability of the transport is directly related to the stratospheric wave driving.

The wave driving of stratospheric temperatures may also impact the water vapor concentrations in the stratosphere. As was shown in Plate 1, the wave driving is anti-correlated with the mean tropical tropopause temperature. A weakening of the wave driving may lead to warmer tropical tropopause temperatures. One of the possible factors determining the water vapor concentration is the "freeze drying" of air entering the stratosphere at the tropical tropopause [see *Dessler,* 1998, and references therein]. Predictions of weakened wave driving of the stratosphere by *Shindell et al.* [1998] in a greenhouse climate may warm the tropical tropopause region, consequently increasing the water vapor concentration in the stratosphere. Since polar stratospheric cloud formation is dependent on the water vapor concentration, a stratospheric water vapor increase will increase the occurrence of PSCs in the polar stratosphere. This increased occurrence of PSCs will permit earlier and more extensive activation of chlorine, with consequent greater ozone loss [*Kirk-Davidoff et al.,* 1999]. Hence, from our observations we infer that a predicted reduction of the stratospheric wave driving may increase water in the stratosphere with a consequent loss of polar ozone.

Numerous studies have tied known atmospheric variations such as the Arctic oscillation (AO) and the quasi-biennial oscillation (QBO) to the Arctic stratosphere. Both the statistical and causal relationship of the eddy heat flux to the polar lower-stratospheric temperatures provides a new framework for testing these known variations. Since the eddy heat flux explains the majority of the stratospheric polar temperature variance, other variations must either directly effect the eddy heat flux and indirectly the polar temperature, or must be related to the residual of polar stratospheric temperatures after removing the eddy heat flux influence. Given the large impact of the wave driving on the stratospheric temperatures, future research efforts need to be concentrated on understanding the processes that control the tropospheric wave driving.

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Figure 1. Temperature at 50 hPa, averaged for 60°–90°N and March 1–15, plotted against the total eddy heat flux at 100 hPa, averaged for 45°–75°N) and for (a) January 15 to February 28 and (b) March 1–15. The individual years are as indicated.

Figure 2. Temperature at 50 hPa, averaged for 80°–85°N and for the 22-year period 1979–2000 (thick solid line). The dashed lines show the radiatively calculated temperature using ozone, water, and carbon dioxide for lower boundaries at 100 hPa (upper dashed line) and 200 hPa (lower dashed line). The white line shows the estimated "dynamics free" temperature using the heat flux relationship to temperature. The gray shading shows the 95% confidence limits on the "dynamics free" temperature estimates.

Figure 3. Schematic illustration of 1) planetary waves propagating into the stratosphere, 2) slowly refracting towards the equator, 3) depositing easterly momentum, and 4) inducing a residual circulation that causes uplift in the tropics and sinking in the polar region. The short arrows illustrate the wave propagation while the thick line with arrows shows the residual circulation. The thin solid lines show the wind speed, the dotted line shows the tropopause, and the dashed lines show the E-P flux divergence or wave-driven wind deceleration.

Plate 1. Correlation of the 22-year time series (1979–2000) of the (a) total and (b) waves 1–3 components of the eddy heat flux at 100 hPa, averaged for 45°–75°N and for January 15 to February 28, with the temperature fields, averaged for March 1–15. Superimposed is the tropopause (thick purple line) as determined from the 22-year March 1–15 averaged Brunt-Vaisalla frequency. Also shown as the white lines are the 22-year January 15 to February 28 averaged zonal-mean zonal wind on (a) and waves 1–3 components of the eddy heat flux on (b).

Plate 2. As in Plate 1 but for the correlation of the temperature at 50 hPa, averaged for 60°–90°N and March 1–15, with the (a) total and (b) waves 1–3 components of the eddy heat flux fields averaged for January 15 to February 28.

Plate 3. Correlation of the 22-year time series (1978/1979–1999/2000) of the waves 1–3 components of the eddy heat flux at 100 hPa with the temperature fields. The heat flux is averaged for 45°–75°N and (a) October 17 to November 30 and (b) December 2 to January 15. Temperatures are averaged for (a) December 1–15 and (b) January 16–31. Superimposed is the tropopause (thick purple line) as determined from the 22-year (a) December 1–15 and (b) January 16–31 averaged Brunt-Vaisalla frequency and the 22-year (a) October 17 to November 30 and (b) December 2 to January 15 averaged zonal-mean zonal wind (white lines).

Plate 4. Correlation of the 22-year time series (1979–2000) of the eddy heat flux at 100 hPa, averaged for 45°–75°N, with the (a) eddy momentum flux fields and (b) E-P flux divergence fields. All quantities are waves 1–3 components and averaged for January 15 to February 28. Superimposed is the tropopause (thick purple line) as determined from the averaged Brunt-Vaisalla frequency and the averaged zonal-mean zonal wind (white lines), both averaged for the January 15 to February 28 period.

Plate 5. Correlation of the 22-year time series (1979–2000) of the waves 1–3 components of E-P flux divergence at 30 hPa, averaged for 30°–75°N and January 15 to February 28, with the March 1–15 averaged (a) zonal-mean zonal wind and (b) temperature fields. Superimposed is the tropopause (thick purple line) as determined from the averaged Brunt-Vaisalla frequency and the averaged zonal-mean zonal wind (white lines), both averaged for the March 1–15 period.