S P A R C

Stratospheric Processes And their Role in Climate

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The Stratosphere in the Climate System

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Introduction

Until relatively recently, the stratosphere has been viewed as a passive part of the climate system, with a small mass and relatively minor influence on surface climate. The primary effect of the stratosphere on surface climate was felt to be the radiative effect of the amount and vertical distribution of stratospheric ozone, which was more of concern for its health effects, rather than its effect on surface temperature and precipitation. In recent years, however, the stratosphere has been shown to have a significant influence on surface weather and climate, and this has become a major research topic.

What are the reasons for the enhanced interest in the stratosphere as a player in climate change? First, at the present time changes in stratospheric temperature associated with ozone depletion are large compared to changes thought to be associated with other greenhouse gases (GHGs) (Ramaswamy *et al.*, 2001). Moreover, temperature changes expected in response to increasing GHGs are larger in the stratosphere than in the troposphere, albeit of opposite sign. So the magnitudes of the temperature changes in the stratosphere associated with human activities are much larger than those expected near the surface. The interactions between a warming troposphere and cooling stratosphere are potentially very significant.

Secondly, recent research has shown that changes in the stratosphere can have significant impacts on surface weather and climate (Kodera and Yamazaki, 1994). The most obvious example of this is the sudden stratospheric warming, which has been shown to have a robust and consistent effect on surface weather patterns (Baldwin and Dunkerton, 1999; Baldwin *et al*., 2003; Kodera *et al*., 2000). This connection is expressed primarily through the dominant natural structures of variability, which tend to be quasi-zonally symmetric and are often called the annular modes of variability (Thompson and Wallace, 1998; 2000). A similar pattern of events occurs in the Northern Hemisphere (NH) during a midwinter major warming event, and in the Southern Hemisphere (SH) during the Spring warming. Both stratospheric and tropospheric climate appear to have linked secular trends over the past 30 years (Thompson and Solomon, 2002; Thompson *et al*., 2000).

The effects of the annular mode variability appear to extend into the tropics, both through the effects on the Brewer-Dobson Circulation, that flows upward across the tropical tropopause and toward the winter pole (Holton *et al*., 1995), and through the connection of the stratosphere to tropospheric modes of variability. The tropospheric annular modes can influence the meridional location of eddy activity and, thereby, influence tropical circulation (Thompson and Lorenz, 2003). The tropical tropopause is a particularly interesting and critical area for stratospheric chemistry and possible climate-chemistry interactions (Mote *et al*., 1996). Here we focus principally on the extratropical dynamical interactions between the stratosphere and troposphere and the possible influences of global change on those interactions.

Stratospheric Warmings and Downward Control

Annular modes explain a large fraction of the intraseasonal and interannual variability (Kidson, 1988; Nigam, 1990; Thompson and Wallace, 2000) and appear to arise naturally as a result of internal interactions within the troposphere and stratosphere (Feldstein and Lee, 1998; Limpasuvan and Hartmann, 2000; Lorenz and Hartmann, 2001; Lorenz and Hartmann, 2003; Robinson, 1991; Yu and Hartmann, 1993). In the stratosphere most of the variability is associated with variations in the large-scale wave driving, which is strongly associated with the occurrence of stratospheric warmings. In the troposphere wave driving is also necessary to move the westerly jets in latitude, but this driving is provided by transient baroclinic waves as well as quasi-stationary planetary waves. Although stratospheric annular variability and tropospheric annular variability are coupled at times, tropospheric annular variations also occur independent of stratospheric annular variations (Kodera and

Kuroda, 2000). It appears that low-frequency quasi-barotropic waves are most important for producing changes in the polarity of annular modes of variation in the troposphere, but high-frequency baroclinic waves are most important for maintaining the persistence of these anomalies (Feldstein and Lee, 1998; Lorenz and Hartmann, 2001; Lorenz and Hartmann, 2003).

Limpsuvan *et al.* (2003) have taken the 44 years NH data from the NCEP/NCAR reanalysis and composited the flow relative to stratospheric warming events. The stratospheric warming events are selected as anomalies of the first mode of 50 hPa zonal flow – times when the zonal vortex is especially weak. These dates correspond approximately to major or minor stratospheric warmings. These composites show statistically significant forerunner and follower structures to the NH wintertime warming events.

Figure 1 shows the composite geopotential anomaly height fields at 50, 250 and 1000 hPa at five temporal phases relative to the warming time. The lifecycle of the stratospheric warming was divided into five 15-day periods: *onset* (days -37 to –23), *growth* (days -22 to –8), *mature* (days -7 to +7), *decline* (days +8 to +22) and *decay* (days +23 to +37). The 1000 hPa field has a statistically significant anomaly at the onset time, which is not zonally symmetric, but is composed mostly of planetary scale waves. The stratosphere develops a strong anticyclone over the pole in the growth phase that continues through the decline phase. The near surface signal (1000 hPa) is unclear in the growth phase, but persists through to the decay phase with a polar structure that is similar to that in the stratosphere (50 hPa). This apparent downward propagation of the polar signal is consistent with the composite analysis of Baldwin and Dunkerton (1999).

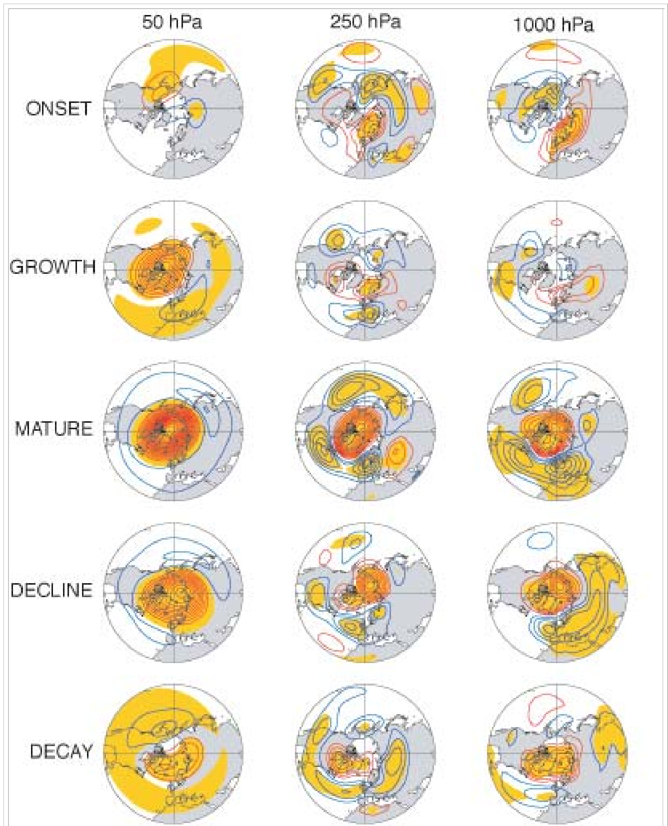

Figure 1. Composite geopotential height anomalies at 50, 250 and 1000 hPa for five phases of stratospheric warmings for 39 warming events from NCEP/NCAR Reanalysis during the period 1958-2001. Contour intervals are 3, 1 and 0.5 decameters. Negative contours are blue, positive red. Zero contours are omitted for clarity and 95% confidence limits are shown as yellow shading.

The meridional mass movements indicated by the geopotential field must be associated with meridional transport of zonal momentum. **Figure 2** shows components of the composite momentum balance from days –40 to +40. Panel (c) shows the zonal mean wind, which is very similar to the plots of Baldwin and Dunkerton (1999). The strongest upward EP flux anomalies occur about day –25, and these are associated with the strongest reduction of zonal wind in the upper stratosphere. This is the preconditioning phase of the warming (Mcintyre, 1982). Later, around day –8 a second burst of upward EP flux occurs, which is associated with a secondary wind reduction that penetrates deeper and eventually reaches the surface around day zero. The composite warming thus seems to have two phases and two time scales. A longer time scale is associated with the preconditioning of the vortex by an earlier wave forcing event, and the shorter time scale is associated with the major warming itself, which penetrates to the surface.The main warming event is associated with very anomalous meridional EP fluxes and associated mean meridional circulations in the troposphere, shown in panels (a) and (d). This tropospheric meridional EP flux anomaly is associated with synoptic-scale waves with zonal wavenumbers greater than 3 (Limpasuvan *et al.*, 2003). Thus, a response by tropospheric synoptic waves seems to be responsible for the final meridional shift of the tropospheric zonal winds and its associated weather and climate consequences. The important role of tropospheric synoptic scale waves is consistent with previous diagnostic studies of the momentum budget of tropospheric annular modes (Lorenz and Hartmann, 2003).

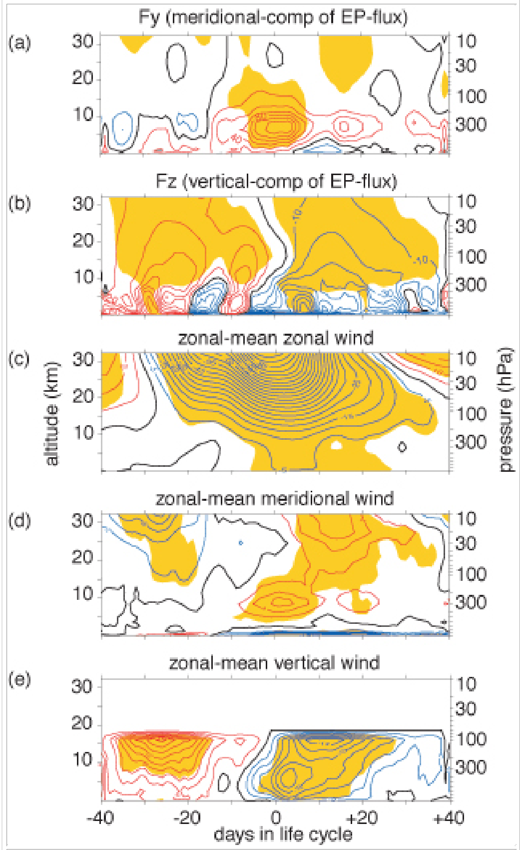

Figure 2. Anomalies of various dynamical quantities averaged poleward of 50°N (except vertical wind, which is averaged poleward of 65°N). Negative contours are blue, positive contours are red and zero contour is black.

a) meridional component of EP flux, c.i. 2 x10^7 kg s^{-2},

b) vertical component of EP flux, c.i. 1 x10^7 kg s^{-2},

c) zonal mean wind, c.i. 5 m s^{-1},

d) mean meridional wind, c.i. 2 ms^{-1},

e) vertical wind, c.i. 2 mm s^{-1}.

The 95 % confidence limits are shown as yellow shading.

The downward propagation of the signal from the stratosphere is likely to be closely associated with the concept of downward control in which wave driving effects are projected downward to the flow below the level of wave driving (Haynes *et al.*, 1991). In major warmings the wave driving propagates downward to the lower stratosphere and forces a response in the troposphere. Once a zonal wind anomaly is projected into the troposphere, the resulting changes in wave propagation and baroclinic instability can result in the positive feedback that reinforces the initial signal through the intermediacy of the synoptic scale waves (Robinson, 2000).

The possibility of rather weak forcing from stratospheric changes producing much larger than expected changes in tropospheric climate has been discussed in several places and has occurred in some global model simulations (Hartmann *et al.*, 2000; Shindell *et al.*, 2001). Because of the nonlinearity of the stratospheric warming dynamics, small changes in wind in the stratosphere or troposphere can lead to changes in the probability of major stratospheric warmings, which can have large effects on stratospheric and tropospheric climate.

Consideration of wave propagation effects on stratosphere-troposphere (ST) coupling is facilitated by index of refraction arguments. The index of refraction for stationary waves is defined in (1) for a zonal wavenumber k (Matsuno, 1970).

$$n^2 = \frac{[q]_\phi}{[u]} - \left(\frac{k}{a\cos\phi}\right)^2 - \left(\frac{f}{2NH}\right)^2 \qquad (1)$$

Where *[u]* is the zonal mean zonal wind, f is the Coriolis parameter, N is the buoyancy frequency, H is the scale height, a is the radius of Earth and ϕ is latitude. The meridional gradient of potential vorticity $[q]_\phi$ is given by,

$$[q]_\phi = \frac{2\Omega}{a}\cos\phi - \frac{1}{a^2}\left(\frac{([u]\cos\phi)_\phi}{\cos\phi}\right)_\phi - \frac{f^2}{\rho_0}\left(\rho_0 \frac{[u]_z}{N^2}\right)_z \qquad (2)$$

where Ω is the rotation rate of Earth, z is height and ρ_0 is the mean density.

Geometric optics arguments suggest that waves should be refracted toward regions of larger index of refraction. A rearrangement of (2) yields,

$$[q]_\phi = \frac{2\Omega}{a}\cos\phi - \frac{1}{a^2}\left(\frac{([u]\cos\phi)_\phi}{\cos\phi}\right)_\phi - \frac{f^2}{N^2}[u]_z\left(\left(\ln N^2\right)_z + \frac{1}{H}\right)_z - \frac{f^2}{N^2}[u]_{zz} \qquad (3)$$

which divides the potential vorticity gradient into contributions from planetary and meridional wind shear (first two terms), vertical shear (third term) and vertical curvature (fourth term in (3)). Both positive vertical shear and positive vertical curvature act to decrease the index of refraction from (3), as does increasing the zonal wind in (1). Therefore, waves should be refracted toward regions of weak westerly winds, weak wind shear and weak or negative wind curvature, with a tendency for equatorward propagation that becomes stronger nearer the equator because of the planetary vorticity gradient (2).

From (1-3) we see the potential for a positive feedback between stronger winds and weaker wave forcing, if the stratospheric jet is poleward of the primary source of planetary wave forcing in the troposphere. If the jet is stronger and has stronger shear, then planetary waves are more likely to be refracted toward the equator and less likely to propagate into the vortex and weaken it. This reasoning may explain why planetary waves are more likely to penetrate the vortex when the tropospheric midlatitude jet is displaced equatorward, which may make stratospheric warmings more likely when the Northern Annular Mode is in its negative phase (Limpasuvan and Hartmann, 1999). In contrast, when the tropospheric jet is displaced poleward, planetary waves are more likely to be refracted toward the equator.

The effect of ozone depletion on tropospheric climate through annular modes seems understandable from basic principles, but the effect of greenhouse cooling of the stratosphere on annular mode variability seems less clear. Ozone depletion is larger in high latitudes if the stratosphere there is sufficiently cold and isolated, so that in the springtime an increased gradient in ozone heating can be expected. Also, since the total ozone column is greater in higher latitudes and the effect of ozone heating penetrates deeper into the atmosphere there, depletion of ozone leads to reduced polar lower stratospheric temperatures in springtime. This acts to stabilize the vortex, resulting in reduced probability of major warmings and delayed spring warming. From these considerations a prediction of ozone depletion leading to a stronger vortex emerges, and this seems to be born out both by observations and by modelling experiments. Greater winds and fewer stratospheric warmings seem to result from ozone depletion, and this appears to have a secondary effect on surface climate that is propagated through the annular modes of variability. Gillett and Thompson (2003) were able to simulate realistic changes in SH climate with only the forcing associated with ozone depletion.

The effect of GHGs on stratospheric warming probabilities is less clear. The polar stratosphere will not obviously cool more significantly than the tropical stratosphere. Moreover, a tendency for polar amplification of warming near the surface would seem to work against enhanced meridional temperature gradients in the stratosphere. One might expect that warming of the tropical upper troposphere and cooling of the polar stratosphere by greenhouse gas increases, would lead to increased temperature gradients on constant pressure surfaces in the upper troposphere – lower stratosphere region, and that this would increase the vertical shear and refractive index in mid-latitudes. It is not clear that these changes would be far enough poleward to produce a positive feedback on the stratospheric polar night jet, however. Moreover, many other changes would occur in the troposphere that would produce effects. Gillett *et al.* (2003) found consistent positive annular mode responses (increased winds at high latitudes) in response to CO_2 increases, but the magnitudes of these changes were not large. Kushner *et al*. (2001) also found that CO_2 increases moved the eddy-driven jet poleward in the SH in a transient coupled model experiment.

In addition to the complications associated with interpreting the sign of refractive index changes, one must consider changes in the wave forcing that may result from changes in zonal wind in the troposphere. Taguchi and Hartmann (2003) found that while increasing the temperature gradient in the stratosphere led to index of refraction changes in the stratosphere that should have lead to a weaker wave drag on the vortex, the increase in high latitude surface winds that resulted from the annular mode shift increased the magnitude of the topographic planetary wave forcing. The increase in wave forcing overwhelmed the effect of the index of fraction, so that the increased upward EP flux compensated for changes in index of refraction that would have ducted waves more toward the equator. The result was a large negative dynamical feedback, in the sense that dynamical heating increased to compensate increased radiative cooling near the pole. A similar enhancement of planetary wave generation was found by Gillett *et al.* (2003).

Summary and Conclusion

Thermally forced temperature changes in the stratosphere associated with human-induced ozone depletion and greenhouse gas increases are larger than temperature changes nearer the surface of Earth. The dynamical response to these changes can be important and can be translated into changes in surface climate that are larger, or structured differently than those expected from direct forcing in the troposphere. The response to ozone depletion seems to consistently give a stronger and more persistent stratospheric vortex, which expresses itself as a positive anomaly of the Annular mode variability.

The stratospheric warming events in both hemispheres seem particularly important in enforcing a ST connection through the annular modes. Compositing of the NH warmings in a 44-year data set suggests that the synoptic-scale waves are especially important in producing the shift in tropospheric wind patterns, which in the stratosphere is driven primarily by wave forcing from planetary scale waves. This further suggests that the stratospheric wave drag and zonal wind responses are able to induce a transition in the naturally-occurring tropospheric mode of variation.

Despite considerable effort, the response of annular modes to carbon dioxide increase seems more uncertain than the response to ozone decreases. Many additional questions remain concerning the response of the coupled ST system to greenhouse gas increases and the associated climate change. Among the key questions are the following. How will winter and spring planetary wave driving change in response to global warming? How will the Brewer-Dobson circulation respond to climate change? How will climate change affect the temperature of the tropical tropopause? How will stratospheric water vapour change in the future, and how will this interact with the climate? Much of interest and importance remains to be done.

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