Internal variability in a perpetual January integration of a troposphere–stratosphere–mesosphere GCM

By STEVEN PAWSON*, ARNOLD MEYER† and SILKE LEDER

Freie Universität Berlin, Germany

(Received 21 September 1993; revised 3 June 1994)

SUMMARY

A 1200-day perpetual January integration of the Berlin troposphere–stratosphere–mesosphere general circulation model (GCM) has been performed with constant boundary conditions. The long-term mean climatology represents the response to the different forcing terms in the zonal-mean momentum budget: the winter stratospheric mean climatology is determined by the balance between the mean meridional circulation (m.m.c.), acting via the Coriolis force, and the Eliassen–Palm flux divergence. The winter stratospheric mean state oscillates between periods of strong wind (undisturbed conditions) and weaker winds (following minor warmings) on periods of several hundred days; this is the response to transient eddy forcing from the troposphere which has an almost-red power spectrum, with maximum power on time-scales of tens of days. As previously found in observations, there is a near cancellation between the wave- and the m.m.c.-induced forcing of the mean flow, but the acceleration is well correlated with the wave forcing. The dominant spatial modes of variability of the GCM are isolated using principal-component analysis. The dominant orthogonal modes of wind and temperature extend from the winter into the summer hemisphere. They represent structures close to thermal-wind balance, with a dipolar structure in the lower stratospheric temperature field associated with variations in the strength of the polar-night jet and weaker wind anomalies in the summer subtropics (22°S, 5 hPa). There is some evidence of long-period fluctuations in the vertical velocity in the lower tropical stratosphere associated with the high-latitude velocity variations; these are consistent with the equatorial temperature anomalies. Time-delayed point correlations with the zonal velocity at 62°N, 1 hPa reveal that the weak anticorrelation of velocity in the northern subtropics (28°N, 1 hPa), which is weak at zero-lag, increases with time to about 58% at 20-days lag. Singular-spectrum analysis of the zonal velocity at the three reference points, which isolates orthogonal modes of temporal variability, reveals that the correlations are increased when only the low-frequency velocity variations are considered.

1. INTRODUCTION

The terrestrial atmosphere displays variability on all time-scales. On time-scales of longer than a few days this variability is often associated with external forcing mechanisms which affect the magnitude or distribution of energy input, leading to an atmospheric response to the changed forcing. An often neglected factor is the inherent nonlinear nature of the circulation, which means that the atmosphere can undergo internal variability that is unrelated to external forcing. Understanding the circulation requires the ability to separate these two mechanisms of variability.

The earth's middle atmosphere displays considerable year-to-year variability in winter, particularly in the dynamically active northern hemisphere. This variability has been associated with various causal mechanisms (e.g. Labitzke 1982) including:

(i) interannual differences in the forcing by planetary long waves from the troposphere, arising from features such as blocking (Labitzke 1965) or the El Niño–Southern Oscillation (van Loon and Labitzke 1987);

(ii) changes in wave-propagation and wave–mean flow interaction in the middle atmosphere, for example due to the phase of the tropical quasi-biennial oscillation (QBO) (Holton and Tan 1980) which affects the location of the zero wind line and hence the direction of wave propagation;

(iii) the controversial issue of 11-year variations in solar forcing (Labitzke 1987); and

(iv) the radiative impact of volcanic aerosol loading (Labitzke and van Loon 1989).

* Corresponding author: Institut für Meteorologie, Freie Universität Berlin, Carl-Heinrich-Becker-Weg 6–10, 12165 Berlin, Germany.
† Present affiliation: Deutscher Wetterdienst, Offenbach am Main.
In order to understand and interpret the effects of these mechanisms it is essential to examine if there is any 'natural' or unforced variability of the middle atmosphere, and to understand any such variability. This is particularly true if external forcing acts to change the frequency with which the stratosphere falls into one of its natural modes, rather than driving anomalous circulations, as has been suggested by Taubenheim et al. (1989), Kodera et al. (1990), and Pierce and Fairlie (1993).

The fundamental mechanism of internal variability of the stratosphere was shown in the one-dimensional (1-D) numerical model of Holton and Mass (1976), also analysed by Yoden (1987); steady planetary-wave forcing imposed at the lower boundary of a simple quasi-geostrophic model, in which the mean state was relaxed back to a reference state, induced a transient response. The model went through a cycle of wave-forced sudden warmings, which led to a reversal of the wind, thereby cutting off the wave forcing by the Charney–Drazin (1961) criterion for planetary-wave propagation; a slow radiative relaxation back towards a westerly wind then occurred, allowing further wave propagation, leading to the possibility of further warmings.

If such mechanisms are relevant to the atmosphere they must be studied by using observed data as well as in more complicated numerical models, which include a more detailed representation of the atmospheric flow and physical processes at work. Work by Pierce and Fairlie (1993) has isolated different structures ('preferred flow regimes') in 10 years of daily stratospheric observations which support the concepts of the Holton–Mass model. Studies with more complicated models of the middle atmosphere have generally been restricted to the simulation of single events (sudden warmings) or have concentrated on the climatology of the system. Little attention has been devoted to the transient behaviour of the stratosphere in three-dimensional models with full physical parametrizations.

This study is thus concerned with the variability of the stratosphere in an atmospheric system with no variations in the external forcing. The simplest possible system has been obtained by neglecting even the annual cycle. A moderately long (1200 day) perpetual January integration of a troposphere–stratosphere–mesosphere general circulation model (TSM GCM) has been performed. The model was integrated with constant boundary conditions (incoming solar flux, specified sea surface temperature, and specified deep-soil temperature and water content) in order to examine the internal dynamical response of the model atmosphere. For the stratosphere this means that all of the variability is either internally generated or is a response to the tropospheric forcing and its variations.

The analysis is directed towards quantifying and understanding the zonal-mean structure of the middle atmosphere in this TSM GCM and its long-period variability. The GCM is described in the following section. In section 3 the long-term zonal-mean structure is presented and the balance of forces leading to this state is analysed; this serves to validate the model performance against reality and allows comparison of the dynamical forcing mechanisms with those of other GCMs. In section 4 some aspects of the transient behaviour in the GCM are analysed; in particular, a transition from a strong jet structure in the northern stratosphere to easterly winds (a minor warming) is examined. The standard deviations of the zonal-mean velocity and temperature fields are presented. This leads the way to a discussion of the long-period, global-scale variations found in the model, which are analysed with the help of statistical techniques, including principal-component (PC) analyses. In section 5 the global-scale modes of variability are presented. Section 6 concentrates on the growth and decay of the dominant flow structures and the relationship between the northern polar regions and the subtropics in each hemisphere. In the final section the major results are discussed and some conclusions are drawn.
2. DESCRIPTION OF THE BERLIN TSM GCM

The TSM GCM includes a full set of physical parametrizations for surface exchange processes, the hydrological cycle and clouds, radiative heating, and subgrid-scale processes. It was modified from ‘cycle 18’ of the 16-level European Centre for Medium-range Weather Forecasts (ECMWF) forecasting model (Heckley 1985; Arpe and Klinker 1986) with an upper boundary at 25 hPa, by the addition of 14 more middle-atmospheric levels and the inclusion of appropriate physical parametrizations for the middle atmosphere. The hybrid vertical coordinate of Simmons and Strüffing (1983) is used; pure pressure levels are defined in the stratosphere with a resolution close to 3.5 km in pressure scale height; the resolution decreases in the three levels below the upper boundary at 0.01 hPa (near 80 km). The stratospheric pressure levels and the approximate height of these levels are tabulated in Table 1. The GCM is formulated spectrally; for economy, a low resolution (T21) was used for the integrations described here. Nonlinear dynamics and parametrized processes are performed on the associated Gaussian grid, which has a resolution of about 5.625° in latitude and longitude.

<table>
<thead>
<tr>
<th>Pressure (Pa)</th>
<th>Scale height (km)</th>
<th>Rayleigh friction coefficients (days⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.00</td>
<td>80.68</td>
<td>0.25</td>
</tr>
<tr>
<td>2.50</td>
<td>74.27</td>
<td>0.24</td>
</tr>
<tr>
<td>4.50</td>
<td>70.15</td>
<td>0.22</td>
</tr>
<tr>
<td>8.00</td>
<td>66.13</td>
<td>0.17</td>
</tr>
<tr>
<td>14.00</td>
<td>62.21</td>
<td>0.11</td>
</tr>
<tr>
<td>25.00</td>
<td>58.15</td>
<td>0.05</td>
</tr>
<tr>
<td>44.50</td>
<td>54.11</td>
<td>0.02</td>
</tr>
<tr>
<td>79.00</td>
<td>50.10</td>
<td>0.01</td>
</tr>
<tr>
<td>140.00</td>
<td>46.09</td>
<td>0.00</td>
</tr>
<tr>
<td>248.50</td>
<td>42.07</td>
<td></td>
</tr>
<tr>
<td>442.50</td>
<td>38.04</td>
<td></td>
</tr>
<tr>
<td>783.50</td>
<td>34.04</td>
<td></td>
</tr>
<tr>
<td>1400.00</td>
<td>29.97</td>
<td></td>
</tr>
<tr>
<td>2500.00</td>
<td>25.91</td>
<td></td>
</tr>
<tr>
<td>4250.00</td>
<td>22.20</td>
<td></td>
</tr>
<tr>
<td>7682.44</td>
<td>18.06</td>
<td></td>
</tr>
<tr>
<td>12784.21</td>
<td>14.49</td>
<td></td>
</tr>
<tr>
<td>18563.65</td>
<td>11.88</td>
<td></td>
</tr>
</tbody>
</table>

The model is the same as in Pawson et al. (1991), except for some aspects of the calculation of radiative heating rates, which are described in Pawson (1992). In the solar part of the spectrum, the models of Shine and Rickaby (1989) and Strobel (1978) are used for ozone (O₃) and oxygen (O₂) heating respectively. The implementation of the O₃ heating has been improved over that in Pawson et al. by the use of model-generated albedos for the heating due to reflected radiation in the Chappuis and Huggins bands in the low stratosphere.

Long-wave radiation is treated with the models of Geley and Hollingsworth (1979) and Shine (1987) in the troposphere and stratosphere respectively. In the stratosphere, a new carbon dioxide (CO₂) transmittance matrix is calculated from the zonal-mean temperature at each latitude every 12 hours (the ‘full radiation time steps’). Ideally, a transmittance matrix would be used for each column, but this is too expensive (requiring 64 times more computing power at the full radiation time steps with T21 resolution).
Although an option for non-LTE (local thermodynamic equilibrium) processes in the 15 \( \mu m \) bands of CO\(_2\) is included in Shine's (1987) radiation scheme, it was not used as it has a negligible effect on the climatology of a model with an upper boundary at 80 km; tests with shorter runs of the GCM showed smaller differences than the fixed dynamical-heating-model calculations of Pawson and Shine (1991), presumably due to the strong influence of the enhanced diffusion and the Rayleigh friction at the upper boundary. Zonal-mean transmittance matrices were also used for O\(_3\) at 9.6 \( \mu m \); the ozone distribution is specified.

The effects of small-scale dissipation in the mesosphere are represented by a linear relaxation (Rayleigh friction) of the wind which damps the wind to zero at the highest model levels. The vertical structure of the dissipation is of the hyperbolic-tangent form introduced by Holton and Wehrbein (1980):

\[
\gamma = -0.125 \times \left\{ 1 + \tanh\left( \frac{z - 63}{7.5} \right) \right\}
\]

where \( z \) is the height in km. These coefficients result in the vertical profile listed in Table 1; they are effective only above 50 km and the upper mesosphere is thus a 'sponge layer'.

The lower boundary condition on the model is that there is now flow through the surface (a half-level of the vertical coordinate system). Likewise, the upper boundary condition is of no flow out of the top of the model. Sea surface temperature and deep-soil moisture and temperature are fixed at climatological values, as in the original ECMWF model.

Note that since these integrations were performed the Berlin TSM GCM has been reformulated using more recently developed physical parametrizations. In particular, a radiation scheme valid at all levels is being implemented to allow GCM studies of climate change. However, the variability of the zonal-mean state of the reformulated GCM is qualitatively similar to that studied here, so that the interpretation of the results is thus expected to remain similar, even if some details would be different if the new GCM had been used.

The integration analysed here was performed for 1200 days, starting from an isothermal, motionless initial field. Only the last 1000 days are analysed since the spin-up time of the model is about 150 days in the low tropical stratosphere, a region characterized by long relaxation times and subjected to a variety of forcing mechanisms whose residual is small; use of these early data in the analysis could contaminate the results. Data were saved daily, though this is an inadequate time resolution for examining changes on time-scales of less than about 5 days, even with the diurnal cycle turned off, due to aliasing which becomes important in the high stratosphere and mesosphere. Some weak filtering was thus applied to the fields analysed in this study to remove the high-frequency variability, leaving only time-scales of 5 days or greater. This is justifiable for this study, which emphasises the low-frequency variability, but it would be inappropriate for studies of the short-term dynamics; it leads to an imbalance in the zonal-mean forcing of several m s\(^{-1}\) day\(^{-1}\) in the mesosphere during highly transient periods.

Note that there are further imbalances in the forcing arising from the diffusion terms in the model (the horizontal diffusion is increased in the mesosphere and the vertical diffusion plays a role in the troposphere) and the subgrid-scale momentum transport by parametrized convection in the troposphere. A further source of error is that the diagnostics were performed on data interpolated to a grid with 27 log-pressure levels using a different numerical differencing scheme to that used on the 30 hybrid levels in the model.
3. THE ZONAL-MEAN STRUCTURE OF THE GCM

In this section the long-term mean structure of the GCM is presented; this serves both as a validation of its performance in comparison with other models and, more importantly, with the real atmosphere. Attention is devoted to the zonal-mean zonal wind ($\bar{u}$) and its forcing.

(a) Mean structure

The mean zonal-wind structure (Fig. 1) has several realistic aspects as well as some deviations from climatologies derived from satellite radiance data under the assumptions of hydrostatic balance and small Rossby number (e.g. Fleming et al. 1990). Generally, the summertime (southern hemispheric) easterly jet is well simulated, sloping from the tropical lower stratosphere to about 45°S in the mesosphere; the jet is not so well closed off as in reality, but this could be amended by increasing the strength and depth of the friction coefficient, though this would have the disadvantage of constraining the stratospheric flow. More serious in the summer hemisphere is that the tropospheric westerlies extend to about 40 km high. This problem could be rectified by including a suitable gravity-wave drag parametrization (Palmer et al. 1986; McFarlane 1987), but this was avoided here because of the uncertain physical assumptions involved in such schemes and the numerous feedbacks which can occur. A similar problem exists in the northern hemisphere, where the subtropical jet is too strong and is not well separated from the polar-night jet (PNJ). The PNJ itself does not slope upwards and equatorwards as in reality, but has a maximum near 70°N at all stratospheric levels. The maximum is too weak (peaking at 47 m s\(^{-1}\) below the stratopause). Such deficiencies are seen in most other TSM GCMs, an exception being that of Rind et al. (1988) who include a comprehensive gravity-wave parametrization in their model.

The corresponding mean temperature field displays the associated features expected from the near thermal-wind balance of the modelled atmosphere; in particular, a cold low polar-night stratosphere is associated with the connected jets, the tropical tropopause is too high and too cold (Pawson 1992), but the summer stratopause is well modelled.

(b) The transformed Eulerian mean meridional circulation

The forcing mechanisms leading to the modelled zonal-mean state of the GCM will be interpreted using the 'transformed Eulerian mean' formulation of the zonally averaged

Figure 1. The 1000-day mean zonal-mean zonal wind field ($\bar{u}$, m s\(^{-1}\)). Negative-valued (easterlies) contours are dashed.
equations of motion, introduced by Andrews and McIntyre (1976). The transformed Eulerian mean meridional circulation (m.m.c.), \((\bar{u}^*, \bar{w}^*)\), is a circulation in the latitude-height plane with a non-divergent mass flux, satisfying the mass continuity equation. Andrews and McIntyre showed that this is a more natural circulation with which to study the dynamics of an atmosphere forced by planetary waves, since it avoids the exact cancellation of the northward eddy heat flux due to vertical temperature advection by steady, linear, conservative quasi-geostrophic waves which is inherent in the standard Eulerian mean equations of motion.

The 1000-day mean m.m.c. from the GCM (Fig. 2) shows northward flow in the mesosphere, peaking at 3.5 m s\(^{-1}\) near 20°S, 70 km, with ascent above 40 km in the summer hemisphere and descent in the winter hemisphere. The summer–winter circulation cell related to the closed jets and reversed temperature gradient is thus modelled. The strength of the northward flow in the mesosphere is slightly weaker than observations suggest (around 5 m s\(^{-1}\) maximum), which is most likely a consequence of the proximity of the upper boundary and the extra dissipation at these levels. The strongest descent (1 cm s\(^{-1}\)) occurs in the polar-night lower mesosphere. There is a region of strong ascent

![Figure 2](image-url)

Figure 2. The 1000-day mean transformed Eulerian mean meridional circulation: (a) \(\bar{u}^*\) (m s\(^{-1}\); contour interval 0.5 m s\(^{-1}\)) and (b) \(\bar{w}^*\) (cm s\(^{-1}\); contour interval 0.125 cm s\(^{-1}\)). Negative values (i.e. southward flow or descent) are shaded.
and some southward flow at the northernmost grid points in the stratosphere region; this is not expected and could arise from the poor separation between the zonal mean and the eddies at these polar latitudes (see Boville 1986). In the troposphere, the main features of the Hadley circulation and the extra-tropical circulation are well captured, although these are quite difficult to discern in Fig. 2.

(c) Forcing of the mean state

The analysis of the forcing of the model has been performed on a logarithmic-pressure vertical coordinate: \( z = H \log(\rho_0/\rho) \), following interpolation of the data from the model grid to standard pressure levels; here, \( H = 7 \) km is the pressure scale height and \( \rho_0 = 1013.25 \) hPa is the reference surface pressure. The basic state density is then \( \rho_0 = \rho_s e^{-z/H} \), where \( \rho_s = 1.25 \) kg m\(^{-3}\) is the surface density.

In the formulation of Andrews and McIntyre (1976) the zonal-mean zonal momentum equation is:

\[
\bar{u}_t = \bar{v}^*(f + r^{-1} \bar{u} \tan \phi) - r^{-1} \bar{v}^* \bar{u}_\phi - \bar{w}^* \bar{u}_z + D + X
\]

(2)

(where \( f \) is the Coriolis parameter, \( r \) the radius of the earth, and \( \phi \) latitude) which expresses the acceleration of \( \bar{u} \) as the response to the various forcing terms: the modified Coriolis term (incorporating the planetary curvature effect), the northward and vertical advection of \( \bar{u} \) by the m.m.c., and the body forces due to 'large-scale' (D) and 'small-scale' (X) zonally asymmetric disturbances. The differentiation between \( D \) and \( X \) in the GCM is that \( D \) incorporates the forcing from the resolved disturbance while \( X \) represents the effects of vertical and horizontal diffusion, as well as the Rayleigh friction at the uppermost levels.

The large-scale eddy forcing, \( D \), can be expressed in terms of the divergence of the Eliassen–Palm (E–P) (1961) flux, \( \mathbf{F} \), a vector in the meridional plane:

\[
D = (\rho_0 f \cos \phi)^{-1} \nabla \cdot \mathbf{F}.
\]

(3)

Within the restrictions of WKBJ theory, when the concept of group velocity is valid, \( \mathbf{F} \) is parallel to the group velocity and its magnitude is proportional to the wave-activity density of the wave packet (Andersson and McIntyre 1976). \( \mathbf{F} \) is defined as:

\[
\mathbf{F} = (0, F^\theta, F^z) = \rho_0 f \cos \phi (0, \bar{u}_z v'/\theta_z - \bar{u}' v^z, (f - \bar{u}_\phi v') v'/\theta_z - \bar{u}' w^z).
\]

(4)

Note that within the limitations of quasi-geostrophic theory, the northward and vertical components are respectively proportional to the northward fluxes of zonal momentum and heat.

For a trend-free system the temporal average of the individual forcing terms in Eq. (2) display a balance which determines the mean structure of the atmosphere. This was assumed by Boville (1986) who studied the eddy forcing of the zonal-mean state in a perpetual January integration of a GCM integrated for 370 days. Boville analysed two 90-day periods separated by a sudden warming. Andrews et al. (1983) used this formalism to analyse the Geophysical Fluid Dynamics Laboratory SKYHI model and found a near cancellation between the Coriolis and large-scale eddy forcing terms. Such compensation was found in observations by Shiotani and Hirota (1985) and in the Goddard Institute for Space Studies TSM GCM (Rind et al. 1988).

There is a similar compensation between \( D \) and \( f \bar{u}^* \) (Figs. 3(a) and (b)) in the wintertime stratosphere of the GCM. In the mesosphere the deficit of up to 15 m s\(^{-1}\) arises from the unresolved terms; the Rayleigh friction contribution to this (Fig. 3(c)) accounts for most of the deficit. The remaining three terms (Figs. 3(d)–(f)) play a much smaller role at all locations; the meridional advection of zonal velocity maximizes in the
Figure 3. The 1000-day mean of forcing terms for $\bar{u}$ (m s$^{-1}$ day$^{-1}$) from Eq. (2) calculated daily and then averaged in time: (a) $D$, as defined by Eq. (3) (contour interval 5 m s$^{-1}$ day$^{-1}$); (b) $f\bar{u}^*$ (contour interval 5 m s$^{-1}$ day$^{-1}$); (c) $X$ (contour interval 2.5 m s$^{-1}$ day$^{-1}$); (d) $-\bar{v}^*\bar{u}_g$ (contour interval 1 m s$^{-1}$ day$^{-1}$; the 0.5 m s$^{-1}$ day$^{-1}$ contour is also shown).
Figure 3. Continued. (e) $a^{-1} \overline{\omega^* u \tan \phi}$ (contours as in (d)); (f) $-\overline{w^* u}$; (contour interval 0.25 m s$^{-1}$ day$^{-1}$); (g) $D$ (as in (a)) between 2 and 12 km (contour interval 5 m s$^{-1}$ day$^{-1}$); (h) $\tilde{f} \overline{u^*}$ (as in (b)) between 2 and 12 km (contour interval 5 m s$^{-1}$ day$^{-1}$). In all panels negative (i.e. westward) forcing is shaded.
subtropical summer mesosphere where $\vec{\nu}^*$ is largest; the geometrical correction assists the Coriolis forcing and the vertical advection is almost negligible.

In the troposphere the dominant terms in the budget are $D$ and $f\vec{\nu}^*$, which again almost cancel; these two contributions are shown in more detail between the approximate heights of 2 and 12 km (Figs. 3(g) and (h)). Note that in the troposphere the approximations involved in using this height coordinate are less good than in the middle atmosphere, at low levels the surface topography begins to be important and the parametrized vertical exchange with the surface (which was not saved during the integration) plays a major role below 2 km.

4. TEMPORAL VARIABILITY OF THE ZONAL-MEAN STATE

During the 1000 days studied here the GCM displayed similar behaviour to that found by Boville (1986), with periods of strong stratospheric winds which break down and recover over long periods. The current integration was long enough for this to occur several times, the strong PNJ recovering between each of the minor warmings, similarly to the idealized model of Holton and Mass (1976). Note that only minor warmings occurred in this GCM; there was no reversal of the zonal-mean wind at 10 hPa, which is one of the criteria of a major warming. In this section the temporal evolution of the zonal-mean state of the stratosphere and the forcing mechanisms are examined in some detail.

(a) An example of a minor warming in the GCM

As an illustration of the temporal variability of the zonal-mean state, the period days 601–750 is examined in some detail in this section. The latitudinal structure of $\vec{u}$ at 1 hPa is shown in Fig. 4. At the beginning of the 150-day period the stratospheric wind exceeds 100 m s$^{-1}$ at 1 hPa) and the 40 m s$^{-1}$ contour extends to 50$^\circ$N. Over the next period there were several instances of a deceleration of the stratospheric westerlies, beginning near days 620, 660 and 720; each of these spread from the jet maximum into the subtropics. The changeover to easterlies occurred near day 665, when they reached 40$^\circ$N at 1 hPa, before retreating polewards. The time–height structure of $\vec{u}$ in the stratosphere and mesosphere at 62$^\circ$N (Fig. 5(a)) shows that these easterlies winds only extended downwards to 40 km so that the event can only be regarded as a minor warming.

The stratosphere is generally considered to be forced by planetary waves propagating from the troposphere. These can be quantified by the time series of $F^{(2)}$ at a level in the low stratosphere; this time series at 100 hPa (Fig. 5(b)) shows the upward propagation of wave energy into the stratosphere in middle latitudes. (Note also the occasional occurrence of weak downward-directed E–P flux vectors in the polar regions and the subtropics.) The decelerations in $\vec{u}$ (Figs. 4 and 5(a)) coincide with pulses of wave activity propagating from the troposphere (Fig. 5(b)). There is no precise correspondence between the upward-propagating wave energy at the tropopause and the acceleration of the mean flow at that latitude: there are several reasons for this, all of which have been discussed in the literature. Firstly, the dependence of the wave propagation on the mean flow structure, observed by Hirota and Sato (1969), Palmer (1981), Al-Ajmi et al. (1985) and Shiotani and Hirota (1985) and modelled by Matsuno (1971) and Karoly and Hoskins (1982), who showed the general tendency of planetary waves to propagate upward and equatorward in a spherical atmosphere, which is modified by the potential-vorticity gradients of the zonal-mean state that are included in Matsuno’s refractive index. Secondly, the dissipation of the waves in the stratosphere means that $F$ is not conserved. Thirdly, $D$ is not the only forcing mechanism; Shiotani and Hirota (1985) showed that
Figure 4. Time series, days 601–750, of $\bar{u}$ (1 hPa). The contour interval is 20 m s$^{-1}$. Regions of easterly wind are lightly shaded and regions with westerlies of 40–60 > 80 m s$^{-1}$ are heavily shaded.

Figure 5. Time series, days 601–750, of: (a) $\bar{u}$ (62$^\circ$N) between 20 and 80 km. The contour interval is 20 m s$^{-1}$. Regions of easterly wind are lightly shaded and regions with westerlies of 40–60 > 80 m s$^{-1}$ are heavily shaded. (b) $P^{\text{GO}}$ (100 hPa) (kg s$^{-1}$day$^{-1}$) as defined by Eq. (4) in the latitude range 30–90$^\circ$N. The contour interval is 0.5 kg s$^{-1}$day$^{-1}$ and negative values (i.e. downward-directed flux) are shaded.
the indirect wave forcing due to the induced m.m.c. (dominated by \( f \tilde{v}^* \)) tends to cancel \( D \), as discussed for the long-term mean of this GCM in the previous section.

Figure 5 shows that the stratosphere of the GCM is subject to a series of forcing pulses from the troposphere, which progressively erode the zonal-mean wind in the stratosphere. This culminates around day 665 in the reversal of the mean wind to easterlies which extends down to 40 km; the pulse which causes this warming is slightly larger than the previous pulses at 100 hPa but is about 25% larger at 10 hPa. This is due to the so-called preconditioning of the stratosphere from the previous wave pulses which erode the polar vortex and progressively allow the wave pulses to enter the polar region, causing the mean flow reversal there.

(b) Forcing mechanisms

The corresponding time series of the forcing at 62°N, 10 hPa are shown in Fig. 6. As for the long-term mean case there is often a near cancellation between the two dominant terms, the Coriolis forcing and \( D \) (Fig. 6(a)). The remaining three terms are generally <10% as strong as \( D \) (Fig. 6(b)), and \((\tilde{u} \tilde{v}^*/a) \tan \phi \), which always acts to enhance \( f \tilde{v}^* \), is usually the largest. However, the near cancellation between the two dominant terms means that these smaller contributions are important in the momentum budget.

As in Shiotani and Hirota (1985), the modelled acceleration (Fig. 6(c)) tends to follow \( D \) closely (Fig. 6(a)) and is well correlated with it.

(c) Episodic behaviour in the GCM

This concept of pre-conditioning preceding the warming has often been noted in observations, e.g. Palmer (1981) and Shiotani and Hirota (1985). Shiotani and Hirota (1985) presented a concept, which is further developed here, that the planetary-wave

![Figure 6. Time series, days 601–750, of the forcing of \( \tilde{u} \) (62°N, 10 hPa) (m s\(^{-1}\)day\(^{-1}\)) due to (a) \( D \) (solid) and \( f \tilde{v}^* \) (dash-dot-dot) and (b) \( a^{-1} \tilde{v}^* \tan \phi \) (solid), \(-\tilde{v}^* a^{-1} \tilde{u}_y\) (dash-dot-dot) and \(-w^* \tilde{u}_z\) (dot). (c) The acceleration, \( \partial \tilde{u}/\partial t \) (m s\(^{-1}\)day\(^{-1}\)).](image)

forcing from the troposphere is quasi-periodic, on a time-scale of about 2 weeks, but that the mean wind variations occur on a longer time-scale, with a maximum of two warmings in each winter due to the time required to pre-condition the stratosphere, undergo the warming, and then to recover before the process is repeated. This perpetual January integration allows this concept to be investigated without the ‘contamination’ of the annual cycle, although the restrictions must be borne in mind.

Power-spectrum analyses of the 1000 days of the integration allow quantification of the time-scales of variability in the forcing and response. To enable determination of the significance of these power spectra, the red-noise spectrum, \( R(\nu) \), for a time-series of length \( N \) with mean spectral power \( P \) is calculated as first-order Markov process (Chatfield and Collins 1980):

\[
R(\nu) = P \left( \frac{1 - x^2}{1 - 2x \cos \left( \frac{2\pi \nu}{N} \right) + x^2} \right)
\]

where \( x \) is the autocorrelation coefficient at time-lag 1 day. Significance levels can be determined by the inclusion of 5% and 95% confidence levels, obtained using a \( \chi^2 \) distribution (Press et al. 1986).

For this application, Eq. (5) was applied to the 1000-day time series using their mean spectral power \( P \) (Fig. 7). The power of \( F^{(c)} (60^\circ N, 100 \text{ hPa}) \) has two fairly dominant peaks at periods close to 40 days and 80–100 days, but only the latter exceeds the 95% significance level of departure from the red-noise spectrum; it is thus difficult to distinguish this spectrum of stratospheric forcing from a random signal. The mean wind at 1 hPa has significant power on the 100–300 day time-scales, which are the longest resolved periods in the data set; from this model run, it cannot be determined if longer periods are present. This confirms that the forcing events in the model occur on a rather shorter time-scale than the response, as in the observations (Shiotani and Horita 1985), which allows for the preconditioning of the flow and the subsequent development of (minor) stratospheric warmings. It is rather undesirable that the only significant power in the GCM occurs at time-scales that are barely resolved; if computer resources were available, it would be profitable to extend the model integration for at least another 1000 days.

The broad peaks of the power spectra of the forcing and the response suggest an episodic rather than purely periodic nature of the system. This contrasts with the idealized solutions of the Holton–Mass model discussed by Yoden (1987), which are essentially periodic responses, the period depending on the amplitude of the steady lower-boundary forcing. This suggests that the transience of the forcing from the troposphere is an essential part of the stratospheric response in the GCM. This is also true in the real atmosphere, whose interannual variability is largely dependent on the occurrence and timing of sudden warmings.

(d) Standard deviations of the zonal-mean fields

The latitude–height structure of the variability of the zonal-mean state is characterized by the standard deviation of the daily data fields about the 1000-day mean (Fig. 8), which assist in the following statistical analysis. The maximum standard deviation in \( \bar{u} \) (exceeding 35 m s\(^{-1}\)) is located above and slightly poleward of the climatological jet maximum. This arises from variations in the strength of the fast-jet flows and the existence of the weak-jet structure following minor warmings. Variability at other locations is much smaller but it is relevant in the following discussions. In particular, note the local maximum in the high stratosphere near 30°N, 50 km.
Figure 7. Log$_{10}$-log$_{10}$ plots of the power spectra of (a) $\bar{u}$ (62°N, 1 hPa) (power is in m/s$^{-2}$) and (b) $F^{1/3}$ (62°N, 100 hPa) (power is in kg/s$^{-3}$day$^{-2}$). The dash-dotted lines are red-noise spectra calculated for a Markov process (Eq. (5)) with the mean power of the relevant time series; the dashed curves show the 5% and 95% significant levels of the spectra according to a $\chi^2$ test.
The standard deviation of the daily zonal-mean temperatures about the 1000-day mean field (Fig. 8(b)) also maximizes in the polar night, exceeding 10 degC near 38 km, at the base of the region of maximum variability of \( \bar{u} \). The variability in the remainder of the modelled atmosphere is smaller, except near the South Pole and in the mesosphere.

5. GLOBAL STRUCTURES: SPATIAL PATTERNS

In this section an analysis of the spatial and temporal behaviour of the evolution of the GCM is presented. The dominant pattern of variability has been isolated using empirical orthogonal function (EOF) analyses of the zonal-mean velocity and temperature fields. We begin with a brief introduction to EOF (principal component) analysis, more comprehensive developments of which are given by Koutzback (1967) or Chatfield and Collins (1980).

(a) Summary of the EOF analysis

Patterns of atmospheric variability can be found by point correlations between one point and the rest of the field, which reveal how variations at each location in the
atmosphere are connected to those at the reference point. However, simple calculation of the point correlation requires prior knowledge of the location of the ‘action centres’ that are used as the reference points. One method of locating these centres is to perform an EOF analysis of the correlation (or covariance) matrices, which isolates significant atmospheric oscillation patterns. EOF analysis identifies orthogonal vectors depicting the spatial patterns of variability of the data set. In the same manner as a 1-D data set can be represented by amplitudes of a set of orthogonal basis functions (e.g. a Fourier series), a multi-dimensional field can be expressed in terms of its EOFs. For instance, the zonal-mean zonal velocity field can be expanded in terms of the orthogonal spatial modes, $\psi_m^u$ (the EOFs), multiplied by their time-dependent amplitudes (or principal components, PCs), $a_m^u(t)$:

$$\bar{u}(x, y, t) = U + \sum_{m=1}^{M} a_m^u(t) \psi_m^u(x, y)$$  

(6)

(where $U$ is the long-term mean of $\bar{u}$). The superscript ‘u’ has been used to represent the zonal velocity; likewise, ‘T’ will represent temperature.

Kutzbach (1967) shows how to manipulate algebraically equations such as Eq. (6) to introduce an eigenvalue problem, which can be solved using standard matrix-inversion techniques. The $\psi_m^u$ and $a_m^u(t)$ are determined in the course of the solution. The routines used in this study were developed by Wang (1994) using matrix-inversion methods based on Press et al. (1986).

The summation in Eq. (6) is truncated at order $M$ since such analysis is appropriate only when a substantial amount of the variability is represented by the leading-order eigenvectors, otherwise one has almost as many eigenfunctions as original data points. Further, a certain degree of separation of the eigenvalues is required to ensure that the eigenvectors are not degenerate (Chatfield and Collins 1980) which would lead to physically meaningless eigenvectors.

Results of the EOF analysis of the zonal-mean fields are discussed here. The stability of the results was assessed by analysing restricted parts of the fields. This revealed that the dominant eigenvectors, $\psi_1^u$ and $\psi_2^u$, which respectively peak in the northern and southern hemisphere stratosphere and represent 11.9% and 7.4% of the variability of the fields, were stable in the sense that their patterns remained essentially unchanged when only the stratosphere or only one hemisphere was analysed. (Note also that use of the covariance instead of the correlation matrices in the analysis led to higher significance but slightly noisier patterns.) In contrast, $\psi_3^u$ and $\psi_4^u$, which have stronger signals in the low stratosphere and troposphere, did not remain unchanged when the EOF analysis was repeated for levels below 24 km only. This is because the troposphere undergoes a much wider range of motions than the stratosphere, and the 1000-day time series analysed here is not sufficient to allow statistically significant tropospheric patterns to be isolated, as in James and James (1992), for example. The first ten EOFs account for about 45% of the variability of the time series; the tropospheric velocity required many eigenvectors, each representing a small part of the variance, to be adequately described. Attention is therefore concentrated on the stratospheric variability.

(b) The dominant pattern of variability in the stratosphere

Figure 9(a) shows that $\psi_1^u$ maximizes in the stratosphere poleward of 60°N, where the variability is largest (Fig. 8), with a downward extension of this positive anomaly into the troposphere. There is a secondary maximum, about half as strong as the peak values, in the southern subtropical stratosphere, with a region of negative values in the
northern subtropical low stratosphere and troposphere. This pattern closely resembles
the linear point correlation of the whole field with \( \bar{u} \) (62°N, 1 hPa), which is close to the
maximum centre of \( \psi^T \).

Physically, the structure of the positive phase of \( \psi^T \) describes a global wind structure
that has a positive anomaly in the winter polar region, a weaker negative anomaly at
lower northern latitudes, and a positive anomaly at low latitudes of the southern
hemisphere. Note that \( \psi^T \) does not describe the absolute magnitudes of the velocity
variations at any point, but their variations relative to their absolute variance (see Fig.
8). Even though the absolute velocity variations in the southern hemisphere subtropics
are quite small (Fig. 8(a)) they are quite highly correlated with the variations in the PNJ.

The near thermal-wind balance of the structure is reflected in the dipolar structure
of \( \psi^T \) (Fig. 9(b)), which accounts for 13.8% of the variance of \( \bar{T} \). (Note that the time
series of \( a^T \) and \( a^T_1 \) are highly anticorrelated.) The temperature anomaly has a positive
maximum poleward of 60°N and a minimum (>90% of the magnitude of the maximum)
in the tropics; this enhances the equator–pole temperature gradient in the negative phase
of \( a^T_1 \) and is consistent with the stronger velocities when \( a^T_1 \) is positive. At higher levels
this temperature gradient is reversed, consistent with the upward decay of the anomaly in $\psi^*_r$, but it is less strongly (50%) correlated to the main pattern. The southern hemispheric features of the two patterns are also in good agreement.

Since the time series of $a^*_t$ and $a^*_r$ are highly anticorrelated, this presentation is restricted to $a^*_r$ (Fig. 10(a)). The largest oscillations are of long period, with some shorter-term variability superimposed on it. Reconstruction of the time series from the mean and this EOF (Figs. 10(b) and (c)) shows that the long-term variations in $\Bar{u}$ at 62°N, 1 hPa are described by $\psi^*_r$ but at 10 hPa, where $\psi^*_r$ is weaker, there are larger differences. This reconstruction of the time series describes 78.2% of the total variance, including the day-to-day fluctuations, of $\Bar{u}$ (62°N, 1 hPa) while at 10 hPa it is considerably less (67.9%). Power-spectrum analysis of $a^*_t$ (Fig. 11) confirms that the dominant power is at the longest periods and exceeds the significance levels of departure from the red-noise spectra, in contrast to the unprocessed velocity field at 62°N, 1 hPa (compare with Fig. 7(a)). The low-frequency variability of the zonal-mean zonal wind field in the winter stratosphere is thus described by $\psi^*_r$; similarly, the dominant mode of variability of the lower and middle stratospheric temperature is described by $\psi^*_r$.

(c) Extremes in the atmospheric structure

The global-scale variations bear some similarity to those observed in satellite radiance data by Fritz and Soules (1970) and modelled by Hsu (1980), who found anticorrelations between the winter polar region and the tropics in their studies. The state of the atmosphere in the extreme phases of $\psi^*_r$ can be shown by averaging the reconstructed

![Figure 10](image_url)

Figure 10. Time series, days 201–1200, of: (a) $a^*_r$ (dimensionless) (b) the original time series of $\Bar{u}$ (thin line) and the reconstruction $U + a^*_r\psi^*_r$ (thick line) at 62°N 1 hPa (m s$^{-1}$), and (c) as for (b) but at 62°N 10 hPa.
zonal-mean wind and temperature fields over the 100 days when $a^q$ has its maximum values and the 100 days when it has its lowest values (Fig. 12). The 100-day mean wind distributions for the maxima of $a^q$ (Fig. 12(a)) resembles the long-term mean (Fig. 1) but the PNJ is stronger, exceeding 75 m s$^{-1}$. The opposite anomaly (Fig. 12(b)) has weak easterly winds in the polar region above 10 hPa and a much weaker PNJ (25 m s$^{-1}$) displaced equatorwards and upwards from its location in the long-term mean structure to near 40°N, 60 km. These two structures are typical for the model and are apparent when one examines individual 30-day (monthly) mean periods, which resemble the fast more often than the slow PNJ structure. There are also qualitative similarities between these extreme states and the interannual differences in the zonal-mean wind derived for different winters from satellite radiance measurements (e.g. Geller et al. 1984).

The difference between these extreme states (Fig. 12(c)) shows its strongest values in the PNJ region, where the fast-PNJ state exhibits strong winds and the weak-PNJ structure has easterly weak winds, with weaker negative values centred on the subtropical stratopause. There is a weak positive difference in the summer hemisphere stratosphere. The temperature difference between these two periods (Fig. 12(d)) is consistent with thermal-wind balance. The polar-night stratosphere is 50 degC warmer at 30 km and 20 degC cooler at 60 km in the weak-PNJ periods. The temperature differences have the opposite sign equatorward of 60°N but are much weaker (about 3 degC at 30 km) since the absolute anomalies in this region are considerably smaller.

In the previous section it was shown how the wave forcing in the winter stratosphere causes deceleration of the PNJ and the eventual occurrence of a minor warming, after several 'preconditioning' pulses of wave activity from the troposphere. The extent of these EOF patterns into the southern hemisphere is a matter of some interest and an attempt is made to interpret it here.
Figure 12. The 100-day mean $\dot{z}$ (m s$^{-1}$) fields for the days when $\psi^0$ had (a) its largest values and (b) its smallest values (contour interval 10 m s$^{-1}$). (c) The difference ($a-b$) (contour interval 10 m s$^{-1}$) and negative differences are shaded. (d) The corresponding temperature difference (deg C) (contour interval 5 deg C) with the inclusion of the $\pm 2$ deg C contours (dotted); negative differences are shaded.
The thermal budget is well approximated (Andrews et al. 1987) by:

\[ \frac{\partial \tilde{\theta}}{\partial t} = \frac{Q}{Q} - \tilde{v} \frac{\partial \tilde{\theta}}{\partial \phi} - \tilde{w} \frac{\partial \tilde{\theta}}{\partial z} \]

where \( Q \) is the diabatic heating rate, arising from radiation transfer in the stratosphere. The thermal stratification of the stratosphere means that the most likely cause of the temperature anomalies in the lower equatorial stratosphere is vertical advection by the mean meridional circulation, coupled with changes in the radiative heating rate. Unfortunately, the heating rates were not saved during the model integration, so a complete study of the thermal budget cannot be performed. Further complications arise from the rather noisy structure of \( \tilde{w} \), whose time series at 10 hPa at the equator (Fig. 13(a)) shows rather large variations on all time-scales. Even smoothing this time series with a 29-day running mean (Fig. 13(b)) does not enable the straightforward identification of features which correlate well with \( \phi \) (Fig. 10(a)). Despite this noisy structure, the 100-day composites of the forcing in Eq. (7) are broadly consistent with expectations. The 100-day means of the vertical advection of potential temperature in the tropical lower and middle stratosphere generally cause less cooling in the fast-PNJ periods than in the slow-PNJ days (Fig. 14). However, directly at the equator and near 20°S this is not the case, the difference between the two cases being slightly negative. There is thus tentative evidence that the temperature anomalies in the tropical lower and middle stratosphere are related to changes in vertical advection due to the induced m.m.c.

![Figure 13. Time-series at 10 hPa at the equator of \( w^* \) (mm s^{-1}). (a) Raw data and (b) the 29-day running mean.](image-url)

6. THE GROWTH AND DECAY OF DISTURBANCES

(a) Time development and teleconnectivity

Attention is now devoted to the temporal development of the main features of \( \psi^* \). Since the time development of \( \bar{u} \) (62°N, 1 hPa) describes the long-term evolution of \( \psi^* \), this location is used as a reference point for calculations of the linear point correlation with \( \bar{u} \) at each latitude and longitude.
Leading and lagged correlations were calculated for time delays $\tau$ of $-29 \leq \tau \leq 29$ days. The latitude versus time-delay sections at 1 hPa and 5 hPa are shown in Fig. 15. The values at zero lag are comparable with those of $\psi_1^q$, with strong correlations near 60°N, regions of weak negative correlation in the northern hemisphere subtropics, and further positively correlated velocities in the southern hemisphere. The time evolution shows that the correlation at 1 hPa tends to grow and decay relatively rapidly (the 50% correlation is reached 15 days before and after the maximum), and that the signal tends to propagate from slightly higher latitudes (near 70°N). The region of positive correlations centred near 39°S increases quite slowly with the main centre and decays much more rapidly afterwards; this correlation is more marked at 5 hPa, where it is located closer to the equator. At 10 hPa (not shown) the correlation reaches 56% at 3–4 days lag.

Of considerable interest is the delay in the growth of the anticorrelated centre of $\psi_1^q$ located near 28°N; this reaches its maximum 15 days later than the main disturbance. At 5 hPa this anticorrelation peaks slightly later (at 20-days lag), lies slightly closer to the equator (consistent with the structure of $\psi_1^q$), and is slightly more persistent. The region of anticorrelation at 1 hPa means that a region of decreasing westerlies at 62°N is accompanied by increasing westerlies at 28°N (as shown in Fig. 12(c)) and that this subtropical increase becomes a maximum 15–20 days later. This time delay in the development of this anticorrelation centre is investigated further here.

(b) Orthogonal modes of temporal variability

Figure 15 shows the lagged correlations using unfiltered velocity fields, which retain all temporal scales. Since the longest periods are of greatest interest, some form of time smoothing is appropriate. There are various possible approaches to this problem. One of these is to apply low-pass filters to the data; this has the disadvantage of pre-specifying
the periodicities of the velocity fluctuations, which is undesirable given the relatively large range of time-scales evident in Fig. 10. The method of singular-spectrum analysis (SSA), which places fewer \textit{a priori} assumptions on the time-scales to be isolated, has been used to investigate the low-frequency variability of $\bar{u}$ in several different locations. In this study, the fluctuations in the filtered time series obtained by applying SSA are extremely similar to those obtained by simple techniques, such as running means. However, the advantage of such analysis is that the time series represent the evolution of temporally orthogonal patterns, in the same manner as the EOFs describe the spatial fields.

SSA proceeds by applying a window of length $W + 1$ to the data set and performing an eigenvalue analysis of the patterns of time development. The eigenvector expansion of a single-point time series ($u_p$) whose long-term mean is $U_p$ is:

$$ u_p(t) = U_p + \sum_{w=0}^{W} s_w^* (t) \chi_w^* \quad t = 1, \ldots, T - W + 1. \quad (8) $$

The eigenvectors $\chi_w^*$ have length $W + 1$ and $s_w^* (t)$ are the time-dependent amplitudes, analogous to the $a_\ell$ in Eq. (6). Again, more details of such analysis techniques are given in Chatfield and Collins (1980).

Analysis of $\bar{u}$ ($62^\circ N$, 1 hPa) first required the choice of an appropriate window. Experimentation with several values led to the choice of $W = 29$ as an optimum value, since this removes the high-frequency (defined here as several days) variability but is short enough to retain the low-frequency variations of the flow. Longer windows (e.g. $W = 69$) lead to more smoothing than necessary for the present study, while shorter windows do not adequately separate the different time-scales of variability. The first eigenvector with $W = 29$ accounts for 62.5\% of the variance of the time series and is characterized by long-period variations (Fig. 16(a), thick line) which describe the low-frequency variability of $a^*_1$ (Fig. 10(a)), as is also confirmed by power-spectrum analysis (not shown).
Figure 16. (a) The time series (days 201–1200) of the first principal components of singular spectrum analysis of $u$ at: 62°N, 1 hPa; 28°N, 1 hPa; and 22°S, 5 hPa. (b) The lagged correlation between $\psi_1^\circ$ (62°N, 1 hPa) and $s_1^\circ$ (28°N, 1 hPa). (c) The lagged correlation between $s_1^\circ$ (62°N, 1 hPa) and $s_1^\circ$ (22°S, 1 hPa).

(c) Correlations between the orthogonal modes

SSA has also been applied to the velocity time series at the other two ‘centres of action’ of $\psi_1^\circ$: 28°N, 1 hPa, and 22°S, 5 hPa. Their leading PCs (Fig. 16(a)) show the dominance of low-frequency variations with periods greater than about 50 days. The lagged correlations of the latter two points with $s_1^\circ$ (62°N, 1 hPa) (Figs. 16(b) and (c)) broadly agree with the correlations in Fig. 15, but there are important quantitative differences in the time evolution and strength of the correlations. The maximum correlation in the southern hemisphere (Fig. 16(c)) reaches 0.68 at zero lag, in comparison with the unfiltered value of 0.53. The anticorrelation in the northern subtropics (Fig. 16(b)) reaches −0.87 at 21 days lag, compared with 0.5 in the unfiltered data; it also reaches −0.2 at 13 days lead and is −0.53 at no lag when the anticorrelation of the unfiltered time series barely exceeds −0.2. The use of orthogonal representations of the low-frequency variability thus isolates rather larger correlations than were obtained without any time filtering, even when small-scale, localized features are included in the time series.

The cause of this anticorrelation has proven difficult to isolate, just as the structure of $\psi_1^\circ$ in the lower tropical stratosphere was impossible to explain with the stored model data. It is tempting to attribute these changes to the m.m.c., which is noisy. SSA failed to isolate significant features owing to the predominance of large oscillations on timescales of several days. A 49-day running mean was applied to $\vec{u}^*$, which successfully
isolates long-period oscillations (Fig. 17). These troughs and peaks can sometimes be related to the occurrence of changes in $\vec{u}$ at this latitude, but the correspondence is not perfect. The search for the cause of this time delay in the onset of the negative correlation thus remains inconclusive. A new version of the Berlin TSM GCM has now been implemented and several integrations performed, from which more data have been saved in a more suitable form for further analysis than in the current study; this topic will, therefore, be investigated in the future with the new model.

Figure 17. Time-series of the 49-day running mean of $\vec{v}^*$ (28°N, 1 hPa).

7. DISCUSSION AND CONCLUSIONS

In this study evidence is presented that the dominant variability of the wintertime stratosphere in a perpetual January integration of a GCM with fixed boundary conditions is represented by a spatial pattern of anticorrelation between the polar region and the wintertime extratropics. Weak anticorrelation between these regions extends down to the surface. The pattern of variability extends into the subtropics of the summer hemisphere, which is correlated with the winter polar region. Analysis of the temperature fields reveals that the anomalies are close to thermal-wind balance. The temporal behaviour of the stratospheric variability pattern is dominated by low-frequency variations. This result bears some resemblance to the 'stratospheric vacillation cycles' isolated in a simple 1-D quasi-geostrophic $\beta$-plane model of wave–mean-flow interaction discovered by Holton and Mass (1976) and analysed by Yoden (1987). The basic generating mechanism of wave forcing from the troposphere, leading to deceleration of the polar-night jet followed by recovery to westerlies and subsequent repetition, is fundamentally similar to that of the Holton–Mass model. The major difference with the GCM is that the full global circulation is simulated, which allows other forcing to act; in particular, the mean meridional advection tends to counteract the wave forcing. It is significant that the vacillatory behaviour of the Holton–Mass model extends to a primitive-equation GCM incorporating realistic physical parametrizations.

One fundamental difference between the GCM and the Holton–Mass model is the transient nature of the forcing from the troposphere in the GCM. This is in accord with observations, illustrated by Shiotani and Hirota (1985) as an episodic forcing for the lower boundary of the stratosphere. The forcing in the GCM occurs with a periodicity of about 20–100 days and is difficult to distinguish from a random process. The GCM only simulates minor warmings; the zonal wind at 60°N, 10 hPa is never easterly, as required for a major warming. Nevertheless, the model behaviour—a low-frequency response to medium-frequency forcing—is representative of the atmosphere. Some benefit may result from the further use of simpler stratospheric models, such as the 1-D
Holton–Mass model or, more appropriately, a 3-D primitive-equation model, with time-dependent lower-boundary forcing to examine the sensitivity of the stratosphere to the strength and transience of the energy input from the troposphere.

The time development of the dominant pattern ($\psi(t)$) of spatial variability closely follows the longer-term oscillations in $\bar{u}$ at 62°N, 1 hPa. This is reminiscent of the results of Wallace and Chang (1982), who found that the stratospheric wintertime flow could be characterized by any single parameter defining the broad-scale features of the circulation. That this ‘low dimensionality’ of the stratosphere extends to high levels, at least in this idealized GCM experiment, is also a significant result. This in itself poses an important problem. The stratosphere is believed to be influenced by several mechanisms, which either directly change the forcing (e.g. the troposphere) or influence the manner in which it can react to any given forcing (e.g. the QBO). If the stratospheric response to any of these is simply a weakening and/or shifting of the PNJ it may be impossible to determine the causality of interannual variability in the real stratosphere by data analysis alone, particularly with the relatively short data time series currently available.

The perpetual January integration displays a long-period oscillation which would act over seasonal time-scales in an atmosphere with an annual cycle. It is, therefore, essential to perform similar analyses for long integrations of a TSM GCM, including the annual cycle but with no interannual variability in the boundary conditions; such an analysis is currently being performed on a 35-year integration of the Berlin GCM. The problem for the stratosphere is essentially whether any anomalies forced in one year can survive the summer (when the forcing is turned off) and influence the circulation the following autumn. This study has isolated significant correlations at lags of up to about 25 days in the GCM. Kodera et al. (1990, 1991) have shown how the state of the subtropical stratopause in early winter can affect higher latitudes and lower levels in later months. Such a memory may only arise from the chemical composition of the atmosphere which affects the radiative properties, since the small heat capacity of the middle atmosphere precludes any interannual variability in runs of a simple model (Yoden 1990). This requires the use of models with interactive chemistry. Relevant trace gases include low stratospheric ozone, which has a moderately long lifetime and a large effect on the radiation budget.

The current study has shown that the low-frequency oscillations found in simple models carry over to more complex GCMs. The moderately long integration allowed some statistical analysis, whose significance requires further investigation. Much longer integrations are also required to examine statistically the coupling between the troposphere and stratosphere. Simplified models, such as that employed by James and James (1992) but extending into the mesosphere, could be applied to such problems, with the assumption that their inherent simplifications are acceptable. However, the use of GCMs allows most of the feedbacks between radiation and dynamics to be accounted for, and the ultimate studies of the forced and unforced variability of the atmosphere must take all aspects of the problem into consideration. In the long term, there is no alternative to multi-year integrations with state-of-the-art GCMs.

**Acknowledgements**

We are grateful to our colleagues K. Labitzke, U. Langematz and P. Strauch for their encouragement and assistance with various aspects of this work. The EOF and SSA routines were provided by R.-S. Wang, who also provided invaluable advice on their application. Discussions with K. Fraedrich, K. Kodera, and S. Yoden motivated the continuation of the analysis. Suggestions by A. Beck and J. Taubenheim on an early
version of the manuscript led to improvements in the relationship to other results. The anonymous referees and the associate editor, J. Slingo, provided constructive comments which led to some more detailed explanations of the methods used and the interpretation of the results being included during revision. The GCM integration and some diagnostics were performed on the CRAY of the Konrad-Zuse-Zentrum für Informationstechnik Berlin. The research was partly supported by the climate research program of the German Federal Ministry for Research and Technology (BMFT), grant 07KFT306.

REFERENCES


Kodera, K., Yamazaki, K., Chiba, M. and Shibata, K.

Kodera, K., Chiba, M., Yamazaki, K. and Shibata, K.

Kutzbach, J. E.

Labitzke, K.

Labitzke, K. and van Loon, H.

Matsuno, T.

Palmer, T. N.

Palmer, T. N., Shutts, G. J. and Swinbank, R.

Pawson, S.

Pawson, S. and Shine, K. P.


Rind, D., Suozzo, R., Balachandran, N. K., Lacis, A. and Russell, G.

Shine, K. P.

Shine, K. P. and Rickaby, J. A.

Shiotani, M. and Hirota, I.

Simmons, A. J. and Strüfing, R.

Strobel, D. F.

Taubenheim, J., Entzian, G. and von Cossart, G.


