The Berlin troposphere–stratosphere–mesosphere GCM: Sensitivity to physical parametrizations

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SUMMARY

The sensitivity of a troposphere–stratosphere–mesosphere general-circulation model to changes in the radiation transfer and subgrid-scale drag parametrizations is investigated. The large interannual variability of the middle atmospheric circulation necessitated a methodological approach. A set of 1260-day, perpetual-January integrations was performed; this enabled significant signals to be extracted from the model variability at a reasonable cost. A Rayleigh-friction scheme was used for the mesospheric drag. Reducing its strength led to changes in the zonal-mean structure of the model which resembled the leading mode of variability. Increasing its depth led to significant changes throughout the stratosphere and extending into the troposphere. Including the diurnal cycle induced no significant changes in the zonal-mean state below the stratopause. Radiation-transfer calculations are not performed fully every time-step; decreasing the frequency of computation from 2 to 12 hours caused marked changes in the mean structure and variability. Including a topographic gravity-wave drag parametrization (TGWD) restored the zonal-mean structure of the 2-hour radiation frequency in the lower stratosphere at the expense of changes in both the temporal variability and the planetary-wave structures. The sensitivity to the frequency of radiation calculations arose from the highly coupled nature of tropical cloud–radiation interactions: the modified structure of the upper tropospheric divergence led to changes in the Rossby-wave source term in the extratropics. The major conclusions are: (1) the natural variability must be properly included in interpretations of the model sensitivity; (2) the adequacy of the tropospheric simulation can profoundly affect the stratosphere; (3) changes in the mesospheric drag can modify the tropospheric circulation; and (4) deficiencies in one parametrization (in this case radiation transfer) can be compensated by other changes (TGWD), but such ‘improvements’ may not apply to all aspects of the simulation.

KEYWORDS: Gravity-wave drag Middle atmosphere Radiation Rayleigh friction Troposphere–stratosphere interaction

1. INTRODUCTION

The continuing interest in the effect of stratospheric trace-gas changes on the environment and human health justifies studies of the climatology of the terrestrial middle atmosphere. General circulation models (GCMs) are generally used for this, since they include state-of-the-art parametrizations of physical processes and radiative–dynamical links between the stratosphere and troposphere. Some aspects of comprehensive climate simulations with troposphere–stratosphere–mesosphere (TSM) GCMs are discussed in section 2.

When examining the possibility of climatic change, it is essential to understand the performance of the GCM being used. As well as documenting the climatological structure and transient behaviour, the sensitivity to the assumptions in the physical parametrizations also needs to be studied. That is the intention of this study: some aspects of the physical parametrizations of the Berlin TSM GCM which can influence the middle atmosphere are investigated. Langematz and Pawson (1997) discuss the climatology of this model. The processes examined here are the calculation of radiative-heating rates, including the importance of the diurnal cycle, and small-scale drag (due to unresolved gravity waves). It turns out that the largest response in the middle atmospheric circulation arises from degrading the calculation of radiative-heating rates and it is argued that this is not a direct consequence of the middle atmosphere—rather, it comes from interactions between convection and radiation in the tropical troposphere.

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The model is described in section 3. An attempt has been made to construct a framework in which significant changes in the simulation can be isolated from the large natural variability; this, along with some details of the reference integration, is described in section 4. The results of the sensitivity experiments are then presented. First (section 5) the consequences of changing the mesospheric drag are shown. The dependence of the model performance on the accuracy of the radiation-transfer calculations is discussed, along with an effect of topographic gravity-wave drag (TGWD) in the lower stratosphere (section 6). In section 7 the influence of changes in the tropical troposphere on high latitudes and the stratosphere is interpreted. The results and conclusions are then summarized.

2. A brief review of some previous GCM studies

Most current TSM GCMs are able to simulate the basic structure of the middle atmosphere but some deficiencies remain. One of these is the 'cold Pole' problem: the modelled temperature of the lower polar stratosphere in winter is much too low. This can be alleviated, at least in the northern hemisphere, by the introduction of TGWD schemes, which provide additional easterly forcing to the low stratosphere, closing the subtropical jets and leading to an increased meridional heat transport (e.g. Palmer et al. 1986; McFarlane 1987; Rind et al. 1988a). Alternatively, Ramanathan et al. (1983) showed that improving the treatment of radiative processes could lead to a better thermal structure in the lower stratosphere; these results need to be interpreted cautiously, since their model had an upper boundary at 10 hPa and the integration was extremely short in comparison to the variability inherent in the stratosphere—even in numerical models (e.g. Boville 1986). Other possibilities include increasing the horizontal resolution: Mahlman and Umscheid (1987) and Hamilton et al. (1995) found that the performance of the GFDL* SKYHI model improved with increasing resolution, presumably due to the ability of the model to simulate a more realistic gravity-wave spectrum at higher resolution (e.g. Hayashi et al. 1989). Studies with the National Center for Atmospheric Research CCM2 model (Boville 1995) also showed a slight reduction in the strength of the polar-night jet (PNJ) as the resolution increased, although only one winter season was analysed at the higher resolutions (T63 and T106).

The importance of parametrized gravity waves for the performance of the Goddard Institute for Space Studies model was emphasised by Rind et al. (1988a,b), who found considerable improvements to the transient and mean behaviour when small-scale drag was included. They emphasised the role of travelling gravity waves, such as may be forced by convection and shear instabilities. Even though more recent parametrizations of such waves (e.g. Hines 1997) are considerably more detailed, the difficulties of coupling the wave spectra to the sources remain, not least because the generation mechanisms are also parametrized. In the current study a simple, linear (Rayleigh) friction (Holton and Wehrbein 1980) was applied in the mesosphere. One objective here is to assess the model's sensitivity to this drag before more sophisticated parametrizations are tested in future studies. The relatively simple treatment of mesospheric drag of Rind et al. (1988a) considerably improved the structure of the mesospheric PNJ, which often lies too far polewards in TSM GCMs. The first theme of the current study is an examination of the sensitivity of the model to the mesospheric drag, which is imposed in a simple manner.

The original version of the Berlin TSM GCM (Pawson et al. 1991) had several deficiencies, including a cold Pole and an unrealistic PNJ. One of the major aspects requiring attention was the tropospheric radiation scheme (Geley and Hollingsworth 1979) which McRostie (1991) found to be inaccurate in the upper troposphere and lower stratosphere.

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Pawson (1992) suggested that the cold bias in the tropical low stratosphere of the Berlin TSM GCM arose from inadequacies in this radiation scheme. The new model includes a
new radiation scheme (Morcrette 1991) as well as many other modifications. The second
dominant theme of this study is the response of the TSM GCM to changes in the radiation
calculations.

3. Model description

The Berlin TSM GCM was modified from the ECHAM1 climate model (Rochecker
et al. 1992). The spectral code from the European Centre for Medium-Range Weather
Forecasts (ECMWF) is used with a triangular truncation; all results presented here were
performed at T21 horizontal resolution. The Gaussian grid, on which the nonlinear
dynamics and the parametrized tendencies are calculated, has 32 latitudes and 64 longitudes
and a resolution of roughly 5.6°.

The hybrid vertical coordinate (Simmons and Strüfing 1983) is used. It is defined at
model half-levels by:

\[ \eta_{k+1/2} = \frac{a_{k+1/2}}{p_s} + b_{k+1/2} , \quad \text{with } k = 0, 1, ..., n_k , \]  

(1)

where \( p_s \) is the surface pressure, \( n_k \) is the number of levels, and the \( a_k \) and \( b_k \) are parameters
which define the vertical grid. Pure sigma coordinates are used in the lower troposphere
\( (a_{k+1/2} = 0) \) with a gradual transition to pressure levels \( (b_{k+1/2} = 0) \) in the stratosphere. The
full-level pressures are determined by linear interpolation between the half-levels, using
the local surface pressure. The pressure-gradient correction (Simmons and Chen 1991) is
included to improve the model performance in regions of steep terrain.

As in Pawson et al. (1991), the model was extended vertically: the upper boundary
was raised from 10 hPa in ECHAM1 to 0.0068 hPa. The 35 half-levels of the model are
tabulated in Table 1. The full-level pressures, assuming \( p_s = 1013.25 \) hPa, and the pressure
scale heights, \( z = H \ln(p/p_s) \), calculated assuming a mean scale height \( H = 7 \) km for an
isothermal atmosphere at 240 K (which is the height coordinate used in the plots), are also
given. The approximate vertical resolution is also shown: this restricts the vertical scale of
waves which can be represented.

A semi-implicit time-stepping scheme (Robert 1982) is used. Tests revealed that a
time-step of 15 minutes was adequate for the simulations: this is considerably shorter
than the 40-minute time-step used in ECHAM1 (Rochecker et al. 1992) in order to maintain
numerical stability in the stratosphere and mesosphere, where higher velocities are reached,
through the Courant–Friedrichs–Lewy criterion.

Horizontal diffusion is represented with the ‘spectral chopping’ scheme of Laursen
and Eliassen (1989), which damps the higher wave numbers \( n > 15 \) by a factor proportional
to \( (n - 15)^2 \). The damping time-scales adopted are 1.12 days for vorticity, 0.22 days for
the divergence and 5.59 days for temperature, humidity and cloud liquid water. These
coefficients are constant with height throughout the troposphere and stratosphere but are
increased at the top five model levels (i.e. at \( p < 0.1 \) hPa); they are successively doubled
at levels 5–2 and the 16-fold increase from the standard value is also used at the highest
level.

The tropospheric physics, which are mainly based on ECHAM1, are summarized in
Table 2 and will not be described in detail here. The major difference between this model
and ECHAM1 is that the radiation scheme of Morcrette (1991) is used at the lowest levels
of the TSM GCM. Long-wave cooling is approximated by a cooling-to-space formulation
### Table 1. The coefficients defining the 35 half-levels of the model, the full-level pressures, assuming a surface pressure of 1013.25 hPa, the pressure scale heights of these full-levels, and the height increments.

<table>
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<th>$a_{k+\frac{1}{2}}$ (Pa)</th>
<th>$b_{k+\frac{1}{2}}$ (hPa)</th>
<th>$p_k$ (hPa)</th>
<th>$z_k$ (km)</th>
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See text for definition of symbols.

for $p < 10$ hPa. In the Morcrette scheme, long-wave radiative fluxes are calculated using a broad-band flux emissivity method, with six spectral intervals, for CO₂, H₂O, O₃, and aerosols. Additionally, the scattering and absorption by clouds are included.

In the mesosphere the solar heating by O₂ at wavelengths between 125 and 205 nm is calculated with the model of Strobel (1978). At $p < 70$ hPa the parametrization of Shine and Rickaby (1989) is used to determine the solar heating rates by O₃ and O₂ in four spectral intervals from 206.186–852.500 nm. At 70 hPa, the fluxes from these calculations are merged with the appropriate spectral interval of the solar heating model of Fouquart and Bonnel (1980), a two-stream formulation with photon path distribution method. In turn, the backscattered fluxes from this model are used as a lower-boundary condition for the Shine and Rickaby scheme. Solar infrared heating is calculated with the Fouquart and
<table>
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<th>Quantity</th>
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<td>Sea surface temperature</td>
<td>Specified</td>
<td>Reynolds (1988)</td>
</tr>
<tr>
<td>Sea-ice extent</td>
<td>Specified</td>
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<tr>
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<td>Envelope orography</td>
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<td>Horizontal diffusion</td>
<td>Scale selective</td>
<td>Laursen and Eliaisen (1989)</td>
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<td>Drag coefficients</td>
<td>Louis (1979), Miller et al. (1992)</td>
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<td>Surface exchange</td>
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<td>Tiedtke et al. (1988)</td>
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<td>Stratiform cloud</td>
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<td>Absorption fit</td>
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<td>Rayleigh friction</td>
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Bonnel (1980) parametrization. Scattering and gaseous absorption due to H₂O, O₃, and uniformly mixed gases are taken into account.

Radiative fluxes at model half-levels are not calculated at every time-step because of the extreme computational expense involved. They are calculated at all model grid points at the so-called ‘full-radiation steps’, which are variously 2 or 12 hours in the current study; at intermediate time-steps the heating rates are calculated using the local temperatures and the ‘effective emissivity/transmissivity’ approach for the long-wave/short-wave calculations, discussed by Geleyn and Hollingsworth (1979). This means that the cloud and water-vapour distributions used in the radiative-heating rate calculations are fixed between the full-radiation time-steps (and are thus inconsistent with the latent heating, which is recalculated at each time-step). The solar zenith angle is updated at every time-step for the heating-rate calculations.

The prognostic scheme of Roeckner and Schlese (1985) is used to determine the stratiform cloud distributions. Convective clouds are represented by the schemes described in Tiedtke et al. (1988); they showed the importance of shallow convection in maintaining the moisture budget for the deep convective zones; this deep convection is represented using a modification of the Kuo (1974) scheme, where the moistening parameter and the cloud-base definition are revised (Tiedtke et al. 1988). The cloud optical properties are those used in Morcrette (1991).
The modelled H_2O and cloud distributions are used, but concentrations of the remaining gases are prescribed; O_2 (20.95\%) and CO_2 (330 parts per million by volume) are assumed to be well mixed. A zonal and monthly averaged O_3 climatology is specified; a smooth annual cycle is fitted to these climatological values. For perpetual-season integrations, the appropriate mid-month values are used. At p < 141 hPa the data from Keating et al. (1990) are used. In the lower stratosphere they are supplemented using data from the Solar Backscattered Ultraviolet radiometer (McPeters et al. 1984), and extended into the polar night using an algorithm provided by K.P. Shine (personal communication, 1989). At higher pressures, zonal-mean values from the ECMWF model were used (as in Pawson et al. (1991)).

Two fairly simple representations of gravity-wave drag were used in this study. The first, intended to represent the effects of dissipating gravity waves in the mesosphere, is a linear relaxation (Rayleigh friction). It is discussed in more detail in section 5, when the sensitivity of the modelled climatology to this process is discussed. The second, used in one of the experiments discussed, is TGWD parametrization designed for the lower stratosphere by Palmer et al. (1986).

4. EXPERIMENTAL DESIGN AND THE REFERENCE INTEGRATION

One of the major problems associated with determining the sensitivity of a model to changes in the physical parametrizations is the separation of changes from the natural variability. This is particularly pronounced in the stratosphere, which displays considerable variability from year to year (e.g. Labitzke 1982). The interannual variability exhibited by comprehensive climate models (e.g. Rind et al. 1988a; Hamilton 1995a) means that long integrations of at least a decade must be performed in order to obtain stable climatologies from TSM GCMs. This is computationally unreasonable if several integrations are to be performed. The sensitivity studies presented here were thus performed with perpetual-January integrations. The temporal variability (e.g. Pawson et al. 1995) means that these must be long enough to enable a clear separation of the signal of the change in physics from the internal variability. Results from 1080 days (36 30-day months) have been analysed, after allowing adequate time (180 days) for the spin-up of the model.

The experiments performed (see Table 3) have the following naming convention: PJ (perpetual January); DC or DM (for diurnal cycle or diurnal mean); TT = 02 or 12 defines the radiation frequency; further appendages refer to the drag: RD and RW describe deep and weak Rayleigh friction and TG denotes the use of TGWD. The first three experiments (PJDC02, PJDC02RW, PJDC02RD) differ only in the specification of the Rayleigh friction;

<table>
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<th>Name</th>
<th>Annual cycle</th>
<th>Diurnal cycle</th>
<th>Full-radiation frequency</th>
<th>Rayleigh friction</th>
<th>Gravity-wave drag</th>
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<td>Off</td>
<td>12 h</td>
<td>normal</td>
<td>off</td>
<td>1260 d</td>
<td>6 h</td>
</tr>
<tr>
<td>PJDM12TG</td>
<td>Off</td>
<td>Off</td>
<td>12 h</td>
<td>normal</td>
<td>on</td>
<td>1260 d</td>
<td>6 h</td>
</tr>
</tbody>
</table>

The names are composed in the form: PJ (perpetual January); DC/DM (diurnal cycle or diurnal mean solar forcing); a two-digit integer representing the frequency of full-radiation steps (in hours). The additional characters TG designate the use of the Palmer et al. (1986) topographic gravity-wave drag scheme while RW and RD denote the weakened or deepened Rayleigh friction.
the next two runs (PJDM02, and PJDM12) calculate the radiative-heating rates with a different accuracy; PJDM12TG includes TGWD (Palmer et al. 1986), so that $G$ in Eq. (2) is not zero.

The climatological structure of the reference integration, PJDC02, is summarized. The temporal variability is similar to that of the previous version of the model (Pawson et al. 1995); the current TSM GCM undergoes major (as opposed to minor) stratospheric warmings (Erlebach et al. 1996). Eight major warmings occurred in the 1080-day period examined here. As in Pawson et al. (1995), more rapid variations were superimposed on this predominant low-frequency variability.

The 1080-day mean of the zonally averaged zonal wind, $\overline{\bar{u}}$ (Fig. 1a), is more realistic than in Pawson et al. (1995). The most obvious improvement is the clear separation of the subtropical jet (STJ) from the PNJ in the winter hemisphere. However, the PNJ is located too far polewards and $\overline{\bar{u}}$ is too weak in the winter stratosphere. The summer stratosphere is dominated by easterlies, with a jet of $-70$ m s$^{-1}$ centred just below the stratopause, but weak westerlies prevail up to 45 km south of 60°S. The stratopause temperature is quite realistic.

After Andrews et al. (1987), in spherical, log-pressure coordinates $(\lambda, \phi, z)$ on a sphere with radius $r$, the zonal-mean momentum equation is:

$$
\frac{\partial \bar{u}}{\partial t} + \frac{\bar{u}^*}{r} \frac{\partial \bar{u}}{\partial \phi} + \bar{w}^* \frac{\partial \bar{u}}{\partial z} - \left( f - \frac{\bar{u} \tan \phi}{r} \right) \overline{\bar{v}^*} = D + X + G,
$$

(2)

where $\bar{v}^*$ and $\overline{\bar{u}^*}$ describe the transformed Eulerian mean meridional circulation, $f$ is the Coriolis parameter, $D$ is the forcing due to resolved eddy motions, $G$ is the TGWD (which is identically zero in most of the integrations discussed here, including PJDC02) and $X$
represents the effects of mean-flow forcing due to the other parametrized processes in the GCM (and unresolved processes in the diagnostics). The dominant components of the time-mean forcing are \([D]\) (Fig. 1b) and \([f \mathbf{v}^*]\), which partially cancel (cf. Shiotani and Horiya 1985). In the extratropical troposphere the baroclinic eddies can clearly be seen in the structure of \([D]\), with positive forcing in the lower troposphere and negative values at higher levels (Edmon et al. 1980). In the stratosphere and mesosphere the easterly forcing by planetary waves dominates \([D]\), with a magnitude of more than 10 m s\(^{-1}\) near 60 km. There is westerly forcing in the summer mesosphere.

In summary, the zonal-mean structure of the new TSM GCM (run PJDC02) performs considerably better than the previous version (Pawson et al. 1991, 1995). Since the horizontal resolution is unchanged and the vertical resolution is similar in both versions, most of the stratospheric changes must arise from the different treatment of radiation transfer and the tropospheric parametrizations. More fields from this integration will be introduced in the following discussions.

5. THE INFLUENCE OF THE MESOSPHERIC DRAG

(a) Formulation of the Rayleigh friction

Small-scale mesospheric drag is represented by a Rayleigh-friction parametrization. Two experiments addressed the effects of the strength and structure of the drag profiles. Following Holton and Wehrbein (1980), the height-dependent drag coefficient is specified as:

\[
\gamma = \gamma_1 + \gamma_2 \times \left\{ 1 + \tanh \left( \frac{z - z_1}{z_2} \right) \right\},
\]

where \(\gamma_1\) and \(\gamma_2\) define the strength and \(z_1\) and \(z_2\) determine the depth of the dissipation. The horizontal drag is then: \((\widehat{X}, \widehat{Y}) = -\gamma \times (u, v)\). It affects \(D\) in Eq. (2) through its influence on the zonally asymmetric part of the flow field, as well as contributing directly to \(X\) (the zonal mean of \(\widehat{X}\)).

In PJDC02 the coefficients in Eq. (3) were chosen to give weak drag at the highest model levels (Table 4). In PJDC02RW, \(\gamma\) was weakened by 50%; in PJDC02RD, a deeper \(\gamma\) profile, extending into the upper stratosphere, was imposed.

(b) Response to weakening the drag

Weakening \(\gamma\) led to a velocity change, as shown by \(\Delta \mathbf{u}\) in Fig. 2. Both the northern hemisphere westerlies and the southern hemisphere easterlies in the mesosphere are stronger. These differences are quite small, just reaching 16 m s\(^{-1}\) near 80 km, 70°N, and the general pattern of \([\widehat{u}]\) is similar to that in PJDC02 (Fig. 1a). In the northern subtropics the westerlies in PJDC02RW are weaker than in PJDC02—the PNJ is displaced slightly to the north. In the absence of other changes in the forcing, a reduction in \(\gamma\) should simply

<table>
<thead>
<tr>
<th>Run</th>
<th>(\gamma_1) (d(^{-1}))</th>
<th>(\gamma_2) (d(^{-1}))</th>
<th>(z_1) (km)</th>
<th>(z_2) (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>PJDC02</td>
<td>0.0</td>
<td>0.25</td>
<td>78.0</td>
<td>6.3</td>
</tr>
<tr>
<td>PJDC02RW</td>
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<td>0.125</td>
<td>78.0</td>
<td>6.3</td>
</tr>
<tr>
<td>PJDC02RD</td>
<td>0.0</td>
<td>0.125</td>
<td>63.0</td>
<td>7.5</td>
</tr>
</tbody>
</table>

Experiment names as in Table 3. Values given for PJDC02 were used in all other integrations.
lead to stronger winds. This is the case in the southern hemisphere mesosphere and the northern high latitudes. In the northern hemisphere $\Delta[u]$ has a dipolar structure, resembling the structure of the leading mode of variability (Pawson et al. 1995); the differences may be attributed to the tendency of the model to favour the stronger PNJ state when $\gamma$ is weakened. The wind differences extend down to the lower stratosphere with a small, probably insignificant, effect in the troposphere.

In a trend-free system with $G = 0$, the time mean of Eq. (2) is approximately:

$$-f[\vec{v}^*] = [D] + [X] = [D] - \gamma[u]. \tag{4}$$

Thus, as long as $[D]$ were unaffected, the leading-order change in the $[\vec{v}^*]$ would be given by:

$$f \Delta[\vec{v}^*] = \gamma \Delta[u] + \Delta \gamma[u], \tag{5}$$

so that in regions where $\Delta \gamma = -0.5 \gamma$, the first term on the right of Eq. (5) is negligible because $\Delta[u] \ll 0.5[u]$ in PJDC02RW (Fig. 2). Thus, in the southern hemisphere mesosphere, where $[\vec{u}]$, $f$, and $\Delta \gamma$ are all negative, one expects $[\vec{v}^*]$ to decrease. Similarly, at high northern latitudes a reduction in the northern flow is anticipated. In fact, $\Delta[\vec{v}^*] < 0$ in the entire upper stratosphere and mesosphere (Fig. 3(a)), validating this very crude argument in those regions where it is applicable. In the northern subtropical mesosphere the neglected contributions all play an important role in the decrease in $[\vec{v}^*]$. This decrease in $[\vec{v}^*]$ throughout the mesosphere is consistent, through mass conservation, with weaker upwelling/downwelling $[\vec{w}^*]$ in the summer/winter polar regions (Fig. 3(b)). Changes in the mean meridional circulation become very small below the stratopause.

(c) **Response to deepening the drag**

Larger, more complex changes in $[\vec{u}]$ are evident for PJDC02RD (Fig. 4). The deeper drag acts down to about 45 km but $\Delta[\vec{u}]$ is positive at most latitudes in the stratosphere and mesosphere, with a negative band sloping upwards and polewards from the subtropical troposphere to about 60°N in the mesosphere. The structure of $\Delta[\vec{u}]$ shows PNJ changes which do not resemble the dominant mode of variability of PJDC02, indicating a fundamental change in the atmospheric circulation extending into the troposphere. The $\Delta[\vec{u}]$ between 20 and 50 km polewards of 60°N signifies a downward extension of the PNJ. In the mesosphere the PNJ extends further towards the subtropics, with a hint of a ‘double-peaked’ structure in PJDC02RD that is not evident in PJDC02 (Fig. 1); this agrees better
Figure 3. The differences in the meridional circulation for PJDC02RW minus PJDC02 (see text). (a) Δv* (contour interval 10^{-1} m s^{-1}) and (b) Δw* (contour interval 0.5 mm s^{-1}). See text for explanation of symbols. Negative values are shaded in each case. Note that these figures cover the height range 35–75 km.

Figure 4. (a) As Fig. 2 but for PJDC02RD minus PJDC02 (see text). (b) [u] for PJDC02RD (contour interval 10 m s^{-1}, with negative contours dashed). See text for explanation of symbols.
with observations (e.g., Fleming et al. 1990) where the the stratospheric PNJ and the mesospheric subtropical jet are separated.

The change in $\bar{u}$ is partially a direct response to the structure of the eddy forcing. At 1 hPa (near 48 km) the latitudinal structure of $[D]$ (Fig. 5) changes markedly polewards of about 40°N: the easterly forcing is more than 1.5 m s$^{-1}$ d$^{-1}$ stronger near 60°N, leading to a decrease in $\bar{u}$ equatorwards of 60°N. However, polewards of 60°N $\bar{u}$ is more westerly, showing the importance of the other forcing terms in Eq. (2). At 10 hPa (near 32 km) the generally positive $\Delta \bar{u}$ (Fig. 4) is a direct consequence of the weaker negative $[D]$ in PJDC02RD (Fig. 5).

There are also changes in $(\bar{u}^*, [\bar{u}^*])$ between PJDC02RD and PJDC02 (Fig. 6). The changes in $[D]$ mean that Eq. (5) is invalid in the northern winter. In the summer hemisphere of PJDC02RD, the stratospheric descent is weaker and the lower mesospheric (55–65 km) ascent is about 25% stronger. The stronger $[\bar{u}^*]$ in the mesosphere reflects the extended latitudinal range and smoother structure of this branch of the meridional circulation. In northern high latitudes the descent rate is reduced by about 30% above 70 km and increases by about 0.5 mm s$^{-1}$ (about 25%) down to 40 km polewards of 40°N, compensating for the increased depth of the northward flow. This would lead to an adiabatic temperature increase, but the direct eddy forcing of the lower stratosphere causes a substantial increase in $\partial \bar{u} / \partial z$ in the lower and middle stratosphere, leading to a stronger meridional temperature gradient, so the polar region is up to 13 deg C colder in PJDC02RD than in PJDC02 (Fig. 7). This brings the polar temperature much closer to the 30-year mean observed temperature in January. This should be compared to the 3 deg C difference between PJDC02 and PJDC02RW.
(d) Discussion

The time-averaged fields responded to the changes in the imposed mesospheric drag in different ways. The transient behaviour is not sensitive to the mesospheric drag: sudden warmings occur sporadically in all three runs, although their relative frequency may change. When $\gamma$ is weakened, the time-averaged changes are at least qualitatively consistent with the concept of downward control (Haynes et al. 1991). Weakening $\gamma$ in the mesosphere causes small changes in $[\vec{u}]$ throughout the stratosphere, and the mean meridional circulation differences are consistent with the temperature changes required to maintain thermal wind balance. However, the response of the stratospheric wind field is simply the dominant mode of natural variability of the TSM GCM which also maintains thermal wind balance, and includes changes in the mean meridional circulation (Pawson et al. 1995) which extend into the lower stratosphere. A causal mechanism of the changes in
the zonal-mean structure in PJDC02RW is thus suggested: in the mesosphere $\overline{u}$ responds directly to the change in $\gamma$ but this causes the model to favour a strong PNJ which is associated with a colder lower polar stratosphere. Thus, although the reponse is consistent with the concept of downward control, this is not a causal mechanism.

Planetary-wave propagation and forcing is an inherently non-local process; even in linear, small-amplitude theory (the basis of the diagnostics presented here although it is not assumed in the model) $D$ has a functional dependence on $\overline{u}(\phi, z)$ because of the propagation properties of planetary waves (e.g. Andrews et al. 1987; Matsuno 1970). More complete theories allow for weak or strong nonlinearities, facilitating wave reflection or resonance (e.g. Tung and Lindzen 1979) which can change the wave structures at lower levels according to the strength or the location of the PNJ. Such processes could be invoked to explain the changes in $[D]$ in the middle and lower stratosphere which occurred when $\gamma$ was deepened. This illustrates the importance of non-local effects, whereby the imposition of additional forcing at higher levels (in the mesosphere) can lead to changes in the wave structures throughout the middle atmosphere and ultimately in the troposphere, where clear changes in $[\overline{u}]$ are found. In this case the 'downward-control' effect is smaller than the local changes in eddy forcing.

The current TSM GCM is sensitive to the mesospheric drag, which was simply specified to facilitate a straightforward diagnostic study. Studies with the GFDL SKYHI model (Hayashi et al. 1989) showed that resolved gravity waves can directly influence the circulation. Hamilton (1995b) found that imposing additional drag in the southern mesosphere in wintertime has important consequences for the success of the simulations. Clearly, studies with realistic parametrizations of travelling, transient gravity waves are required: although Hines's (1997) scheme may go a long way towards resolving the propagation and dissipation properties of gravity waves, much effort is still required to quantify the strength and locations of their generation, despite the progress in recent modelling studies (e.g. Alexander et al. 1995). Rayleigh friction may not be physically correct but it provides a controlled framework in which to examine the model response to mesospheric drag, in the hope that future studies with a more comprehensive drag parametrization can be better directed.


(a) The experiments compared

The effects of changing the frequency of full-radiation calculations are now investigated by comparing PJDC02 with PJDM02 and PJDM12, where the radiative-heating rates were calculated with a different accuracy (Table 3). PJDC02 included the diurnal cycle and updated the radiative fluxes every two hours. PJDM02 was identical, except that diurnal mean solar insolation was used. In PJDM12 the radiative fluxes were only calculated every 12 hours. For reasons which will become apparent during the discussion, a further experiment (PJDM12TG), which includes a TGWD parametrization (so that $G$ in Eq. (2) is not zero), is also included here.

(b) Time dependence in the integrations

The different ranges of the 10 hPa zonal-mean zonal velocity at 58°N, $\overline{u}_{58}$, and the North Pole temperature $T_{\text{NP}}$ in the four integrations are illustrated by their binned distributions (Fig. 8). In runs PJDC02 and PJDM02 these distributions are very similar, concentrated close to $\overline{u}_{58} = 20 \text{ m s}^{-1}$ and $T_{\text{NP}} = -50 ^\circ \text{C}$ with limited excursions from this modal point. In contrast, in PJDM12 the distribution is more dispersed but generally slower
and warmer. Introducing the TG WD overcorrects this bias, causing a clustering of values about their mode which has a higher $T_{sp}$ and lower $\bar{u}_{58}$ than in runs PJDC02 and PJDM02.

(c) Stationary-wave structure

The 10 hPa geopotential heights ($Z_{10}$) reveal systematic differences between the four runs (Fig. 9). In PJDC02 the polar vortex is displaced from the Pole and centred over Scandinavia, with a well-formed Aleutian high; the vortex is quite elongated. A similar pattern is found for PJDM02, where both the Aleutian high and the cyclonic centre are weaker. Decreasing the radiation frequency to 12 hours has a large effect on $Z_{10}$: in PJDM12 it is dominated by a wave-number-1 structure, the Aleutian high is 64 geopotential decametres (gpdm) higher and the low is 64 gpdm deeper than in PJDC02; the orientation of the vortex also changes, the dominant features moving further east (by about $60^\circ$ at $40^\circ$N). In PJDM12TG the general shape and location of the vortex resemble those in PJDC02 and PJDM02 rather than those in PJDM12, although the field is slightly flatter and the cyclonic centre is moved to the west.

(d) The effect of tidal forcing

Neither the transient behaviour of the zonal-mean flow nor the time-averaged eddy structure change significantly when the diurnal cycle is turned off. The latitudinal structure
of \([\bar{u}]\) at 1 hPa, 10 hPa and 100 hPa (Fig. 10) is similar for PJDC02 and PJDM02; in the northern extratropics there is a height-dependent increase of \([\bar{u}]\) in PJDM02. The small differences near 1 hPa are consistent with the changed meridional circulation and the tidal forcing in PJDC02. The stronger negative \([D]\) in PJDC02 is evident at 1 hPa (Fig. 10). These small changes differ from the large changes in the tropical mesospheric winds of the UK Universities Global Atmospheric Modelling Programme model when the diurnal cycle was neglected (Jackson 1994). The current results agree better with other studies (e.g. Miyahara and Wu 1989), where the largest response occurs above the mesopause, which is outside the domain of the current TSM GCM. The role of tidal forcing and the structure of the tides will not be discussed further.

(e) Radiation and topographic gravity-wave drag

Attention is now devoted to the effects of the radiation frequency and the TGWD. In the southern hemisphere and the tropics the latitudinal structure of \([\bar{u}]\) and \([D]\) form two pairs, according to the frequency of the radiative-flux calculations (Fig. 10). For instance, at 100 hPa the southern STJ is about 5 m s\(^{-1}\) weaker in PJDC02 and PJDM02 than in the other two runs, consistent with \([D]\) being easterly there. In the northern hemisphere the situation is quite different: \([\bar{u}]\) in PJDM12TG is apparently forced towards the state of PJDM02, correcting the deficiencies in PJDM12. The differences in \([D]\) cannot be described so simply. The marked reduction in the magnitude of \([D]\) in PJDM12TG is consistent with the weak stationary planetary waves and the reduced transient activity.
However, the changes in $[D]$ between PJDM12 and PJDM12TG cannot explain how the mean wind in PJDM12TG is pulled towards that of PJDM02; here, the direct gravity-wave forcing [$G$] must be examined (Fig. 11).

There are two easterly-forcing maxima by [$G$], both in the northern hemisphere. The first, just above and to the north of the STJ (almost 5 m s$^{-1}$d$^{-1}$) provides the additional easterly forcing which narrows the STJ in PJDM12, returning its structure to that in PJDM02, more than compensating for the westerly planetary-wave forcing there (Fig. 10). At 10 hPa $[D]$ in PJDM12TG is considerably weaker than in the other runs but again $[G]$ at least partially compensates, changing $[u]$ (which agrees less well with that in PJDM12 at this level than at 100 and 1 hPa). The second maximum is in the mesosphere, close to the centre of the PNJ in PJDC02. At 1 hPa $[G]$ totally compensates for the minimum in $[D]$ and the magnitude of $[u]$ in PJDM12TG reaches that of PJDM02.

The latitude–height structure of $[u]$ in PJDM12TG (Fig. 11) shows a broader mesospheric jet, similar to $[u]$ in PJDC02RD (Fig. 4)—the PNJ is located further towards the subtropics (as suggested by Fig. 10); however, these changes do not extend down into the middle and lower stratosphere as those in PJDC02RD did. Again, the importance of an
adequate representation of gravity-wave drag for the upper stratosphere and mesosphere is highlighted: these results suggest that TGWD contributes at least some of the required forcing, consistent with previous studies (e.g. Rind et al. 1988a). (Note that detailed comparison of the TGWD parametrization with other schemes at high levels may be unfair, as it was developed to act as a simple scheme for the lower stratosphere.) The longitudinal structure at various levels clearly echoes the major topographic features of the northern hemisphere (although their relative contributions change with altitude).

(f) Discussion

In this section several important aspects of the GCM sensitivity to the radiation transfer and TGWD were presented. First, including the diurnal cycle in the incoming solar radiation leads to an additional easterly forcing which increases with increasing altitude; this has the expected effect on the zonal wind, reducing the strength of the wintertime westerlies, but it has little influence on the model performance—the transient behaviour is similar and the planetary-wave structure almost unchanged. Decreasing the accuracy of the radiative-flux calculations had a more dramatic effect: the zonal-mean climatology and the stationary waves both changed dramatically while the transience increased. This systematic bias was reduced considerably by including the TGWD, which is an important process in the northern lower stratosphere. This can be interpreted as the correction of
a systematic error due to degradations in one physical process by the introduction of a different mechanism. However, the transient behaviour of the TSM GCM when the TGWD was included was considerably different from that when the radiation transfer was 'properly' represented.

Regarding the effect of the TGWD, recall that Miller et al. (1989) showed that at low resolution (close to T21) there is a systematic underestimate of the northward eddy-momentum-flux convergence $M_e$ in the upper stratosphere which balances a similar underestimate in the mountain torque, leading to a realistic climatology. Only at higher resolutions, when $M_e$ becomes more realistic, was the missing drag important. This issue is not to be confused with the importance of including some representation of mesospheric gravity-wave drag, which is done very crudely in the current model. The current results suggest that one must interpret the effects of TGWD in low-resolution models, such as that of Rind et al. (1988a,b) who use a $8^\circ \times 10^\circ$ grid, with some caution.

The results of the experiments with different full-radiation time-steps are quite surprising. There is a clear change in the model performance when the accuracy of the radiation-transfer calculations is changed. The differences are marked in the winter stratosphere. However, this cannot be a direct effect because the treatment of the dominant component of the solar forcing and the cooling to space are not so closely coupled to the radiation frequency as the calculations of radiative fluxes in the troposphere. This suggests that the middle atmosphere responds to the tropospheric forcing: this will be investigated in the following section.

7. Changes in the Tropospheric Circulation

(a) Differences in the upper troposphere

The zonal part of the rotational velocity component, $[u_\phi]$, at 200 hPa (Fig. 12) for PJDC02 agrees quite well with climatology (e.g. Hoskins et al. 1989). The strong STJs are located over the eastern edges of the northern hemispheric land masses, and weaker, more zonally symmetric winds lie close to 40°S. In the tropics the climatological easterlies extend from South America, over Africa, the Indian Ocean and Indonesia to the eastern Pacific, although their latitudinal extent varies. Turning off the diurnal cycle partly results in a decrease of the strength of these easterlies in most locations in the tropics and also an increase in the strength of the subtropical westerlies; these changes never exceed 5 m s$^{-1}$. Rather larger changes are found when the radiation frequency is reduced; these are quite pronounced in the tropics but also show up at higher latitudes, the maxima of the STJ being shifted polewards by a few degrees. These differences suggest that the parametrization changes directly affect the tropical atmosphere, which consequently modifies the links with the extratropical circulation. This will be investigated further here.

(b) Differences in the tropical troposphere

It is well known that the interactions between the radiative heating and the convection in the tropics combine to drive the large-scale Hadley and Walker circulations, as well as influencing the intraseasonal oscillation and other features. In numerical models, several recent studies have reported the importance of the interactions between the various physical processes and the sensitivity to the parametrizations. Morcrette (1990) showed how the improvements to the radiation scheme in the ECMWF forecasting model led to a more vigorous convectively driven circulation due to the improved treatment of the transmission through the atmosphere. In that study the clouds were modelled after Slingo (1987). The current model used the scheme of Roeckner and Schlese (1985) and has the advantage
Figure 12. (a) Rotational component of the zonal velocity at 200 hPa for PJDC02 (contour interval 10 m s$^{-1}$) and the differences (b) PJDM02 minus PJDC02 and (c) PJDM12 minus PJDC02 (contour interval 2 m s$^{-1}$). Negative differences are shaded.
of treating liquid water as a prognostic variable. A modified version of the Kuo (1974) scheme is used for deep convection.

Several dynamical features of the tropical troposphere are compared. The tropical easterlies at 200 hPa (Fig. 12), with a dominant centre south of the equator, are reflected in $\bar{u}$ (Fig. 13) where they are confined to a narrow band centred on 12°S, with weak westerlies over the equator; the STJ is evident. The temperature structure shows the expected decrease with increasing altitude to a minimum of less than $-75$ °C at 16–18 km. The Hadley circulation is evident, with upwelling to the south of the equator (exceeding 4 mm s$^{-1}$ at 200 hPa, near 11 km), strong northward flow under the tropopause (exceeding 1.5 m s$^{-1}$ at 200 hPa) and descent centred on 20°N.

These zonal means show a clear response to the changes in the physics. Turning off the diurnal cycle (PJDM02) leads to very small changes in the thermal structure and slight (less than 2 m s$^{-1}$) increases in $\bar{u}$ at almost all locations. There is a slight change in the Hadley circulation—the ascent becomes slightly more concentrated (the 0.5 mm s$^{-1}$ changes in $\bar{w}^*$) but the mass flux across any level in this latitude band is virtually unchanged. This is not the case for PJDM12, where the transport across the pressure levels in the tropical troposphere is less than in PJDC02. The ascending branch of the Hadley Cell is about 25% weaker and moves equatorward; there is a corresponding reduction in the northward flow in the upper troposphere, which shows mainly as a decrease in the vertical extent rather than as a large decrease in the maximum strength near 11 km. The reduction in $\bar{w}^*$ leads to less adiabatic cooling in the upper troposphere and lower stratosphere (where the latent-heat release is very small); the temperature difference reaches 3 deg C at 100 hPa (16 km) in the subtropics. Over the equator $\bar{u}$ becomes westerly in the upper troposphere (cf. Fig. 12) but it decreases north of 20°N, consistent with the small change in the Coriolis forcing. Along with these changes in the quantities displayed, the zonally averaged cloud cover and the precipitation both change by around 10% near the centre of the upwelling in PJDM12. There are also some small differences in the geographical distributions of these quantities.

(c) Communication to higher latitudes

A dynamically consistent approach to studying the mechanisms of tropical–extratropical interactions is the notion that the large-scale divergence fields associated with the outflow from the convective regions can interact with the vorticity field. If $\zeta$ is the absolute vorticity and $\Lambda$ is the divergence, the vorticity equation can be written (Sardeshmukh and Hoskins 1988):

$$\left( \frac{\partial}{\partial t} + \mathbf{v}_\psi \cdot \nabla \right) \zeta = -\nabla \cdot \mathbf{v}_\zeta - \Lambda \zeta + F,$$  \hspace{1cm} (6)

where $F$ represents the effects of frictional motions (and unresolved processes in the numerical model) on the vorticity. The horizontal velocity has been partitioned into two components; the rotational, divergence-free component $\mathbf{v}_\psi$ (the zonal component of which is shown in Fig. 12) and the irrotational, divergent component $\mathbf{v}_\zeta$, where $\psi$ and $\zeta$ represent the stream function and the velocity potential. The important aspect of this formulation (Eq. (6)) is that even though $\mathbf{v}_\zeta$ may be weak it can be directed across the $\zeta$ gradients, whereas $\mathbf{v}_\psi$ tends to be directed along these gradients, so that the divergent flow can lead to a significant modification of the vorticity field.

The changes in the zonally averaged Hadley circulation are the zonal-mean signal of changes in the three-dimensional divergent mass flow, which is partially illustrated by $\chi$ at 200 hPa (Fig. 14). For PJDC02 there is a global-scale ‘wave-number-1’ type of structure, in good agreement with assimilated observations (Hoskins et al. 1989), with negative
Figure 13. Zonal-mean cross-sections, 30°S–30°N, 2–18 km, showing $\bar{u}$ (contour interval 5 m s$^{-1}$, top row), $\bar{T}$ (contour interval 5 °C, second row), $\bar{v}'$ (contour interval 2.5 dm s$^{-1}$, third row), and $\bar{w}'$ (contour interval 1 mm s$^{-1}$, bottom row) for run PJDC02 (left column). The differences PJDM02 minus PJDC02 (second column) and PJDM12 minus PJDC02 (third column) are also shown with respective contour intervals 0.5 m s$^{-1}$, 0.5 °C, 0.5 dm s$^{-1}$, and 0.25 mm s$^{-1}$. In these plots negative values are shaded. See text for explanation of symbols and definition of models.
Figure 14. (a) The velocity potential ($\chi$) (thick lines, contour interval $2.5 \times 10^{-6}$ m$^2$s$^{-1}$) and the stream function ($\psi$) (thin lines, contour interval $2.5 \times 10^{-5}$ m$^2$s$^{-1}$) at 200 hPa for PJDC02. The differences of $\chi$: (b) PJDM02 minus PJDC02 and (c) PJDM12 minus PJDC02 (contour interval $0.5 \times 10^{-6}$ m$^2$s$^{-1}$, negative values shaded). See text for definition of models.
values centred over the west Pacific and Indonesia, slightly to the south of the equator, and positive values centred in the northern tropics over Africa. This is consistent with the south–north zonal-mean flow in Fig. 13. The differences between PJDC02 and PJDM02 are generally quite small; the pattern remains the same with some quantitative differences in the strength of the gradients over the regions of convection, but the extreme values (and hence the global-scale divergent circulation) remain essentially unchanged. In contrast, the differences between PJDM12 and PJDC02 are larger. Particularly, the gradients of \( \chi \) near 60\(^\circ\)E are weakened considerably, so that the divergent flow there is reduced. This is the region of the jet-stream entrance (Fig. 12), so this reduction in divergent flow reduces the cross-streamline advection of the rotational flow and hence decreases the production of relative vorticity in the jet-stream entrance, leading to the weaker jet in PJDM12 compared with PJDC02.

Following Sardeshmukh and Hoskins (1988), the right-hand side of Eq. (6) can be regarded as a Rossby-wave source (RWS):

\[
\text{RWS} = -v_x \cdot \nabla \zeta - \Lambda \zeta,
\]

which can illustrate the generation of extratropical planetary Rossby waves due to the interactions between the rotational and divergent components of the flow. This viewpoint at least allows a consistent dynamical interpretation of the time-averaged structure of the atmosphere or the model. RWS for the three model runs PJDC02, PJDM02 and PJDM12 (Fig. 15) shows again the similarities between PJDC02 and PJDM02 but the large differences in PJDM12. These lie over the North Atlantic, where RWS is much stronger in PJDM12, and over the entire Pacific and Himalayan region where RWS in PJDM12 is weak. Indeed, RWS in PJDM12 has a more predominant wave-number-2 structure than in the other two runs, where it is more confined in longitude and has a suggestion of a wave-number-3 structure. This is consistent with the increased strength of the longest planetary waves in PJDM12.

\((d)\) Interpretation

Tracing the sensitivity of the winter stratosphere in the TSM GCM to the frequency of full-radiation computations has led to the tropical troposphere. It is there that a delicate balance between the cloud cover and the radiation transfer exists. Convective activity is generally short-lived, so that reducing the frequency of the radiation computations presumably leads to an imbalance between the actual cloud cover and that used in the calculations of transmissivity used for the radiation. Although the latent heating is updated at every time-step, the optical properties of the clouds are constant between the full-radiation steps, so the cloud distributions are clearly inconsistent in these two parts of the physics calculations. Consequently, the Hadley and Walker circulations spin down when the radiation is calculated less frequently, reducing the strength of the divergent outflow in the upper troposphere. This is communicated to higher latitudes through the physical effect that the cross-streamline divergent flow can affect the generation of vorticity anomalies (the Rossby-wave source); given the constraint that the extratropical orography remains fixed, thereby limiting the extent to which the direct extratropical forcing can be perturbed, this change to the RWS directly leads to a response in the zonally asymmetric structure. The high-latitude planetary waves in the upper troposphere are thereby modified. This in turn affects the strength of dynamical forcing of the middle atmosphere.

This interpretation is at least dynamically consistent, within the diagnostic framework presented. An important point is that the radiative fluxes obviously need to be calculated sufficiently often for the model to give a realistic simulation: every 12 hours is inadequate.
Figure 15. The Rossby-wave source (RWS), defined according to Eq. (6), for (a) PJDC02, (b) PJDM02, and (c) PJDM12 (see text). The dark/light shaded regions in (b) and (c) highlight regions where positive/negative differences of more than 10 units from PJDC02 occur.
Introducing these inconsistencies into the experiments (and noting that the studies in Pawson et al. (1991, 1995) were run under these conditions, albeit with different parametrizations of the hydrological cycle) has illustrated how errors which originate in one region of the modelled atmosphere (in this case, the tropical troposphere) can be communicated to remote regions (here, the extratropical stratosphere via the troposphere) by the dynamical processes at work. More generally, although the degradations in this study were deliberately introduced, they may be applicable to other models which unintentionally use deficient physical processes in the tropical troposphere.

8. SUMMARY AND CONCLUSIONS

The formulation of the Berlin TSM GCM has been described in some detail and its performance evaluated. A reference integration revealed considerable improvements over the previous version of the model (Pawson et al. 1991, 1995). These must arise from the improved physical parametrizations, largely in the troposphere, which have an indirect effect on the stratosphere due to dynamical and radiative coupling.

Results from a series of sensitivity experiments using the comprehensive TSM GCM were presented. To avoid the necessity of performing several multi-annual simulations, the experiments were performed in the framework of a set of moderately long perpetual-January integrations, designed to enable the signal of changes to be separated from the internal variability of the model. This was achieved by separating responses which resemble the dominant mode of variability of the zonal-mean structure (and are a consequence of the model occupying one state more readily) from those which introduce a totally new pattern of variability. The results have shown the following sensitivities:

- Decreasing the strength of the idealized drag at the uppermost model levels (in the mesosphere) led to local changes in the circulation but little significant change in the zonal-mean structure below the upper stratosphere (experiments PJDCO2 and PJDCO2RW). The changes resembled the first mode of variability of the model.
- Increasing the depth over which this idealized drag acted led to major changes in the structure of the modelled PNJ; in particular, the strong, vertical PNJ in PJDCO2 was replaced by a broader feature which bears more resemblance to the observed split jet and extends lower, leading to more realistic temperatures in the lower polar stratosphere. These changes do not resemble the natural variability patterns in the model. This suggests that a deeper profile of $\gamma$ may have been more appropriate for the reference integration (PJDCO2). The induced changes extended into the troposphere.
- Turning off the diurnal cycle had the expected changes in the mesosphere: slightly stronger net westerly flow due to the lack of easterly forcing by the tidal drag, but there were no appreciable effects on the stationary waves, the transient behaviour, or the troposphere (experiments PJDCO2 and PJDMO2).
- Reducing the frequency of 'full-radiation computations' from 2 hours to 12 hours had a dramatic effect on the modelled climatology: the strength of the PNJ decreased dramatically and the stationary-wave field became much stronger. These differences could be traced back to large changes in the strength of the Hadley circulation and the divergent flow in the tropical troposphere which affect the vorticity generation in the subtropics, causing shifts of the upper-tropospheric subtropical jet and changes in the tropospheric structure which propagate into the stratosphere (experiments PJDCO2, PJDMO2 and PJDMO2).
- The use of a TGWD parametrization 'corrected' the deficiencies introduced by degrading the radiation calculations, at least in the upper troposphere, leading to a
restoration of the zonal-mean state. The transient behaviour was, however, suppressed and the lower stratospheric planetary waves were degraded.

The use of perpetual-January experiments is naturally no substitute for long climate integrations including the annual cycle; this could be particularly problematic if ozone were to be determined self-consistently in the model. However, the experimental framework provided here was a relatively efficient means of conducting these sensitivity tests. Such tests provide the basis for understanding the climatological structure of the model and how it may react to imposed perturbations or other changes in the physical parametrizations. A discussion of the climatology of a long, multi-annual reference integration of the Berlin TSM GCM is presented by Langematz and Pawson (1997).

Two aspects of the study are quite alarming: the extreme sensitivity of the model to the frequency of full-radiation computations, and the fact that changes induced by modifying one parameter can be amended by introducing a new parametrization. That such GCMs can show large responses to small changes was first shown by Fels (1985). The current results probably show that such GCM experiments must still be interpreted with caution. Another aspect, which may not have been fully accounted for in previous studies of TSM GCM responses to external or internal change, is the analysis in terms of the structure of the dominant oscillation patterns in the model.

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