Tides in the Extended UGAMP General Circulation Model

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SUMMARY

In this paper the effect of tides on the zonal mean flow in the Extended UK Universities Global Atmospheric Modelling Programme General Circulation Model (EUGCM) is examined. Previous modelling studies have shown that omitting the diurnal cycle of solar radiation gives rise to changes in the zonal wind of the order of 10–20 m s\(^{-1}\) near the mesopause, and to larger changes in the lower thermosphere. In the EUGCM, which has a top level near the mesopause, turning off the diurnal cycle gives rise to model zonal winds in July that are up to 65 m s\(^{-1}\) stronger near the equatorial mesopause. It is shown that such zonal wind changes are reduced by up to 20 m s\(^{-1}\) when the model horizontal diffusion is reduced by a factor of 100, but are increased by up to 10 m s\(^{-1}\) when the model vertical diffusion is switched off above the lower stratosphere. Such large differences in zonal wind between diurnal and non-diurnal runs cannot be attributed to excessive model tidal amplitudes, since they are found to be in good general agreement with both observations and other modelling studies. An exception is the diurnal tide in the subtropics, which decreases with height above about 0.01 mb instead of increasing to the model top. It is shown that this behaviour is insensitive to changes in the model diffusion, but it is suggested that it could be caused by spurious reflections at the model top.

1. INTRODUCTION

In order to model diurnal and semi-diurnal thermal tides accurately in the upper mesosphere and lower thermosphere, many two-dimensional tidal models have used upper boundaries of between 300 and 550 km (e.g. Lindzen and Hong 1974; Miyahara 1981; Forbes 1982; Vial 1986). This eliminates distortions in the tidal solutions at lower levels caused by reflections from the upper boundary (e.g. Lindzen et al. 1968). For example, Vial (1986) found that tidal solutions below 110 km were insensitive to changes to the default upper boundary of 500 km. For model validation purposes, it is particularly important that solutions below about 100 km are not affected by reflections from the upper boundary, since the bulk of tidal observations have been made there; rocketsonde observations primarily cover the 30–60 km region (e.g. Reed et al. 1969; Reed 1972; Nastrom and Belmont 1976), whilst radar observations chiefly cover the 70–100 km region (e.g. Bernard 1981; Manson et al. 1981; Carter and Balsley 1982; Ahmed and Roper 1983). Three-dimensional tidal models (e.g. Kähler 1989; Wu et al. 1989) can use lower upper boundaries than two-dimensional models, since the horizontal diffusion present in these models reduces the problems with reflection from the upper boundary. The upper level of the Wu et al. model is at 165 km, whilst the Kähler model, which also uses enhanced horizontal diffusion at upper model levels, has an upper boundary at 120 km. It is important that these models extend to such heights because it has been shown that momentum deposition by tides affects the zonal mean flow in the upper mesosphere and lower thermosphere. For example, the quasi nonlinear two-dimensional model of Miyahara and Wu (1989) showed that omitting the diurnal cycle of solar radiation increased zonal-mean zonal winds by a maximum of 10 m s\(^{-1}\) in the upper equatorial mesosphere, and by a maximum of 40 m s\(^{-1}\) in the extratropical lower ther-

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mosphere, whilst Wu et al. (1989) showed that momentum deposition due to tides led to large induced winds in the lower thermosphere, the largest being under -135 m s\(^{-1}\) near 110 km over the equator.

In this paper we examine the effect of tides on the mean flow in the Extended UGAMP* General Circulation Model (EUGCM). Such effects have been largely ignored in middle-atmosphere GCMs. Miyahara et al. (1993) ran a GCM with a top boundary of 165 km under perpetual October conditions. They showed that dissipating diurnal tides induced zonal mean easterlies and a cellular meridional wind structure in the equatorial upper mesosphere and lower thermosphere. The tidal-induced easterly acceleration was of the order of 10\(^{-4}\) m s\(^{-2}\) in the equatorial lower thermosphere. Hunt (1986, 1990) included a diurnal cycle in a GCM with a top level of 100 km. Comparison of diurnal and non-diurnal versions showed that including the diurnal cycle decreased the equatorial zonal wind by up to 65 m s\(^{-1}\) at 100 km, and by up to 30 m s\(^{-1}\) at 90 km. However, part of this difference may be due to the fact that the diurnal version includes a gravity-wave parametrization scheme, whilst the non-diurnal version does not. Any substantial differences in tidal amplitude or associated momentum deposition between tidal models (e.g. Lindzen and Hong 1974; Vial 1986; Miyahara and Wu 1989; Wu et al. 1989; Miyahara et al. 1993) and the EUGCM may be due to the lower top level in the EUGCM (about 90 km). Hunt and Manabe (1968) and Zwiers and Hamilton (1986) used GCMs with diurnal cycles that had top levels in the middle stratosphere and which employed a 'rigid lid' upper boundary condition. Zwiers and Hamilton found a tendency for the effects of the rigid lid to compensate for the omission of the thermal forcing above the top level of the model. It is likely that this problem will be less serious in the EUGCM, since the omitted thermal forcing above the top level of the EUGCM is much weaker than that omitted above the top level of the Zwiers and Hamilton model. In addition, the imposition of a 'rigid lid' can produce seriously distorted results (e.g. Lindzen et al. 1968; Boville and Cheng 1988). However, problems with reflection from the upper boundary may manifest themselves differently in the EUGCM, since it does not have a rigid lid upper boundary condition, but instead uses a 'sponge layer' of enhanced horizontal diffusion in the top five model levels in order to reduce reflection of waves from the model top.

The aim of this paper is firstly to describe the effect of tides on the mean flow of the EUGCM and, secondly, to explain this effect via sensitivity studies and a documentation of modelled tidal amplitudes and phases. The layout of the paper is as follows. The model is described in section 2, whilst in section 3 the effect of tides on the model mean flow is examined. Such results show that differences in zonal-mean zonal wind in the upper mesosphere between diurnal and non-diurnal simulations are larger than in other studies (e.g. Miyahara 1981; Miyahara and Wu 1989; Wu et al. 1989). Miyahara and Wu (1989) have shown that the size of such differences may depend on the amount of vertical diffusion present in the model. In addition, Boville (1985) has shown that decreasing horizontal diffusion decreases wave forcing in a troposphere/stratosphere GCM. Accordingly, the latter part of section 3 examines the sensitivity of these zonal wind differences in the EUGCM to both vertical and horizontal diffusion. Another possible reason for such large zonal wind differences is that the EUGCM tidal amplitudes are much larger than observed. Hence, in section 4 diurnal and semi-diurnal tidal amplitudes and phases are compared with observations and other modelling studies. The sensitivity of these amplitudes to changes in the model diffusion is also examined. Conclusions appear in section 5.

* UK Universities Global Atmospheric Modelling Programme.
2. Model Description

The EUGCM is a troposphere/stratosphere/mesosphere model with 47 levels from the surface to about 90 km, and has a vertical resolution of between 2 and 3 km in the middle atmosphere. The model is derived from the cycle 27 European Centre for Medium-range Weather Forecasts (ECMWF) forecast model (e.g. Tiedtke et al. 1988) and includes a full set of parametrizations for surface exchange processes, the hydrological cycle, and clouds, radiative heating and subgrid-scale processes. The model prognostic equations calculate vorticity, divergence, temperature, specific humidity and surface pressure, and spectral coefficients are used to represent the horizontal variables. Simulations described here have been run at T21 horizontal resolution (i.e. approximately $5^\circ \times 5^\circ$) with a seasonal cycle of solar radiation and specified, seasonally varying, sea surface temperatures.

Two radiation schemes are used in the EUGCM. The scheme used in the troposphere and lower stratosphere (Geleyn and Hollingsworth 1979) produces a warm bias in lower stratospheric temperature (Mocquet 1990), and also does not represent the transition from Lorentz to Doppler broadening in the mid-stratosphere; therefore it cannot be confidently used in the lower stratosphere or above. Accordingly, it and the middle atmosphere radiation scheme are merged between 20 and 74 mb by using a heating rate that is a linear combination of the heating rates given by the two schemes. The middle atmosphere solar-radiation scheme was developed by Shine and Rickaby (1989) and considers absorption by ozone and oxygen, whilst the infrared scheme (Shine 1987) calculates infrared cooling and exchange from the 15 $\mu$m carbon dioxide and 9.6 $\mu$m bands, and from water vapour. The carbon dioxide and ozone cooling rates are calculated using latitudinally varying Curtis matrices, whilst the water vapour cooling rates are calculated using an approximation similar to the cooling-to-space approximation. Departures from local thermodynamic equilibrium are also included. The model ozone climatology uses ozone values from the COSPAR* International Reference Atmosphere (CIRA) (Keating et al. 1990) at pressures less than 20 mb, and the 1979–81 mean values from the Nimbus-7 Satellite Backscattered Ultra Violet (SBUV) instrument (see, for example, McPeters et al. 1984) between 20 and 74 mb; at higher pressures the ECMWF model ozone climatology is used (see Gray et al. (1993) for more details).

The model simulations shown here are summarized in Table 1. The principal simulation (run GW) is run for one year from an initial data set based on an ECMWF analysis for 15 January 1987 (see Gray et al. (1993) for more details of this data set). All other simulations described here are started from run GW fields at the end of June (for July simulations) or the end of December (for January simulations). Run GW uses a gravity-wave parametrization scheme that parametrizes two gravity waves: the first is an orographically forced gravity wave with a phase speed of 0 m s$^{-1}$ (see Palmer et al. 1986); the second has a phase speed of 20 m s$^{-1}$, a globally uniform momentum flux of $3 \times 10^{-4}$ N m$^{-2}$, and is generated at the 252 mb level. The reasons for choosing these values appear in Jackson (1993). Effects of tides on the mean flow are examined by comparing run GW with a similar simulation in which the diurnal cycle is switched off (run GWOFF).

Subsequent simulations investigate the effect of vertical and horizontal diffusion on tides. In run GW eighth-order horizontal diffusion is applied to all the model prognostic variables (except surface pressure). The damping time-scale at the shortest resolvable length-scale is 1.2 hours for divergence and 3 hours for the other variables (this implies damping times for the migrating diurnal tide of $1.42 \times 10^8$ days and $3.56 \times 10^8$ days,

* The Committee on Space Research of the International Council of Scientific Unions.
TABLE 1. SUMMARY OF THE EXPERIMENTS

<table>
<thead>
<tr>
<th>Run identifier</th>
<th>Description</th>
</tr>
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<tbody>
<tr>
<td>GW</td>
<td>Default run. Diurnal cycle on. Upper-level horizontal diffusion is enhanced by successively decreasing damping time by a factor of 2 from the fifth from top level to the second from top level, above which it is constant. Horizontal diffusion damping time-scale (at shortest resolvable length-scale) is 1.2 hours for divergence and 3 hours for the other variables. Vertical diffusion on.</td>
</tr>
<tr>
<td>GWOFF</td>
<td>Like GW, except the diurnal cycle is switched off.</td>
</tr>
<tr>
<td>HDF</td>
<td>Like GW, except the upper-level horizontal diffusion is enhanced by successively decreasing the damping time by a factor of 2.5 from the third from top to top model level.</td>
</tr>
<tr>
<td>HDFOFF</td>
<td>Like HDF, except the diurnal cycle is switched off.</td>
</tr>
<tr>
<td>LHDF</td>
<td>Like GW, except the damping time for horizontal diffusion is increased by a factor of 100 at all model levels.</td>
</tr>
<tr>
<td>LHDFOFF</td>
<td>Like LHDF, except the diurnal cycle is switched off.</td>
</tr>
<tr>
<td>VDF</td>
<td>Like GW, except the vertical diffusion is switched off above the 15 mb level.</td>
</tr>
<tr>
<td>VDFOFF</td>
<td>Like VDF, except the diurnal cycle is switched off.</td>
</tr>
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</table>

respectively). This damping is enhanced in the top five model levels by successively decreasing the damping time by a factor of two from the fifth from top level to the second from top level, above which it is constant. The impact of changing this enhanced diffusion is examined in run HDF, in which the diffusion is instead enhanced at upper levels by successively decreasing the damping time by a factor of 2.5 from the third from top model level to the top model level (run HDFOFF is the corresponding non-diurnal run). The effect of reducing the horizontal diffusion everywhere is examined in run LHDF, in which the damping time is increased by a factor of 100 at all model levels (run LHDFOFF is the corresponding non-diurnal run). Run GW employs a Richardson number-dependent vertical-diffusion scheme (Louis 1979; Louis et al. 1982). The effect of vertical diffusion on the modelled tides and on the mean flow is examined in run VDF, in which the vertical-diffusion scheme is switched off above the 15 mb level (run VDFOFF is the corresponding non-diurnal run).

3. EFFECT OF TIDES ON THE MEAN FLOW

In this section the effect of tides on the modelled mean flow is investigated by comparing simulations run with and without the diurnal cycle of solar radiation. (Only time-averaged zonal mean fields are shown here. A further study could examine longitudinal and temporal variations of the effects of tides in the simulated middle atmosphere. Such a study could be incorporated into a more general and comprehensive description of the EUGCM climatology: model latitude/longitude temperature fields in the stratosphere (not shown) are in good qualitative agreement with observations (e.g. Fleming et al. 1990), but this examination has not been applied to a wide range of model variables, model levels, or to differing model simulations. In addition, short-time-scale model variability has not been investigated.)

The mean meridional wind for run GW in July (Fig. 1(a)) is mainly winterward (northerly) in the upper equatorial mesosphere with a minimum of less than \(-12\) m s\(^{-1}\) near 0.003 mb and 15°S. There is also a small region of southerly flow in the northern (summer) tropics near 0.002 mb and between 0° and 20°N, with maximum winds of less than 4 m s\(^{-1}\). This wind pattern is in agreement with other models. Miyahara and Wu (1989) ran a quasi-linear two-dimensional model under northern winter solstice
conditions: near 0.002 mb poleward flow was up to 2 m s\(^{-1}\) on the summer side of the equator and greater than 10 m s\(^{-1}\) on the winter side of the equator. The GCM of Miyahara et al. (1993) produced poleward winds near the equator of up to 2 m s\(^{-1}\) near 0.002 mb, although it should be noted that this simulation was for October rather than solstice conditions. In both these models the region of poleward flow near the equator is present between about 0.003 and 5 \(\times\) 10\(^{-5}\) mb (85 and 110 km), with a maximum near 2.3 \(\times\) 10\(^{-4}\) mb (100 km). This therefore implies that the vertical extent of the EUGCM is not large enough to simulate such meridional wind structure fully.

Model results (Miyahara and Wu 1989; Wu et al. 1989) show that these winds are part of a two-cell circulation, which also includes ascent near the equator and descent in the subtropics. The magnitude of such modelled vertical velocities is in general less than 50 mm s\(^{-1}\) below the 0.001 mb level. The vertical velocity for run GW in July (Fig. 1(b)) is in reasonable agreement with these models. There is ascent near the equator, with maximum values of over 30 mm s\(^{-1}\) near 10\(^{\circ}\)S, and regions of descent in the subtropics, with minimum values of \(-60\) mm s\(^{-1}\) near 25\(^{\circ}\)S, and less than \(-20\) mm s\(^{-1}\) near 10\(^{\circ}\)N. Miyahara and Wu (1989) also performed a non-diurnal simulation which produced an essentially one-cell circulation in the upper mesosphere and lower thermosphere, with ascent in the summer hemisphere, descent in the winter hemisphere, and strong winterward flow in the upper mesosphere. EUGCM results are in good agreement with this.

The mean meridional wind for run GWOFF in July (Fig. 1(c)) is almost exclusively northerly in the mesosphere, whilst the vertical-velocity field (not shown) shows ascent (descent) in the summer (winter) extratropical mesosphere and near-zero winds in the tropics.

Figure 2(a) shows the zonal-mean zonal wind field for run GW for July. This field is in reasonable agreement with the climatology of Fleming et al. (1990) (not shown). Comparison with the corresponding field for run GWOFF (Fig. 2(b)) shows that the zonal-mean zonal winds for run GWOFF are similar to those in Fig. 2(a) below the stratopause, but are clearly much more westerly in the upper equatorial mesosphere. Such differences are better illustrated in Fig. 2(c), which shows the difference between Figs. 2(a) and 2(b). This figure confirms that switching off the diurnal cycle increases the zonal wind in the upper equatorial mesosphere dramatically; the largest change in zonal wind in that region is over 60 m s\(^{-1}\). In addition, it is asymmetric about the equator, its maximum being on the winter side of the equator. This may be related to the diurnal tidal amplitude being larger in the winter subtropics than in the summer subtropics (see later). There are also small decreases in zonal wind in the upper mesosphere near 50\(^{\circ}\)S and 30–50\(^{\circ}\)N. The corresponding plot for January (not shown) has similar features to July, including a region of increased zonal wind that is asymmetric about the equator, with a maximum of over 30 m s\(^{-1}\), and regions of decreased zonal wind in the subtropics of up to \(-5\) m s\(^{-1}\) (winter hemisphere) and \(-25\) m s\(^{-1}\) (summer hemisphere).

Other authors (e.g. Miyahara and Wu 1989; Wu et al. 1989; Miyahara et al. 1993) have suggested that the zonal wind differences such as those in Fig. 2(c) are caused by momentum deposition due to tides. This is examined further here. Figure 3 shows the forcing due to Eliassen–Palm flux divergence (EPFD) from resolved model waves (including tides) from run GW in July. The pattern of EPFD in the upper southern and equatorial mesosphere broadly matches the wind differences in Fig. 2(c) (but with opposite sign since Fig. 2(c) shows (run GWOFF – run GW) values). In addition, the total forcing of the zonal wind has a rather similar pattern, since the contributions to this forcing from the residual circulation and gravity-wave drag (both not shown) broadly cancel each other. In the equatorial and southern subtropical upper mesosphere (where zonal wind differences in Fig. 2(c) are largest) EPFD values in Fig. 3 are in general
Figure 1. Monthly mean zonal mean fields in July. (a) Meridional wind for run GW, (b) vertical velocity for run GW, and (c) like (a) except values for run GWOFF are shown. Positive values represent southerly winds in fields (a) and (c), and ascent in field (b). The contour interval is 2 m s⁻¹ for fields (a) and (c), and 10 mm s⁻¹ for field (b). Negative values are represented by dashed contours.
stronger than corresponding values for run GWOFF (not shown). Therefore, it appears that the changes in zonal wind in these regions may be caused by momentum deposition due to tides. In contrast to values in the summer and equatorial upper mesosphere, EPFD in the northern hemisphere extratropics in Fig. 3 is very large and is not similar to the zonal wind difference there in Fig. 2(c). Jackson (1993) showed that the parametrized gravity-wave drag is very large and westerly in this region. This large gravity-wave drag is partially balanced, not only by the residual circulation but also by the EPFD, indicating resolved wave/gravity-wave interactions. Thus the change in zonal wind in the northern extratropical mesosphere in Fig. 1(c) is probably caused not just by differences in EPFD (chiefly due to the absence of tides in run GWOFF) and gravity-wave drag between runs GW and GWOFF, but also by different resolved wave/gravity-wave interactions in these two runs.

The changes to the upper mesospheric EUGCM zonal winds when the diurnal cycle is removed are much larger than in other modelling studies. Results from two models are shown here. Miyahara and Wu (1989) use a quasi nonlinear two-dimensional model with a specified profile of vertical diffusion that has a maximum of approximately $3 \times 10^6 \text{cm}^2\text{s}^{-1}$ near 110 km. By contrast, Wu et al. (1989) use a simplified GCM in which vertical diffusion is not specified (apart from a background value of $10^4 \text{cm}^2\text{s}^{-1}$), but is instead dependent on the Richardson number. The Wu et al. model also parametrizes convective instability via a dry convective-adjustment scheme. Figure 4(a) shows the impact on the zonal wind of removing the diurnal cycle in the Miyahara and Wu model (note that this plot is for January rather than July and that the sign of the zonal wind changes is opposite to that in Fig. 2(c)). The pattern of the zonal wind differences below 90 km is similar to the EUGCM, with a region of asymmetrical wind differences near
Figure 2. Monthly mean zonal-mean zonal wind in July. (a) Run GW, (b) run GWOFF, and (c) field in (b) minus field in (a). Westerly winds are positive, easterly winds are negative. The contour interval is 10 m s⁻¹ for fields (a) and (b) and 5 m s⁻¹ for field (c). Negative values are represented by dashed contours.
Figure 2. Continued.

Figure 3. Forcing due to Eliassen–Palm flux divergence from resolved model waves for run GW in July. The contour interval is 15 m s\(^{-1}\) day\(^{-1}\). Negative values are represented by dashed contours and negative regions are shaded.
the equator, and changes of opposite sign in the subtropics (not visible in Fig. 4(a) since the zero contour is not plotted). However, the zonal wind differences there are weaker than in the EUGCM by up to 50 m s\(^{-1}\) near the equator and by between 5 and 15 m s\(^{-1}\) in the subtropics. The zonal mean wind induced by the breaking diurnal tide in the Wu et al. model (Fig. 4(b)) is also much weaker in the upper mesosphere than in the EUGCM. The induced wind is symmetric about the equator, unlike the zonal wind differences in Figs. 2(c) and 4(a), which have their maxima on the winter side of the equator. This may be related to the fact that in Wu et al.'s model, upper mesospheric diurnal amplitudes are about equal in the subtropics of both hemispheres, whilst in Miyahara and Wu's model and in the EUGCM (see later) winter subtropical amplitudes exceed those in the summer subtropics.

(a) Effect of diffusion on momentum deposition due to tides

The changes in zonal wind between diurnal and non-diurnal EUGCM simulations (Fig. 2(c)) may be larger than in the Miyahara and Wu (1989) and Wu et al. (1989) models because the EUGCM has its top level near 90 km, whilst the other two both have top levels above 160 km. Indeed, the largest zonal wind changes in these models are in the lower thermosphere above the top level of the EUGCM: in the Miyahara and Wu model the largest zonal wind change is about 40 m s\(^{-1}\), and is located near 110 km and 30°N (Fig. 4(a)), whilst in the Wu et al. model the largest induced zonal wind is less than \(-135\) m s\(^{-1}\) and is located near 110 km and 0° (Fig. 4(b)). Thus, it is possible that the EUGCM momentum that would otherwise be transported by tides to the thermosphere is instead deposited in the mesosphere due to the horizontal diffusion, or more particularly in the upper mesosphere due to the enhanced horizontal diffusion there. In addition, the vertical diffusion may be important in determining the zonal wind changes in Fig. 2(c). Miyahara and Wu (1989) showed that increasing vertical diffusion leads to an increase in momentum deposition due to tides at lower altitudes. Accordingly, the sensitivity of
momentum deposition due to tides on the zonal mean flow to changes in both horizontal and vertical diffusion is examined below.

The effect of horizontal diffusion on momentum deposition due to tides is examined here in two experiments. In runs HDF and HDFOFF changes to the enhanced upper-level horizontal diffusion have little impact on the model mean flow (not shown). The second experiment investigates how the model will respond to larger and more widespread changes in the horizontal diffusion by increasing the damping time by a factor of 100 at all model levels (runs LHDF and LHDFOFF). The zonal-mean zonal wind for July for run LHDF (Fig. 5(a)) is reasonably similar to the corresponding field for run GW, except that upper mesospheric equatorial easterlies are weaker by up to 15 m s$^{-1}$, and the maximum winter middle-atmosphere jet is about 10 m s$^{-1}$ stronger and is located about 20° nearer the pole. The major difference between Fig. 5(a) and the corresponding field for run LHDFOFF (not shown) is that the upper mesospheric winds in run LHDFOFF are westerly, with a maximum of over 40 m s$^{-1}$. Such differences are shown more clearly in Fig. 5(b). In the mesosphere north of 60°S the field in Fig. 5(b) is similar in pattern to the differences in zonal wind between runs GW and GWOFF (Fig. 2(c)). However, the zonal wind differences in the upper equatorial mesosphere are up to 20 m s$^{-1}$ smaller. Both meridional and vertical wind fields for run LHDF (not shown) are broadly similar to those for run GW. An exception is the region of descent in the upper mesosphere centred near 20°S, which is over 20 mm s$^{-1}$ weaker in run LHDF than in run GW. This weaker descent, combined with weaker easterly shear, results in weaker easterly forcing from vertical advection in run LHDF compared with run GW, and the total zonal wind forcing (not shown) is accordingly also less easterly in this region. Such weaker descent and weaker easterly shear implies weaker momentum deposition due to tides. This is confirmed by examining the EPFD for run LHDF (Fig. 6), which shows that in the southern and equatorial upper mesosphere the pattern of EPFD is similar to that for run GW (Fig. 3), but that the values are generally weaker, especially above the 0.01 mb level. This is an extension of the results of Boville (1985), who found that decreasing wave damping in a troposphere/stratosphere GCM led to a decrease in wave forcing. It is not clear whether further reducing the model horizontal diffusion would lead to upper mesospheric zonal wind differences as small as those found by Miyahara and Wu (1989) and Wu et al. (1989), or whether it would instead increase spurious reflection of waves from the model top. Full examination of this would require a version of the EUGCM with a higher top level.

The influence of vertical diffusion on the mean flow is examined by switching off model vertical diffusion above the 15 mb level (runs VDF and VDFOFF). Differences in zonal wind between runs VDF and VDFOFF (Fig. 7) are in general between 5 and 10 m s$^{-1}$ larger than in Fig. 2(c) in the southern subtropical upper mesosphere, but are similar elsewhere. In this region, differences in residual circulation and gravity-wave drag between runs VDF and VDFOFF (not shown) are similar to corresponding differences between run GW and GWOFF values. The larger zonal wind differences in Fig. 7 instead appear to be caused by differences in EPFD between runs GW and VDF. Where the values in Fig. 7 are largest (near 0.001 mb and between 10 and 30°S), the EPFD in run VDF (Fig. 8) is over 45 m s$^{-1}$day$^{-1}$ stronger than the EPFD in run VDFOFF (not shown), and over 15 m s$^{-1}$day$^{-1}$ stronger than for run GW (Fig. 3). In the southern subtropical upper mesosphere the westerly forcing of the zonal wind due to vertical diffusion in run GW is between 5 and 10 m s$^{-1}$day$^{-1}$ (not shown). An effect of this may be to lower the level at which momentum deposition due to tides takes place (consistent with Miyahara and Wu (1989)): in run GW (Fig. 3) the largest easterly EPFD values in the equatorial mesosphere are chiefly located near 0.003 mb and 20 to 30°S, whereas in run VDF (Fig.
Figure 5.  (a) Monthly mean zonal-mean zonal wind in July for (a) run LHDF and (b) run LHDFOFF minus field in (a).  The contour interval is 10 m s\(^{-1}\) for field (a) and 5 m s\(^{-1}\) for field (b).  Negative values are represented by dashed contours.
Figure 6. Like Fig. 3, except values for run LHDF are plotted.

Figure 7. Monthly mean zonal-mean zonal wind in July for run VDFOFF minus the corresponding field for run VDF. The contour interval is $5 \text{ m s}^{-1}$ and negative values are represented by dashed contours.
8) the largest such easterly forcings are situated at the same latitudes, but are concentrated near the model top level. Differences between EPFD in runs GW and VDF are much smaller south of 40°S, where the vertical diffusion in run GW is much weaker.

4. Tide amplitudes and phases

The results from section 3 show that switching off the diurnal cycle has a dramatic effect on the upper mesospheric zonal winds in the EUGCM. Furthermore, such changes in zonal wind are affected by the specification of vertical and horizontal diffusion. However, it is also possible that the momentum deposition due to tides is too great because the diurnal and semi-diurnal tides present in the model have much greater amplitudes than observed. In addition, it is of interest to note the effect on modelled amplitudes and phases of having a model top level near 90 km, instead of in the thermosphere or above as is common in other modelling studies of tides. Accordingly, in this section the modelled tidal amplitudes and phases are described and compared with those obtained from observations and other modelling studies.

The tidal amplitudes and phases shown here are calculated from 16-day periods of the model integration. These results, and the comparisons made with observations, broadly illustrate the model's ability to simulate tides. However, the comparisons should be treated with caution for two reasons. Firstly, observations (e.g. Glass et al. 1978) show that tides can vary significantly over time-scales of the order of a few days. Secondly, whilst most observational data sets comprise data from one station only, model results are averaged over longitude. It should be noted that a significant degree of longitudinal variability can exist in the tidal behaviour, owing for example to tide/gravity-wave
interaction, or to the presence of non-migrating tides. In addition, tidal amplitudes can be influenced by tropospheric temperatures (Lindzen 1968) and tropospheric winds (Wallace and Tadd 1974).

(a) Diurnal tides

Figures 9(a) and 9(b) show the amplitude and phase, respectively, of the diurnal tide in zonal wind for run GW in the first 16 days of July (since the modelled phase at each latitude varies slightly around the latitude circle, the vertical phase propagation at low latitudes can be better demonstrated in Fig. 9(b) by showing phases from one model longitude, rather than a zonal average). The amplitude and phase of the diurnal tide in meridional wind have similar characteristics to the zonal wind tide and, therefore, are not shown here. In Figs. 9(a) and 9(b) the relative lack of amplitude growth at high latitudes (i.e. 58°N and 58°S), combined with the lack of vertical phase propagation at these latitudes, is indicative of the (1,−2) mode. (It is common to describe tides in terms of Hough modes, (L, M), where L is the zonal wave number (equal to 1 for diurnal tides); here M = −2 implies an evanescent mode which is symmetric about the equator.) However, Fig. 9(b) also shows that vertically propagating modes are more dominant than the (1,−2) mode in parts of the middle atmosphere, i.e. in the upper mesosphere (58°N) and near the stratopause (58°S). At lower latitudes the larger growth of amplitude with height, together with a vertical wavelength of around 30 km and a clear vertical phase propagation, suggests that the wave consists chiefly of the (1,1) mode (which is vertically propagating and symmetric about the equator). Modelled diurnal tidal amplitude in both zonal and meridional wind is larger in the winter subtropics (30°S) than in the summer subtropics (30°N) in much of the mesosphere. Such results are in reasonable agreement with both radar observations (e.g. Vincent et al. 1989) and rocketsonde data (e.g. Nastrom and Belmont 1976). The overall pattern of results agrees reasonably well with both observations (e.g. Reed et al. 1969; Ahmed and Roper 1983; Manson et al. 1985) and with other modelling studies (e.g. Forbes 1982; Vial 1986; Miyahara and Wu 1989). An exception is in low latitudes, where model amplitudes decrease with height above the 0.01 mb level, whilst corresponding observed amplitudes increase with height up to about the 0.001 mb level. This is investigated further in section 4(c).

The amplitude of the zonal diurnal tide in January (Fig. 10) is broadly similar to July (Fig. 9(a)), with the exception of the summer amplitude at 58°, which increases with height in the mesosphere, reaching a maximum of about 30 m s⁻¹ near 0.003 mb, before rapidly decreasing above that level. Such a rapid increase in amplitude in the upper mesosphere has been observed by radar at Poker Flat (65°N) (Carter and Balsley 1982). It may be due to weak tide/gravity-wave interaction, since the model gravity-wave drag is much weaker near 58° in the summer mesosphere in January than in July. Another possibility is that model diffusion leads to a ‘broadening’ of the (1,1) mode into high latitudes (e.g. Forbes and Hagan 1988).

(b) Semi-diurnal tide

Figures 11(a) and 11(b) show the amplitude and phase, respectively, of the zonal semi-diurnal tide. The size and behaviour of the amplitudes at 58°S, 30°S and 3°N are in reasonable agreement with those in other models (e.g. Walterscheid and Venkateswaran 1979; Miyahara and Wu 1989). These models also show that summer amplitudes exceed those in winter in the mesosphere. This feature is also present in the EUGCM: Fig. 11(a) shows that amplitudes at 30 and 58°N exceed those at 30 and 58°S by up to 10 m s⁻¹ at 0.01 mb. Corresponding meridional tidal amplitudes are also larger in summer than in winter (not shown). These amplitudes decrease rapidly above the 0.01 mb level; this is
Figure 9. (a) Mean amplitude (in m s\(^{-1}\)) of the diurnal component of zonal wind for the first 16 days of July in run GW. Values are shown for 58°S (short-dashed line), 30°S (solid line), 3°N (dotted line), 30°N (long-dashed line) and 58°N (bold line); (b) like (a), except phase (for 180°W only) is shown (time of westerly maximum in hours).
consistent with observations made at Kiruna (68°N) (Bernard 1981), which show a
decrease in semi-diurnal amplitude between 0.01 and 0.005 mb. Figure 11(b) shows that
at 30 and 58°N (summer) there are large phase shifts between 3 and 0.5 mb, whilst at 30
and 58°S there are much weaker phase shifts near 30 and 1 mb, respectively. Phase shifts
in the meridional tide in summer are also larger than those in winter (not shown). Such
phase shifts have been reported in observations (e.g. Reed 1972) and in models (e.g.
Miyahara and Wu 1989; Miyahara et al. 1993). That the phase shift in the EUGCM takes
place at a higher level in summer than winter is in agreement with some modelling
studies. For example, Walterscheid and Venkateswaran (1979) model phase shifts near
10 mb in winter mid latitudes, but near 1 mb in summer mid latitudes. The zonal semi-
diurnal amplitudes in January (Fig. 12) are similar to corresponding values in July (Fig.
11(a)) up to about the 0.1 mb level, but above that level the difference between summer
amplitudes and those elsewhere is smaller than in July. In January, winter amplitudes
grow faster with height, whilst summer amplitudes at 30 and 58° are weaker than
corresponding values in July. The meridional semi-diurnal amplitudes in January (not
shown) are similar to those in Fig. 12. Like in July, the phase of the zonal semi-diurnal
tide in January (not shown) shifts by a larger amount in summer than in winter.

(c) Effect of diffusion on diurnal tide amplitudes

The most obvious difference between the EUGCM and other results is the behaviour
of the diurnal tidal amplitude at low latitudes. Observed amplitudes increase with height
to about 90 km, and then remain constant in the lower thermosphere, whilst the EUGCM
amplitudes increase with height only to about the 0.01 mb level (about 77 km), and then
decrease above that level (and at some latitudes there are additionally distinct maxima
and minima above the 0.01 mb level). Such effects may be related to the vertical diffusion
Figure 11. (a) Mean amplitude (in m s⁻¹) of the semi-diurnal component of zonal wind for the first 16 days of July in run GW. Values are shown for 58°S (short-dashed line), 30°S (solid line), 3°N (dotted line), 30°N (long-dashed line) and 58°N (bold line); (b) like (a), except phase is shown (time of westerly maximum in hours).
in the model. Forbes and Hagan (1988) and Forbes and Vincent (1989) have shown that the (1,1) diurnal mode, which is the dominant diurnal mode in equatorial and subtropical regions, is sensitive to vertical diffusion: increased diffusion led to weaker subtropical amplitudes, and in addition the altitude of this amplitude was lower in the higher-diffusion case. It is also possible that the decrease in the EUGCM subtropical amplitudes is related to the enhanced horizontal diffusion in the model, since the level at which the enhanced diffusion is switched on approximately matches the level above which the diurnal amplitudes decrease. In addition, the sensitivity of tidal amplitude to horizontal diffusion throughout the middle atmosphere has not been examined. Accordingly, in this section the sensitivity to model diffusion is examined by comparing subtropical diurnal amplitudes for the control run (run GW) with those from run VDF (no vertical diffusion above the 15 mb level), run HDF (different enhanced horizontal diffusion) and run LHDF (horizontal diffusion a factor of 100 weaker).

Figure 13 shows the amplitude of the meridional diurnal tide in July at 30°S. Amplitudes for runs GW and HDF are very similar above the 0.01 mb level. These results, together with a similar result obtained when the upper-level enhanced horizontal diffusion was turned off completely (not shown), indicate that the upper-level enhanced diffusion has little effect on tidal amplitude. Figure 13 also shows that the amplitude for run VDF in the middle mesosphere is slightly larger than those for the other three runs shown. However, since the model vertical diffusion depends on the Richardson number and is therefore often small, run VDF amplitudes may be different owing to small changes in the interaction between tides and the mean flow, resolved model waves and gravity waves. This may also explain the small differences between the amplitude for run LHDF and those for the other three model runs in Fig. 13, since such differences are not systematically present at other tropical and subtropical latitudes (not shown).
The above results show that the decrease in diurnal tidal amplitude near the 0.01 mb level does not appear to be related to the strength of model horizontal or vertical diffusion. Another possibility is that it is a result of spurious reflections from the top level of the model. In a comparison of wave amplitudes from two versions of a GCM with top levels of 10 and 0.1 mb, Boville and Cheng (1988) noted that amplitudes in the former model version had an approximately standing-wave structure in the vertical, with distinct maxima at 14 and 33 mb, and minima in between. This structure was absent in the latter model version. Such a structure is similar to that above the 0.01 mb level in Fig. 13 (it is also present, though more weakly, at other latitudes (see, for example, Fig. 9(a))). However, Boville and Cheng’s results may not be directly applicable to the EUGCM, since their model uses a ‘rigid lid’ upper boundary condition (i.e. vertical velocity is zero at the top model level), whereas in the EUGCM the vertical velocity is non-zero at the top level but a ‘sponge layer’ of enhanced horizontal diffusion is applied near the top of the model to damp any spurious reflections. Rasch (1986) outlined two criteria for an effective sponge layer. Firstly, the model must have sufficient vertical resolution to resolve the waves to be damped adequately (i.e. 8–10 grid points per wavelength), and secondly the sponge layer must extend over at least one vertical wavelength. For the case of tropical and subtropical diurnal tides, the first criterion is met in the EUGCM (the model vertical resolution is between 2 and 3 km in the middle atmosphere), but the second is not, since such waves have a vertical wavelength of around 30 km, whilst the model sponge layer has a depth of only 15 km. It is, however, not totally clear that this is the source of the problem, since the upper-level structure in
Fig. 13 changes little when reductions are made to the depth of the sponge layer (run HDF) or to the strength of the horizontal diffusion (run LHDF). Rigorous examination of this phenomenon would involve extending the depth of the model sponge layer, or extending the vertical domain of the EUGCM.

5. CONCLUSIONS

In this paper it is shown that removing the diurnal cycle in the control run of the EUGCM (run GW) leads to changes in the upper mesospheric equatorial zonal winds of up to 65 m s\(^{-1}\). The pattern of such zonal wind changes is in good general agreement with other modelling studies (e.g. Miyahara 1981; Miyahara and Wu 1989; Wu et al. 1989), but the size of the changes is not: zonal wind differences in the upper mesosphere of the Miyahara and Wu (1989) model are up to 50 m s\(^{-1}\) weaker than corresponding EUGCM differences near the equator, and between 5 and 15 m s\(^{-1}\) weaker in the subtropics. Such large zonal wind differences in the EUGCM cannot be attributed to poorly modelled diurnal and semi-diurnal tides since it is shown that the EUGCM tidal amplitudes and phases are generally in good agreement with both observations and other models. However, the differences in zonal wind between diurnal and non-diurnal runs change when model vertical and horizontal diffusion are changed. Zonal wind differences increase by up to 10 m s\(^{-1}\) when the vertical diffusion is turned off above the 15 mb level, and decrease by up to 20 m s\(^{-1}\) when the horizontal diffusion is increased by a factor of 100. In the former case, such zonal wind changes probably occur because the absence of vertical diffusion raises the level at which momentum deposition due to tides takes place, resulting in a concentration of EPFD near the model upper boundary. The latter result appears to be due to a weakening of momentum deposition due to tides when the horizontal diffusion is reduced, consistent with the results of Boville (1985). It is not clear whether reducing horizontal diffusion further would lead to upper mesospheric zonal wind differences as small as those found by Miyahara and Wu (1989) and Wu et al. (1989), or whether it would instead increase spurious reflection of waves from the model top. Further examination of this would require a version of the EUGCM with a higher top level.

It is also shown that EUGCM diurnal and semi-diurnal amplitudes and phases are in good general agreement with observations and other modelling studies. An exception is diurnal amplitudes in the tropics and subtropics, which decrease above the 0.01 mb level, and in some cases have a standing-wave structure above that level. This behaviour is not altered when the upper-level model enhanced diffusion is changed, or when the model horizontal diffusion is reduced by a factor of 100 at all model levels. Removing model vertical diffusion also has little influence on model diurnal amplitudes, in contrast to the results of Forbes and Hagan (1988) and Forbes and Vincent (1989), probably since the difference in vertical diffusion between runs GW and VDF is much smaller than corresponding differences in the above mentioned studies. Instead, the structure of model diurnal amplitudes above the 0.01 mb level may be caused by spurious reflections from the model top level. In particular, the depth of the model sponge layer may be too shallow. Rigorous examination of this phenomenon, beyond the scope of this paper, would involve extending the depth of the model sponge layer, or extending the vertical domain of the EUGCM.

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