



THE INFLUENCE OF TIDES ON COMPOSITION OF THE THERMOSPHERE

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ABSTRACT

This paper investigates mechanisms by which tidal oscillations in the lower thermosphere can change composition. In particular, the effects of vertical flow are investigated by distinguishing between vertical motion of pressure levels and vertical motion relative to pressure levels. It is shown that only the latter disturbs diffusive equilibrium, while the former changes composition at a fixed height but not on a pressure level, thus causing no departure from diffusive equilibrium. An estimate of diffusion times shows that molecular diffusion in the lower thermosphere cannot restore diffusive equilibrium on the time scale of diurnal and semidiurnal tides. As a result of up- and downwelling relative to pressure levels adiabatic temperature changes occur which coincide with density variations. Theoretical predictions are confirmed by runs with the Coupled Thermosphere-Ionosphere model (CTIM). These runs also successfully reproduce the thermospheric temperature-density phase anomaly.

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INTRODUCTION

Tides are global oscillations of pressure, temperature and winds with periods less than a day. They are generated through absorption of solar radiation by water vapour (near-IR) in the troposphere, by ozone (UV) in the stratosphere, and by absorption of UV and the XUV in the thermosphere [Forbes, 1995]. In the lower thermosphere, between 80 and 150 km height, in-situ forcing of tides is weak, but tides propagating upwards from the troposphere and stratosphere play an important role in the dynamics of the region. Below the turbopause (around 100 km), all gases in the atmosphere are well mixed. In the absence of eddy mixing each constituent will approach a hydrostatic equilibrium where the partial pressure of the species is balanced by its own gravitational weight. The vertical distribution of each species is then in diffusive equilibrium, implying an exponential fall of the particle density with the individual scale height, $H_i = RT/M_i g$, where $i = O, O_2$ or N_2 . In order to examine the influence of tides on composition, one needs to determine how far this equilibrium is disturbed and how quickly molecular diffusion can restore the balance. When trying to identify what mechanisms may cause changes of composition in the thermosphere and ionosphere, Rishbeth et al. [1987] suggested vertical flow, changes in photochemistry, vertical components of centrifugal and Coriolis forces and electrostatic forces between charged particles which change the scale height of ions. This paper will only discuss composition change induced by vertical flow. The dominant photochemical process affecting major species thermospheric composition is photo-dissociation of O_2 [Roble et al., 1987]. Photo-chemical lifetimes are of the order of 5-7 days so photo-chemistry will not play a significant role in tidally induced composition changes.

In order to investigate the influence of vertical flow on composition, it is necessary to distinguish between two components of vertical wind, the *barometric velocity* v_b and *divergence velocity* v_d , where the latter represents motion relative to a fixed pressure-level and the former is the motion of the pressure level itself [Rishbeth et al., 1969]. The total vertical wind v in the height coordinate system is given by their sum, $v = v_b + v_d$. While the barometric wind can be caused by thermal expansion, the divergence vertical wind is

related to the vertical wind in the pressure frame ($w=dp/dt$) by $v_d=w/\rho g$. From the continuity equation it follows that the vertical wind of the pressure coordinate system, w , is related to the divergence of the horizontal wind field:

$$\frac{\partial w}{\partial p} = -\vec{\nabla}_p \cdot \vec{V}_{horiz}$$

Therefore, the divergence wind is related to the divergence of the horizontal wind field.

THE MODEL

To demonstrate the basic theoretical predictions in this study, we use a recently modified version of the Coupled Thermosphere-Ionosphere model (CTIM) [Fuller-Rowell and Rees, 1980, Quegan et al., 1982, Fuller-Rowell et al., 1994, and references therein]. The model solves the equations of momentum, energy and continuity for neutrals (O, O_2, N_2) on pressure levels and ions (O^+, H^+) on a height system. With the model's lower boundary at 80 km height, the chemistry of the middle atmosphere is not calculated and tides formed there need to be specified as a boundary condition. Recently, this external tidal forcing in CTIM has been modified in two significant ways. First, the pressure level of forcing has been moved down from level 4 (around 98 km) to the lower boundary. More importantly, equations from classical tidal theory [Holton, 1975] have been implemented to analytically couple the semidiurnal and diurnal tidal oscillations of temperature and winds to those of pressure at the lower boundary. Since calculations are carried out on levels of constant pressure, pressure oscillations correspond to oscillations of pressure level heights. The latitudinal tidal structure at the lower boundary is given by Hough functions.

VERTICAL FLOW AND COMPOSITION

Tidal oscillations can generate both of the vertical wind components described above. While temperature changes through thermal expansion and contraction cause barometric winds, the horizontal wind gradients introduced by tides are accompanied by vertical divergence winds which are relative to the pressure levels. Both components of vertical wind have a different influence on composition. The hydrostatic distribution of constituents above the turbopause implies that ratios O/O_2 and O/N_2 are constant on any level of fixed pressure as long as diffusive equilibrium holds. Figure 1 shows the globally averaged vertical distribution of O , O_2 and N_2 , as modelled by CTIM. One can see that ratios O/O_2 and O/N_2 decreases as a function of altitude. Since barometric winds are caused by adiabatic heating, temperature variations will change the scale heights of each constituent by the same proportion, the constituent ratios are maintained on a fixed pressure level and barometric winds will not disturb diffusive balance. Only at a fixed height will the ratios change under barometric winds. In contrast, divergence vertical winds will move parcels of gas relative to pressure levels and so cause departures from diffusive equilibrium since constituent ratios change with height, as shown in Fig. 1. Molecular diffusion acts to try to restore equilibrium but its time scale can be slower than the vertical winds causing the change. Therefore, from theoretical considerations we expect that any vertical flow changes composition at a fixed height, but only vertical motion relative to levels of fixed pressure will disturb the diffusive equilibrium.

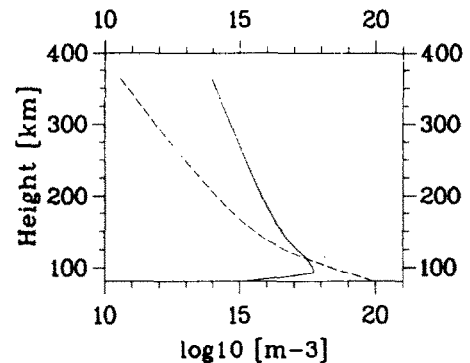


Fig.1. Globally averaged height profiles of O (solid), O_2 (dashed) and N_2 (dotted), as modelled by CTIM, in units of $\log_{10} [m^{-3}]$.

DIFFUSION TIMES

While tidally induced vertical flow will upset diffusive balance, it remains to be shown how quickly molecular diffusion can restore equilibrium. Since we focus on heights above the turbopause, eddy mixing is neglected.

This is justified by the fact that 5 km above the turbopause (set at 100km in CTIM) the eddy coefficient used in CTIM is only 1% of the molecular diffusion coefficient and vanishes completely above 150 km. The time necessary to reach diffusive equilibrium, τ_{eq} , can be estimated by $\tau_{eq} = H^2/D$ [Rishbeth & Garriott, 1969], where H is the scale height and D the diffusion coefficient. Using the molecular diffusion rates in CTIM (taken from [Colegrove et al, 1966]), values of τ_{eq} can be calculated, as shown in Table A for globally averaged conditions.

Height	105km	115km	125km	140km	160km	180km	200km	>200km
τ_{eq}	410h	150h	60h	24h	10h	5h	2h	<1h

Table A: Time scale for molecular diffusion

The table shows that time scales necessary to reach diffusive equilibrium are larger than the periods of tides for heights below around 160 km. So molecular diffusion is unable to restore equilibrium fast enough. Only above 160km does diffusion becomes effective and also ultimately contribute towards the effective dissipation of upwards propagating tides.

MODEL RESULTS

In order to investigate whether CTIM is capable of reproducing the above theoretical results, two runs were carried out, for equinox conditions, using F10.7 solar flux of 85 and Kp=3. At the lower boundary, semidiurnal tidal (2,4) forcing was applied in both runs with an amplitude of 200m geopotential height, setting the phase of local time maximum to 12.0h. In order to investigate the influence of horizontal wind gradients, however, in one of the runs horizontal wind gradients were artificially set to 0. In Fig.2a and b, the dashed lines indicate the height of pressure level 5 (0.019 Pa) and the solid lines are the height of constant $O/O_2 = 0.5$ ratio. While Fig.2a shows results from a normal run, Fig.2b shows the case where horizontal wind gradients were set to 0. The different amplitudes and phases of curves in Fig. 2a show that on levels of constant pressure the O/O_2 ratio is not constant. This demonstrates not only that tides disturb the diffusive equilibrium but also that molecular diffusion is unable to restore equilibrium on the timescale of semidiurnal tides. In Fig.2b, however, the O/O_2 ratio is constant on the pressure level. With the earlier findings about diffusion times we may conclude that no departure from diffusive equilibrium occurs when the only vertical motion is that of the pressure levels themselves. In Fig.3, the temperature of pressure level 5 (solid) is plotted with the temperature at fixed height of 107.3 km (dotted) from the run in which horizontal wind gradients were set to zero. The plot shows that temperature varies diurnally on the pressure level and semidiurnally at the fixed height. In contrast, the temperature on that pressure level is strongly semidiurnal when horizontal wind gradients and divergence vertical wind are non-zero, as shown in Fig.4 (solid line). This shows that temperature on a pressure level is influenced by solar heating (giving primarily a diurnal component) and by adiabatic heating (semidiurnal component) which is caused by up-and downwelling relative to the pressure

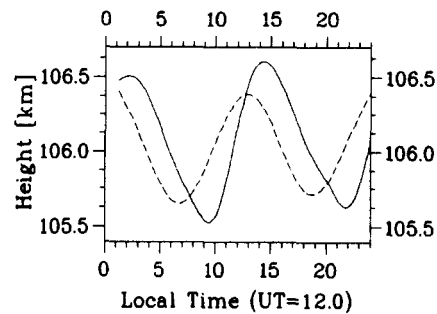


Fig.2a. Height of pressure level 5 (dashed) and constant $O/O_2 = 0.5$ level (solid line) at $\text{lat}=0^\circ$, using (2,4) forcing only, for normal conditions.

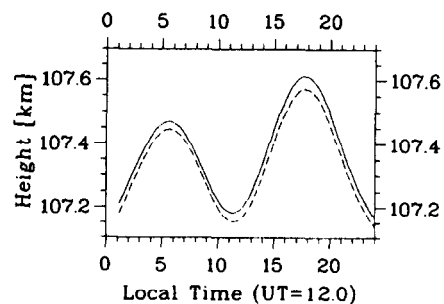


Fig. 2b. Same as Fig.2a, but with horizontal wind gradients set to 0.

level. The dotted curve in Fig.4 is the difference of both curves in Fig.2a. It is a measure of the change of O/O_2 ratio on the pressure level and thus for the degree of departure from diffusive equilibrium. The plot shows that strongest departures from diffusive equilibrium coincide with temperature extrema and confirms that the temperature changes on the pressure level are directly linked to up- and downwelling. The absence of divergence winds in the run from Fig.3 implies that no adiabatic heating on the pressure level occurs, and thus no semidiurnal component. The solar heating will increase with altitude and eventually become stronger than the adiabatic heating from tides, resulting in a primarily diurnal temperature behaviour. Above 200km, most of the tides propagating upwards from the middle atmosphere will have dissipated, thus further reducing the semidiurnal component.

It was shown earlier that diffusion times become shorter than the times of tidal oscillations above 160 km height. Here, departures from diffusive equilibrium will therefore lead to transport of constituents. Mayr et al. [1973] showed that wind-induced diffusion in the thermosphere leads to transport of atomic oxygen towards earlier local times. They found this redistribution of O to be one of the mechanisms causing the *thermospheric phase anomaly* in which temperature maxima occur between 1 and 3 hours after the O density maxima. This anomaly is successfully reproduced by CTIM above 170 km height.

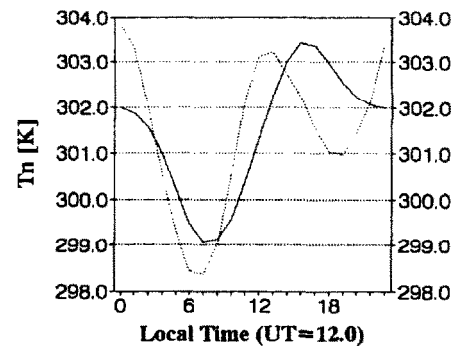


Fig.3 Temperature on p-level 5 (solid) and at fixed height of 107.3 km (dotted), with horizontal wind gradients set to 0.

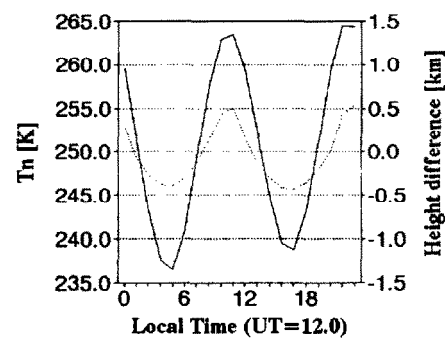


Fig.4 Temperature on p-level 5 (solid) and difference between curves in Fig.2a (dotted), with non-zero horizontal wind gradients.

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