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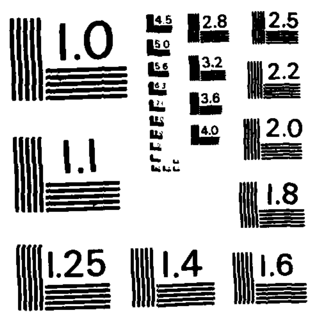

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GLOBAL ATMOSPHERIC WAVES

Mechanical heating of the lower thermosphere

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20. Abstract: Calculations are presented of the flux of mechanical energy into the lower thermosphere from the leading migrating modes of diurnal and semidiurnal atmospheric oscillation. The thermal generation of these modes is evaluated for mean January, April, July and October conditions using revised profiles of water vapour and ozone heating and the resulting vertically upward energy fluxes are evaluated using classical tidal theory. For the diurnal oscillation, the calculations confirm an earlier analysis (Lindzen, 1967) that tidal energy dissipation may be a significant factor in thermospheric heating. Also the relative changes of flux between the above four months are noted to be similar in form to the semi-annual variation of thermospheric air densities. For the semidiurnal oscillation the energy flux is noticeably smaller (by a factor of about three) than earlier estimates (Lindzen and Blake, 1970) due mainly to the revision of ozone densities and an approximate calculation indicates that the semidiurnal tide is a relatively minor source of thermospheric heating.

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Energy fluxes of the (1,1,1) atmospheric oscillation

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Abstract

The global mean vertical energy flux of the (1,1,1) mode of atmospheric oscillation is evaluated at 80 km altitude by classical tidal theory for mean January, April, July and October conditions using revised profiles of water vapour and ozone heating. Fluxes calculated for January and July are lower than those for April and October due to seasonal changes in water vapour, solar declination and Sun-Earth distance. Flux values obtained are compared with a previously stated requirement for maintaining the residual thermosphere and are adequate unless damping, which is ignored in the present calculations, introduces a factor of more than an order of 10 in magnitude. The relative changes of flux between the above four months are noted to be similar in form to the semi-annual variation of thermospheric air densities.

1. Introduction

The upward energy flux propagated into the lower thermosphere from below by diurnal atmospheric oscillations was calculated by Lindzen (1967) to be about half the downward flux of solar radiation absorbed above 105 km when averaged globally. Although many approximations are involved in such a calculation, the result indicates that tidal energy dissipation may be a significant factor in thermospheric heating.

In a similar analysis for the (1,1,1) oscillation the separate wave contributions excited by water vapour and ozone absorption of solar radiation were examined and it was found (Groves, 1976) that the amplitude of the wave excited by water vapour absorption was reduced by 75 per cent by a wave of opposite phase excited by ozone absorption. Within the limits of accuracy of such evaluations, the waves excited by water vapour and ozone absorptions were possibly of comparable magnitude and opposite phase, in which case the upward energy flux could be greatly varied by small relative changes in heating between the two absorbing regions arising for example from the seasonal meridional movement of water vapour.

A detailed examination of diurnal heating was therefore undertaken (Groves, 1977) based on a more comprehensive model of specific humidity, in which the dryness of the upper troposphere was realistically modelled and ozone heating reduced in accord with observational data on ozone densities from Park and London (1974). A further refinement was the change from a constant scale height to a realistic scale height profile for the basic atmosphere. The revised calculations were carried out for mean January, April, July and October conditions. The amplitude of the (1,1,1) oscillation that would be excited by water vapour absorption alone was then found to be reduced by 38 per cent instead of 75 per cent by the presence of the oscillation excited by ozone absorption. With this lower value for the reduction, the idea of a highly variable energy flux arising from two nearly cancelling waves became rather unlikely. The main seasonal variation shown by the results was a 30 per cent lower value in July than at other times, which arose from the increase in Sun-Earth distance and solar declination and the meridional displacement of water vapour into the N hemisphere: for January, the effect of the net meridional displacement of water vapour into the S hemisphere was largely counterbalanced by the reduced Sun-Earth distance. The results of the analysis therefore failed to approximate to a semi-annual type of variation and hence to indicate on these grounds at least, any direct relationship between tidal heating and the well-known semi-annual variation of thermospheric air densities (CIRA, 1972). In two other respects however the results (Groves, 1977) were not incompatible with the thermospheric phenomenon:

(i) energy flux magnitudes were apparently adequate to maintain the residual thermosphere, i.e. the thermosphere after removal of solar heating, and (ii) the 30 per cent July decrease of energy flux was of the same magnitude as the July air density decrease found in the semi-annual variation.

The present paper continues the investigation of the energy flux propagated by the (1,1,1) oscillation and its possible relevance to thermospheric heating. The analysis is based on revised calculations of water vapour and ozone heating (Groves, 1982a,b), but the method of calculating the energy flux is the same as before (Groves, 1977), i.e. by classical tidal theory, no account being taken of dissipation or background winds.

2. (1,1,1) components of water vapour and ozone heating

Detailed accounts of the calculation and evaluation of Hough components of water vapour and ozone heating for mean January, April, July and October conditions have previously been given (Groves, 1982a,b). A significant modification to tropospheric heating rates was made by the introduction of cloud-related scattering, the effect of which has been to reduce heating in the lower troposphere and increase it at mid-tropospheric heights. The continuous lines in Fig. 1 show the heating profiles previously reported which are based on the absorption coefficients of Somerville et al. (1974).

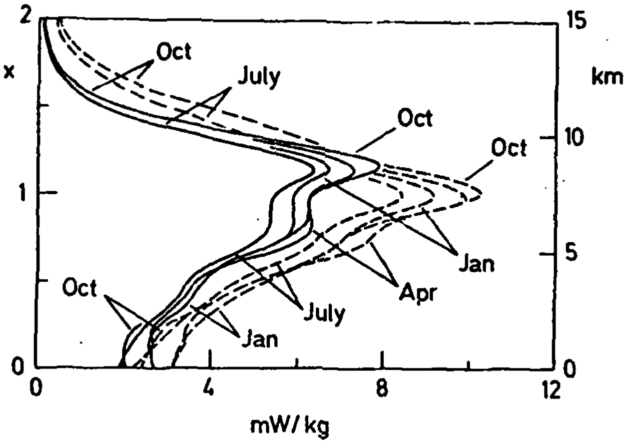

Fig. 1 Heating rate profiles for the (1,1,1) mode due to water vapour absorption. Key: —— from Groves (1982b) and based on the absorption coefficients of Somerville et al. (1974); ----- evaluated with the absorption coefficients of Kerschens et al. (1976). x = ln(surface pressure ÷ pressure). Heights correspond to a constant scale height of 7.5 km.

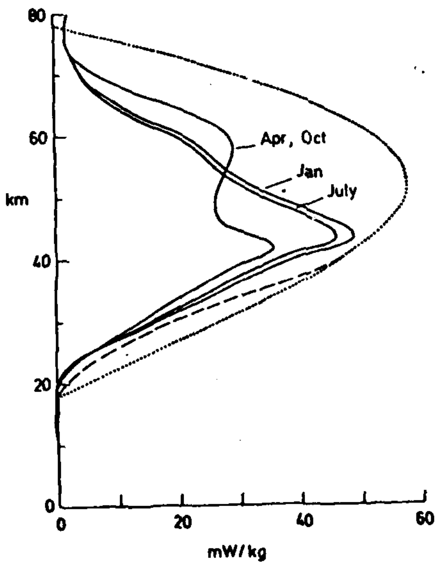

Fig. 2 Heating rate profiles for the (1,1,1) mode due to ozone absorption. Key: ...... Lindzen (1967); ----- Groves (1977); —— Groves (1982a).

Many sets of water vapour absorption coefficients have been published in recent years and, for some of these, maximum heating rates are calculated which are 30 per cent greater than those obtained with the coefficients of Somerville et al. (1974). The energy flux calculations have therefore been carried out for both sets of profiles in Fig. 1, the second set being based on the absorption coefficients of Kerschens et al. (1976).

Fig. 2 shows a comparison between the Hough components of ozone heating (Groves, 1982a) and earlier equinox profiles. The dotted curve is from Lindzen (1967) and was adopted for examining the separate tropospheric and stratospheric contributions (Groves, 1976). The dashed curve is the modification introduced below 40 km in the subsequent re-examination (Groves, 1977). The latest profiles have been derived from latitudinal ozone cross-sections based on observed densities, and below 40 km they extend the previous modification to still lower heating rates. Above 40 km, values are lower than the earlier ones by a factor of more than two at some heights.

3. <u>Contributions of tropospheric and stratospheric heating to the (1,1,1) oscillation at 80 km</u>

The linearization employed in classical tidal theory enables the oscillation at any given height to be expressed as a superposition of incremental oscillations arising from different levels of heating. In the present calculations 80 km is the upper limit of the region of thermal excitation and it is at this height that we evaluate the oscillation and thence the vertical energy flux. The evaluation has previously been undertaken of the (1,1,1) thermal response weighting function $W_t(z)$ which weights the (1,1,1) component of heating at height z according to its contribution to the oscillation at 80 km (Groves, 1975, 1982c). $W_t(z)$ may be calculated from Equ. 11.8 of Groves (1982c). Fig. 3 shows $W_t(z)$ for this mode on an arbitrary scale. The values of pressure scale height adopted at 0(2)26 km are 8.75, 8.46, 8.14, 7.82, 7.53, 6.98, 6.63, 6.22, 5.90, 5.99, 6.14, 6.24, 6.39, 6.56 km and at 28(2)150 km values are taken from the mean CIRA (1972).

A characteristic feature of the (1,1,1) mode which is apparent in Fig. 3 is its oscillatory nature with a wavelength of roughly 25 to 30 km. A second feature of $W_t(z)$ (Fig. 3) is the exponential-like decay of the function with height, which arises as a consequence of the decrease of atmospheric density with height and the associated increase of wave amplitudes with upward propagation. Larger amplitudes are therefore generated at 80 km by lower levels of thermal excitation and accordingly larger values of $W_t(z)$ are found at lower heights. In the troposphere $W_t(z)$ is almost entirely of one sign, whereas at stratospheric/mesospheric heights there are two regions of positive values and two regions of negative values. As radiational heating has the same phase at all heights, maximizing at local noon, the positive and negative contributions to the oscillation at 80 km arising from the thermal excitations at different heights below 80 km are directly additive. The summation involves the integral

$$F(z) = \int_0^z W_t(\zeta)\, J(\zeta)\, d\zeta \qquad (1)$$

where J is the heating rate profile (Figs. 1 and 2). For $z \geqslant z_S = 80$ km, we have $J(z) = 0$ and $F(z) = F(z_S)$.

By Equs. 11.3 to 11.8 of Groves (1982c) the oscillation (of any atmospheric parameter) at height $z \geqslant z_S$ has $F(z_S)$ as a factor (denoted by $\ell_J \Phi(0)$ in Groves (1982c)), and the seasonal variation of the oscillation at this height arising from the seasonal changes of J, as shown for example in Figs. 1 and 2, is proportional to the corresponding variation of $F(z_S)$.

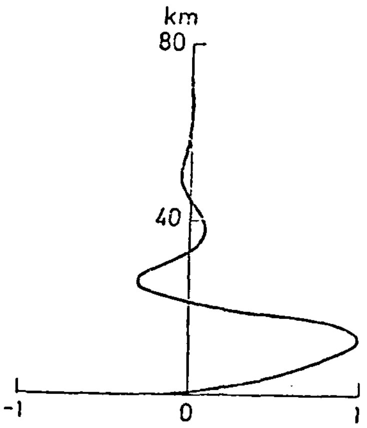

Fig. 3 Thermal response weighting function $W_t(z)$ for the (1,1,1) mode plotted on an arbitrary scale along the abscissa (Groves, 1982c).

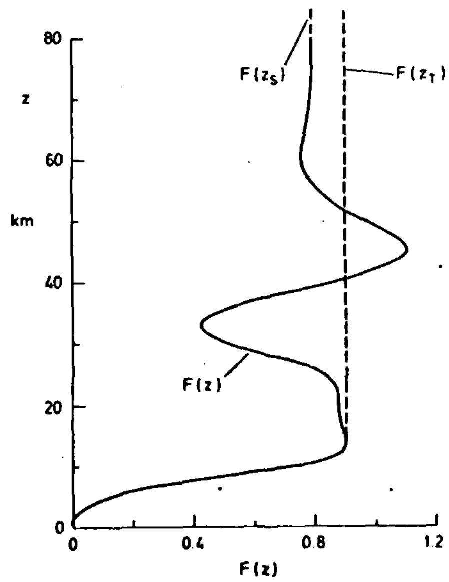

Fig. 4 F(z) (equation 1) plotted on an arbitrary scale. z_T is the upper height limit of tropospheric heating. z_S = 80 km, the upper height limit of stratospheric/mesospheric heating. The heating rate profile, J, used in equn. 1 is for January.

$F(z)$ is shown in Fig. 4 on an arbitrary scale for the January heating rate profile, the tropospheric heating rates being those based on the absorption coefficients of Somerville et al. (1974). Of particular interest is the value of $F(z_S)$ in relation to that of $F(z_T)$, where z_T is the upper limit of tropospheric heating. Fig. 4 shows that $F(z_S) = 0.876\ F(z_T)$, i.e. wave amplitudes at 80 km due to tropospheric heating are reduced by 12.4 per cent by the excitation due to ozone heating. The corresponding reductions for the three other months are between 9 and 14 per cent. These reductions are smaller than the value of 38 per cent previously calculated (Groves, 1977) mainly on account of the lower ozone heating rates now employed (Fig. 2). If the water vapour absorption coefficients of Kerschens et al. (1974) are adopted, the corresponding reduction is 9.1 per cent for January. The values in the three other months lie in the range 8 to 10 per cent. The choice of absorption coefficients does not therefore affect the general conclusion that the reduction of amplitude of the tropospherically excited wave at 80 km by that due to ozone heating is small in all seasons.

Fig. 5 shows the values obtained for $F(z_S)$ and $F(z_T)$ in January, April, July and October. $F(z_S)$ is lower in January and July than April and October when the values are nearly equal; the decrease in January being about half that in July.

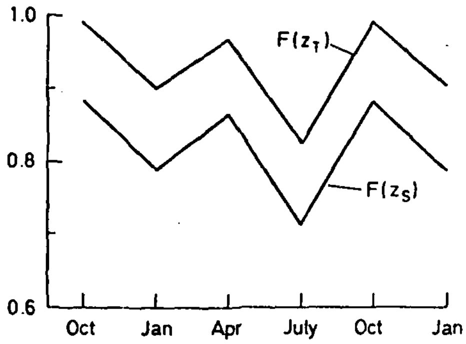

Fig. 5 Monthly mean values of $F(z_T)$, $F(z_S)$ for January, April, July and October.

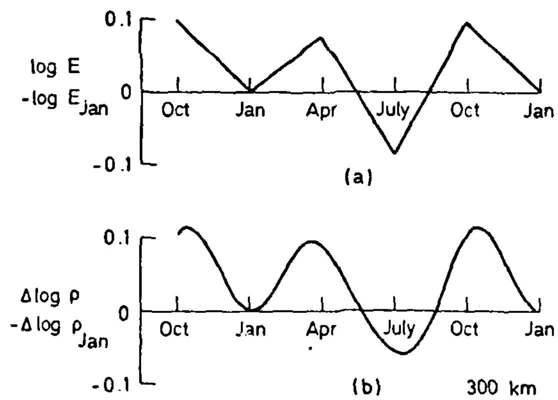

Fig. 6 (a) E, the globally-averaged (1,1,1) undamped energy flux into the lower thermosphere. The values shown are based on the absorption coefficients of Somerville et al. (1974), but are not affected by more than the thickness of the plotted line if those of Kerschens et al. (1976) are taken.

(b) $\Delta \log \rho$, the CIRA (1972) semi-annual variation in log(air density) at 300 km.

Fig. 5 shows that the seasonal variation in $F(z_S)$ arises almost entirely in the troposphere, i.e. from $F(z_T)$, being slightly augmented by a very small variation in the stratospheric contribution, $F(z_T) - F(z_S)$. Data of a seasonal nature that have been introduced into the calculation of tropospheric heating are the global distributions of cloud amount and specific humidity, the solar declination and the Sun-Earth distance. The greater distance of the Sun in July than January would account for most of the decrease of $F(z_T)$ in July with respect to the January value, leaving decreases of approximately the same amount in January and July relative to April and October to be accounted for by other effects. The effect of the seasonal variation in cloud amount has been investigated by recalculating $F(z_T)$ with the same global distribution of cloud for each of the four months (taken equal to the average for the four seasonal distributions). The resulting changes in $F(z_T)$ are found to be inconsequential, being less than 0.7 per cent. The roughly equal decreases in $F(z_T)$ in January and July relative to April and October therefore arise in these calculations from the change in the adopted distributions of specific humidity in conjunction with that of the solar declination between the respective months.

4. Mean (1,1,1) energy flux E at 80 km

E denotes the globally-averaged value of the time-averaged vertical energy flux at 80 km. It is evaluated here for the (1,1,1) mode by classical tidal theory using Equ. 11.9 of Groves (1982c) with the Newtonian cooling constant taken as zero. The continuous line in Fig.6a joins values obtained for log E/E_{Jan}, where E_{Jan} is the January value of E, the absorption coefficients employed being those of Somerville et al. (1974). As $E \propto F(z_S)^2$, Fig.6a is readily constructed from Fig. 5. With the absorption coefficients of Kerschens et al. (1976), Fig. 6a is not affected by more than the thickness of the plotted line. On the other hand E_{Jan} is highly dependent on the choice of absorption coefficients, values of 1.34 and 2.68 mW/m^2 being obtained with the absorption coefficients of Somerville et al. (1974) and Kerschens et al. (1976) respectively. The factor of 2.00 difference between these values can reasonably be attributed to the difference between the two sets of heating profiles in Fig. 1 and the dependence of energy flux on the square of wave amplitude. These values considerably exceed the previous estimate of 0.4 mW/m^2 for the undamped flux at 80 km (Groves, 1977). Higher values are now obtained due to (i) the smaller reduction in wave amplitude by ozone excitation, i.e. 8 to 14 per cent instead of 38 per cent, and (ii) the closer correspondence of the revised troposphere heating profiles (Fig. 1) with the weighting function W_t (Fig. 3) as a result of the inclusion of cloud-related scattering of solar radiation.

The rate of heat input above 100 km needed to maintain a residual exospheric temperature of 500 K, corresponding to a zero level of solar activity, has been estimated as 0.1 mW/m^2 (Volland, 1969). The above estimates of undamped energy flux indicate that tidal energy could adequately supply the required heating, provided the flux reaching the region of 100 km has not been reduced by more than an order of 10 in magnitude due to dissipative processes below 100 km. In the absence of more detailed calculations we refer to those of Lindzen and Blake (1971) in which (1,1,1) amplitudes were reduced by a factor of about 2 at 105 km due to viscosity, eddy conductivity and radiative damping with respect to undamped values: for energy flux the corresponding reduction would be by a factor of about 4. When taken with the above values of E (either 1.34 or 2.68 mW/m^2) an energy flux into the lower thermosphere is indicated which is more than adequate to supply the heating requirement estimated for maintaining the residual th^rmosphere (0.1 mW/m^2).

A feature of the (1,1,1) energy flux shown by Fig. 6a is a semi-annual type of variation. With only four determinations in January, April, July and October the yearly cycle is obviously not well defined, but these four months are expected to be those of maximum or minimum heating with respect to solar distance, declination and water vapour movement and should therefore indicate the general nature of the yearly cycle. Fig. 6b shows the variation in in log(air density) relative to the January.value in the thermosphere (CIRA, 1972) and a high degree of similarity with Fig.6a is evident.

5. Discussion

Previously calculated height profiles of the (1,1,1) Hough components of heating due to water vapour and ozone absorption for mean January, April, July and October conditions (Figs. 1 and 2) have been used to calculate by classical tidal theory the globally averaged flux of (1,1,1) energy propagating upwards at 80 km in a non-dissipating atmosphere. The results obtained (Fig. 6a) show that fluxes for January and July are lower than those for April and October, the relative decrease for January being about half that for July.

Consideration has been given to the origin of the changes of flux between the respective dates in terms of the adopted data. The value for July is lower than for January on account of the greater Sun-Earth distance and January and July values are otherwise lower than those for April and October due to changes in the global distribution of water vapour in conjunction with the change of solar declination: the effect on the energy flux of changes in ozone heating between the respective dates is found to be relatively inconsequential.

Calculations carried out for two different sets of water vapour absorption coefficients (Fig. 6a) show the relative changes of flux for the four dates to be insensitive to the particular absorption data adopted. On the other hand the values obtained for the undamped flux differ appreciably, being 1.34 and 2.68 mW/m^2 for the two sets of data. The extent to which these values would be reduced by damping below, say 105 km, is uncertain, but previous calculations (Lindzen and Blake, 1971) indicate a factor of four in which case the present work yields a flux that is still well in excess of the energy requirement of 0.1 mW/m^2 estimated by Volland (1969) for maintaining a residual exospheric temperature of 500°K.

Compared with previous calculations (Groves, 1977), the yearly cycle defined by January, April, July and October evaluations now shows a much closer similarity with that of the semi-annual variation of thermospheric air densities (Fig. 6). The present calculations are however based on a number of simplifying assumptions such as the neglect of background winds which might also contribute to the yearly cycle of the vertical energy flux, and further analysis is considered necessary before any definite conclusion might be reached on the physical origin of the semi-annual variation of thermospheric air densities.

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Thermospheric energy flux of the semidiurnal tide

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Abstract

The upward energy flux of the (2,2,2) mode of atmospheric oscillation generated by water vapour and ozone radiational heating is calculated at 125 km for mean January, April, July and October conditions. The values obtained for the global mean flux lie close to 0.05 mW/m^2 with a small reduction in July amounting to 13% of the average for the other three months. The effect of semidiurnal tidal heating on exospheric temperature is discussed with reference to the earlier work of Lindzen & Blake (1970) and it is concluded that the semidiurnal tide causes a relatively small increase in exospheric temperature of $\sim$33 K.

1. Introduction

Dissipation of most tidal modes occurs below 130 km with the notable exception of the leading semidiurnal mode, i.e. the (2,2,2) mode, which is dissipated over a wide range of heights primarily above 130 km and should therefore be relatively effective in heating the thermosphere (Lindzen & Blake, 1970). On the basis of previous detailed calculations (Lindzen, 1970), Lindzen & Blake (1970) took equatorial values of 0.3 and 0.4 mW/m^2 for the time-averaged upward energy flux of the (2,2,2) mode (excited by water vapour and ozone radiational heating) and showed using the one-dimensional heat conduction equation that exospheric temperatures of about 600-700 K could be maintained without EUV heating. Their calculations also showed that with a noon EUV flux of 1.4 mW/m^2 the exospheric temperature would be increased by more than 200 K to just over 1300 K by the presence of a 0.4 mW/m^2 tidal energy flux.

Evaluations of the semidiurnal water vapour and ozone tidal excitations have recently been undertaken (Forbes & Garrett, 1978; Walterscheid et al., 1980; Groves, 1982a,b) which significantly differ from the earlier results upon which the above values of energy flux (0.3 - 0.4 mW/m^2) were based. A re-evaluation of the (2,2,2) energy flux entering the lower thermosphere therefore appears to be in order and in the present paper this is undertaken using the calculated heating rates of Groves (1982a,b). The procedure followed is similar to that by which (1,1,1) energy fluxes have been evaluated (Groves, 1982c).

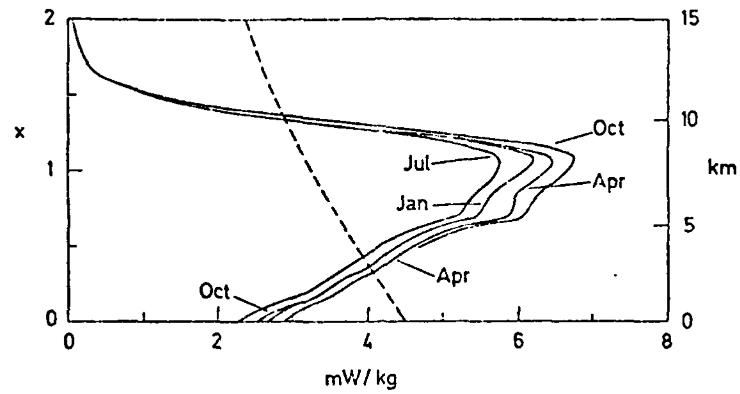

Fig. 1 Heating rate profiles for the (2,2,2) mode due to water vapour absorption. Key —— from Groves (1982a); ---- from Siebert (1961). x = ln(surface pressure ÷ pressure). Heights correspond to a constant scale height of 7.5 km.

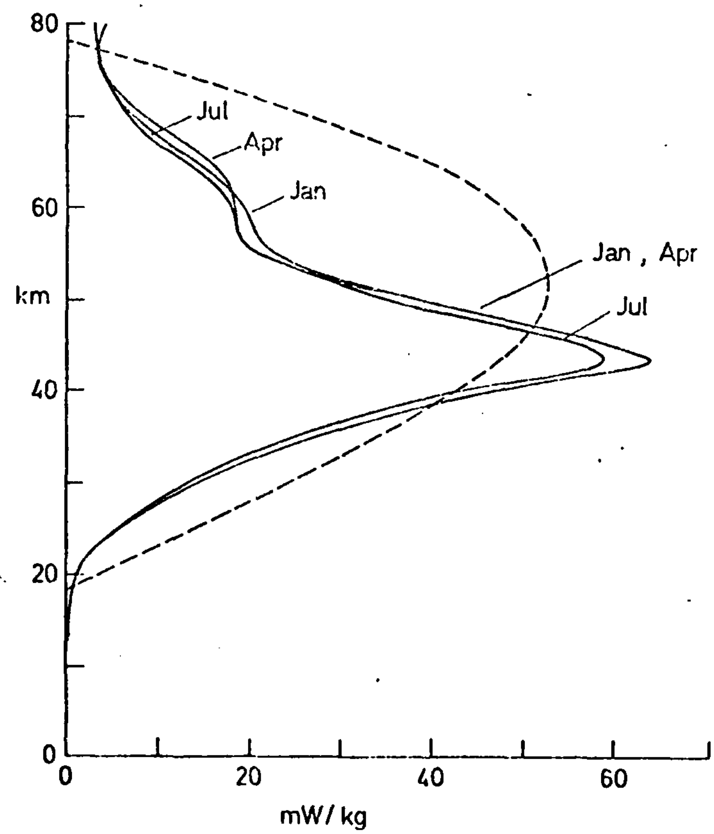

Fig. 2 Heating rate profiles for the (2,2,2) mode due to ozone absorption. Key —— from Groves (1982b); ---- from Chapman and Lindzen (1970).

2. Water vapour and ozone heating

Fig. 1 shows the (2,2,2) heating rate profile for water vapour absorption evaluated for mean January, April, July and October conditions (Groves, 1982a). The dashed curve in Fig. 1 is the earlier evaluation of Siebert (1961) which was adopted by Chapman & Lindzen (1970) and Lindzen & Blake (1971). The lower values now evaluated for the upper troposphere arise from a more realistic representation of the dryness of the upper troposphere; the higher values for the mid troposphere arise from the inclusion of cloud-related scattering; and the lower values nearer to the surface arise from the consequent shielding effect of the clouds.

Corresponding plots for ozone heating are shown in Fig. 2. October values are not shown being nearly the same as the April values (Groves, 1982b).

3. Contributions of water vapour and ozone heating to the (2,2,2) oscillation at 80 km

We take z_S = 80 km as the upper height of thermal excitation J of the (2,2,2) mode, this being the quantity plotted in Figs. 1 and 2, and evaluate

$$F(z) = \int_0^z W_t(\zeta)\, J(\zeta)\, d\zeta \qquad (1)$$

where $W_t(\zeta)$ weights $J(\zeta)$ according to its contribution in the interval $(\zeta, \zeta + d\zeta)$ to the oscillation at a given higher level, e.g. at 80 km. By Equs. 11.3 to 11.8 of Groves (1982d)

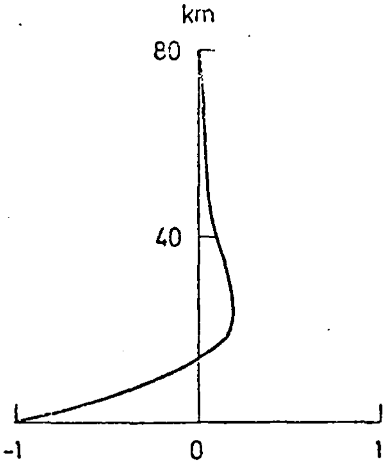

Fig. 3 $W_t(z)$ for the (2,2,2) mode plotted on an arbitrary scale along the abscissa.

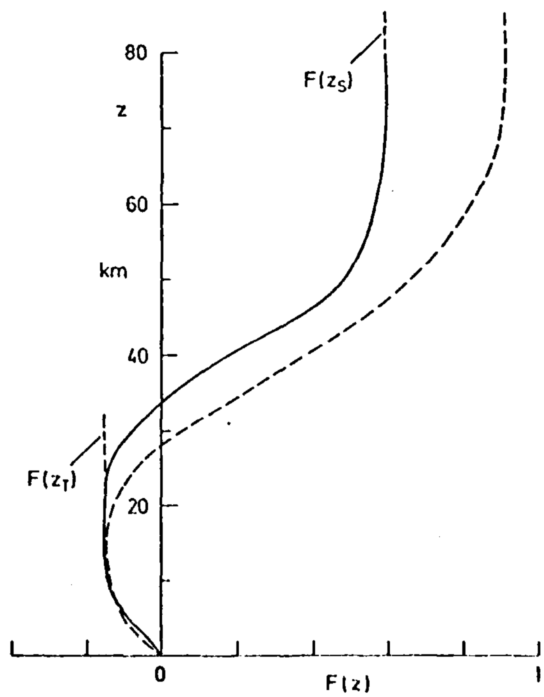

Fig. 4 F(z) (equation 1) plotted on an arbitrary scale. z_T is the upper height limit of tropospheric heating. z_S = 80 km, the upper height limit of stratospheric/mesospheric heating. Key: —— based on the January profiles in Figs. 1 and 2; ---- based on the dashed profiles in Figs. 1 and 2.

the oscillation (of any of the main atmospheric parameters) at a given height $z \geqslant z_S$ has $F(z_S)$ as the J-dependent factor. $F(z_S)$ is equal to $- \ell_J \Phi(0)$ in Groves (1982d). By examining $F(z)$ for $0 \leqslant z \leqslant z_S$ the relative contributions of water vapour and ozone heating to the (2,2,2) oscillation at 80 km (or indeed at any greater height) may be assessed.

Fig. 3, which is taken from Groves (1982d), shows $W_t(z)$ on an arbitrary scale for a non-dissipative atmosphere having pressure scale height values at 0(2)26 km of 8.75, 8.46, 8.14, 7.82, 7.53, 6.95, 6.63, 6.22, 5.90, 5.99, 6.14, 6.24, 6.39, 6.56 km and at 28(2)150 km from the mean CIRA (1972). From Fig. 3 we see that W_t changes sign at 14 km which is a height intermediate to the regions of water vapour and ozone heating, and therefore the contributions of these two regions of heating to $F(z_S)$ will be of opposite sign. This result is apparent in Fig. 4 where the continuous line shows the January profile of $F(z)$ on an arbitrary scale; and furthermore we see that the tropospheric contribution $F(z_T)$ is, in absolute magnitude, a small fraction (about 20%) of the stratospheric contribution $F(z_S) - F(z_T)$. The tide propagating into the thermosphere from below is therefore primarily generated by ozone heating and suffers a partial reduction from water vapour heating. The $F(z)$ profiles calculated for April, July and October lie too close to the January one to be conveniently plotted in Fig. 4.

The main source of error in calculating water vapour heating and hence $F(z_T)$ is considered to be the data for water vapour absorption. The results given here are based on the absorption coefficients of Somerville et al. (1974), whereas heating rates based on alternatively available coefficients may be about 30% greater (Groves, 1982a). The effect of such uncertainties on $F(z_S)$ is however diminished by the predominance of the ozone heating: for example, a 30% increase in $-F(z_T)$ would lead to a change in $F(z_S)$ of about 8%. This situation contrasts with that of the (1,1,1) mode for which water vapour heating provides the major contribution to $F(z_S)$ (Groves, 19.. c).

The main source of error in calculating the ozone heating arises from uncertainties in ozone densities particularly above 50 km, where individual observations at the same latitude may differ by a factor of two or more. Such variations have a relatively small effect on $F(z_S)$ as above 50 km both $W_t(\zeta)$ and $J(\zeta)$ in (1) are tailing away to nearly zero: for example, if J is increased by 25% at 50-60 km and by 50% at 60-80 km, $F(z_S)$ increases by only 6%.

Finally $F(z)$ has been evaluated for the earlier heating profiles (the dashed lines in Figs. 1 and 2) and the result is the dashed line in Fig. 4. The increased value obtained for $F(z_S)$ with the earlier heating arises from the higher ozone heating rates (Fig. 2). As upward energy flux depends on $F(z_S)^2$, it immediately follows that a much smaller energy flux will be obtained with the new heating rates than that previously calculated.

4. Mean (2,2,2) energy flux $\bar{E}$ entering the lower thermosphere

Values of $\bar{E}$, the global mean of the time-averaged upward energy flux have been calculated using equ. 11.9 of Groves (1982d) for a non-dissipative atmosphere with the pressure scale height profile defined in Section 3. Above 105 km, the values obtained for $\bar{E}$ vary by only a few per cent with height as the atmosphere becomes non-reflective to the (2,2,2) mode. The actual height above 105 km at which $\bar{E}$ is evaluated is therefore inconsequential: the values in Table 1 were calculated at 125 km. We shall also refer to E_o, the equatorial time-averaged energy flux. We have

$$E_0 = 2\Theta_0^2 \bar{E} = 2.33 \bar{E} \tag{2}$$

where Θ_o is the equatorial value of the (2,2,2) Hough function Θ. Equ. 2 follows from the definition of $\bar{E}$, the latitudinal dependence of tidal energy flux on Θ^2 and the normalization $\int_{-1}^{1} \Theta^2 \, d\mu = 1$, where μ is the cosine of colatitude.

Table 1 shows that the energy fluxes evaluated for January, April, July and October are less by a factor of about 2.4 than the earlier result. An equatorial flux of about 0.12 mW/m^2 at the equinoxes (April and October values) is

Table 1. Mean (2,2,2) energy flux $\bar{E}$ entering the lower thermosphere and the corresponding equatorial value E_o calculated for the profiles in Figs. 1 and 2. Units are mW/m^2.

	Heating rate profiles from Figs. 1 and 2					
	Jan	Apr	Jul	Oct	Dashed line	Lindzen and Blake (1970)
$\bar{E}$	0.051	0.050	0.045	0.053	0.12	-
E_o	0.119	0.117	0.104	0.123	0.28	0.3-0.4

now predicted compared with 0.28 mW/m^2 which is obtained if the present analysis is applied to the earlier heating. This latter value compares satisfactorily with the lower end of the range of possible values (0.3-0.4 mW/m^2) taken by Lindzen & Blake (1970) who used the same heating but not the same basic profile of atmospheric scale height. The actual result for the energy flux depends on the assumed scale height profile, and it was for this reason that Lindzen & Blake (1970) took values of 0.3 and 0.4 mW/m^2 to correspond with reasonable variations in the basic atmosphere.

5. Annual variation of $\bar{E}$

As $\bar{E}$ has been evaluated for only January, April, July and October its yearly cycle is not well defined, nevertheless these times of the year include periods of maximum and minimum heating and should give a good indication of the yearly cycle. The main conclusion to be drawn from Table 1 is the absence of any appreciable annual variation in $\bar{E}$: a small reduction in July is shown amounting to about 13% of the average of the January, April and October values which lie comparatively close to one another. The yearly cycle of $\bar{E}$ therefore differs appreciably from that for the (1,1,1) mode which shows a semi-annual type of variation with July some 30% lower than April or October (Groves, 1982c).

6. Discussion

The energy flux propagating into the lower thermosphere in the (2,2,2) mode due to water vapour and ozone heating has been estimated to have an equatorial value of close to 0.12 mW/m^2 with only a small annual variation (Table 1). This value is noticeably smaller than the values of 0.3-0.4 mW/m^2 taken by Lindzen & Blake (1970), the reduction arising mainly from the adoption of new ozone heating (Groves, 1982b).

By means of the one-dimensional heat conduction equation Lindzen & Blake (1970) found that an upward tidal energy flux of 0.4 mW/m^2 would increase exospheric temperature by more than 200 K to about 1300 K for a noon EUV flux of 1.4 mW/m^2. On the basis of their calculation we estimate, without giving details here, that a tidal energy flux of 0.12 mW/m^2 with the same EUV flux would increase the exospheric temperature by 78 K. Such estimates apply to the equator and have neglected latitudinal heat conduction which would result in a much smaller global increase of exospheric temperature. If we take a tidal energy flux of 0.03 mW/m^2 which is representative of the global mean energy flux (Table 1), we obtain by the same one-dimensional theory an exospheric temperature increase of 33 K. Although this is only an approximate calculation, it indicates that the semidiurnal tide is a relatively minor source of thermospheric heating.

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