Internal Gravity Waves in Atmospheres with Realistic Dissipation and Temperature Part III. Daily Variations in the Thermosphere

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Abstract—Using the procedures developed in Parts I and II we examine the daily variation of the thermosphere. The daily variation is found to consist in a strong semidiurnal oscillation (amplitude of temperature oscillation (100°K–200°K) excited primarily by ozone heating in the mesosphere, and a diurnal oscillation of similar magnitude excited primarily by EUV heating within the thermosphere. The prediction of a strong semidiurnal oscillation causes our results to differ in form and magnitude from present observationally based models. The nature of the inadequacies in our calculations and in data analyses is discussed in detail.

1. Introduction

We have in Parts I and II (Lindzen, 1970; Lindzen and Blake, 1970) investigated daily variations in the thermosphere only insofar as they are forced by excitations below 80 km. Parts I and II also give, in detail, the atmospheric models and computational formalism used. Our results showed conclusively that diurnal (24-hour harmonic components) oscillations excited below 80 km will not significantly affect the atmosphere above 150 km. Now, thermospheric densities estimated from satellite drag data (King-Hele and Walker, 1961; Harris and Priester, 1965; Jacchia and Slowey, 1967) do exhibit daily variations compatible with large daily temperature variations. These variations appear to be primarily diurnal (as opposed to semidiurnal); they must be almost entirely due to thermal excitation within the thermosphere: to the absorption of ultraviolet radiation (1200–1800 Å) by O₂ and the absorption of extreme ultraviolet radiation (EUV) by O₂, O and N₂ (Craig, 1965). Such sources
will prove effective if they provide excitation at sufficiently great heights (above about 150 km) where molecular transports are very important and where, therefore, diffusion can transport disturbances into the upper thermosphere with little or no loss of amplitude (viz. Sec. 5 of Part I). We examine in this, Part III, tidal oscillations excited by such high altitude sources, using the "equivalent gravity mode" (E.G.M.) formalism described in Parts I and II. This formalism neglects the changes in meridional scale which possibly occur due to interactions of viscosity with coriolis forces. In order to ascertain the errors due to the use of E.G.M.'s we also examine, in Sec. 8, the response of diurnal excitation in an artificial ("ersatz") gravity mode whose meridional wavenumber differs from that appropriate to any E.G.M. The results suggest that errors on the order of 40% in amplitude and one hour in phase may result from using the E.G.M. formalism. Within the limitation imposed by this uncertainty we find that we are able, with reasonable choices for the distributions of excitation and mean temperature, to account for the observed diurnal oscillation in the thermosphere. Amplitudes of from 100 °K–200 °K in the temperature oscillation with maximum temperature occurring between about 16.00 LT and 14.00 LT are predicted—at least above 200 km. Below 200 km there are complicated variations in both amplitude and phase.

At this point our agreement with satellite drag measurements is satisfying if not exciting. Unfortunately, this is not the true state of affairs. The same calculations which predict a "realistic" diurnal oscillation also predict a comparably large semidiurnal oscillation within the thermosphere, but excited primarily by ozone and water vapor absorption below 80 km. While satellite drag data indicate some harmonic distortion in the daily variation (maxima and minima do not occur twelve hours apart), they do not suggest a semidiurnal oscillation of the same magnitude as the diurnal oscillation. However, Jacchia (personal communication; see also Harris and Priester, 1968) has pointed out that the methods of data processing and analysis involve much implicit smoothing in time which might be expected to obscure the semidiurnal component of the daily temperature variation as deduced from satellite drag data. How this might occur is described in Sec. 7 of this paper.

In Sec. 2 I describe, in detail, the models used for excitation above
80 km (i.e., UV and EUV radiation as absorbed by $\text{O}_2$, $\text{N}_2$ and $\text{O}$). There is no need to include excitation below 80 km in studying the diurnal oscillations of the upper thermosphere. However, for the semidiurnal oscillation of the thermosphere we must include all sources of excitation. Indeed, the most important sources of semidiurnal excitation prove to be those below 80 km. These are described in Part II.

In Sec. 3 I describe the atmospheric models to be used. In particular I examine atmospheres with different mean temperature distributions, since daily variations in the upper thermosphere turn out to be particularly sensitive to this distribution. Sections 4 and 5 give our results for diurnal and semidiurnal oscillations. For the diurnal oscillations we find, in agreement with earlier work (Harris and Priester, 1965), that EUV excitation is much more important than UV excitation. For a mean exospheric temperature of 800 °K we find that diurnal exospheric temperature oscillations with amplitudes of about 200 °K are possible. However, for a fixed spatial distribution of diurnal excitation, exospheric response decreases as the exospheric mean temperature increases. Phase also varies somewhat with exospheric mean temperature. For the semidiurnal temperature oscillation we find upper thermospheric amplitudes on the order of 200 °K—almost independent of the mean thermospheric temperature distribution. However, the phase of the semidiurnal oscillation depends markedly on the mean temperature distribution.

Section 6 deals with the calculation of density oscillations following the procedure outlined in Sec. 4 of Part I. An example of such a calculation is given in Sec. 7, where we show superposed diurnal and semidiurnal oscillations in temperature and density as calculated for an 800 °K mean exospheric temperature. These results show some points of agreement with the results of satellite drag analyses. In particular, maximum temperature is predicted to occur near 1400 hours local time. On the other hand the predicted range of density variation is much larger than that estimated from satellite drag data, and it has a large semidiurnal component. Indeed, the presence of a semidiurnal component of the same order of magnitude as the diurnal component accounts for almost all the difference between the present results and data analyses.
The present calculations predict semidiurnal and diurnal oscillations of comparable magnitude in the tropical thermosphere; satellite drag analyses reveal the presence, primarily, of the diurnal component. The crucial question, therefore, is why is the semidiurnal component selectively suppressed. One possibility, discussed in Sec. 7, is that the drag data analyses, themselves, have filtered out the semidiurnal component—and, in the process, have misrepresented mean densities and temperatures. In Sec. 8 the calculations are critically assessed, with special emphasis on the sensitivity of our results to the choice of mean temperature and to the use of the E.G.M. formalism.

2. High Altitude Thermal Excitation

We have based our excitation distributions on the calculations of Johnson (1958) as described by Craig (1965). Johnson (1958) calculated the rate of energy absorption as a function of height, assuming vertical incidence of radiation (noon over the equator at equinoxes). He also took

$$\overline{\epsilon F(\infty)} = 0.7 \text{ ergs cm}^{-2} \text{ sec}^{-1}$$

(1)

where

$$\epsilon = \text{ thermal efficiency of absorption processes,}$$

$$F(\infty) = \text{ incident solar flux,}$$

and the overbar indicates that the quantities result from integrals over the wavelength range 220–900 Å.

Separate results are available for x-ray absorption (which appears to be thermodynamically negligible), EUV absorption (220–900 Å) and UV absorption (1300–1800 Å).

The first step in specifying our tidal excitation was to express Johnson's (1958) absorption distributions in terms of analytic functions. For UV absorption we obtained

$$q_{1300-1800\text{Å}} = q_{\text{UV}} = 2.685 \times 10^{-7} \mu \text{watt/cm}^2 \times f_{\text{UV}}(z)$$

(2)

where

$$f_{\text{UV}}(z) = \exp (1 - z_1 - \exp(-z_1))$$

(3)
and

\[ z_1 = \frac{z - 104.7 \text{ km}}{9.065 \text{ km}}; \]

\( f_{UV}(z) \) has a maximum value of unity at \( z = 104.7 \text{ km} \).

For EUV absorption we get

\[ q_{220-900 \text{ Å}} = q_{EUV} = 8.3 \times 10^{-9} \text{ μwatt/cm}^2 \times f_{EUV}(z) \tag{4} \]

where

\[ f_{EUV}(z) = \sqrt{2} \exp \left\{ -z_2^2 + 0.5 - z_3 - \exp(-2z_3) \right\}, \tag{5} \]

\[ z_2 = \frac{z - 147 \text{ km}}{150 \text{ km}} \]

\[ z_3 = \frac{z - 127.4 \text{ km}}{56.5 \text{ km}}. \]

\( f_{EUV}(z) \) has a maximum of unity at \( z = 147 \text{ km} \).

Harris and Priester (1965) found that for time-dependent calculations of thermospheric structure, it was necessary to adopt a value for \( \overline{\varepsilon F}(\infty) \) of 2.1 ergs/cm\(^2\) sec in order to simulate the observationally based estimate for the mean exospheric temperature. Harris and Priester (1965) considered only direct solar absorption and thermal conductivity—omitting both motion and other sources of heating.\(^\dagger\)

While Harris and Priester's choice for \( \overline{\varepsilon F}(\infty) \) is not out of the realm of possibility, there remains significant controversy over its proper value (Harris and Priester, 1968, Craig, 1965). As a compromise we will adopt one half of the large value; i.e., (4) will be replaced by

\[ q_{EUV} = 1.25 \times 10^{-8} \text{ μwatt/cm}^2 \times f_{EUV}(z); \tag{6} \]

\( q_{UV} \) will remain unchanged. The choice of \( q_{EUV} \) is further complicated by the fact that it varies profoundly with solar activity. We are assuming that (6) is characteristic of moderate activity.

Having specified \( q_{UV} \) and \( q_{EUV} \) we are left with the problem of estimating the latitude and time dependence of our excitation. Given

\(^\dagger\) In a separate note sent to J. Geophys. Res., Lindzen and Blake show that, on the basis of theoretical calculations, the dissipation of the main semidiurnal tide in the thermosphere leads to sufficient deposition of energy (comparable to an \( \overline{\varepsilon F}(\infty) \) averaged over the whole day of 0.4 ergs/cm\(^2\)/sec or a nighttime value of about 1.0 ergs/cm\(^2\)/sec) to substantially relax the need for mean solar heating.
uncertainty over the precise excitation we have felt free to use a rather crude procedure. For the amplitude of the diurnal excitation at the equator we have taken $1/2(q_{UV} + q_{EUV})$. The phase is taken to be such that there is maximum heating at 1200 hours local time. We have chosen the amplitude of the semidiurnal excitation to be 37% of the diurnal amplitude by analogy with the relation between diurnal and semidiurnal ozone excitation (viz. Sec. 3 of Part II; when the excitation is expressed in terms of $\tau$, the temperature oscillation produced in the absence of motion, molecular transport, etc., the semidiurnal excitation is 18.5% of the diurnal excitation. However, $\tau$ is proportional to $J/\sigma$ and $\sigma$, the frequency, is twice as large for a semidiurnal than a diurnal oscillation). Again by analogy, the phase of the semidiurnal oscillation is such that we have maximum heating at 00.00 and 12.00 hours local time.

Finally, we assume that the latitude variation of the high altitude excitation is the same as that for ozone absorption (viz. Sec. 3 of Part II). Thus, we may write the excitation for each of the three main tidal modes as follows:

$$
\begin{align*}
\begin{bmatrix}
J_{\text{high altitude diurnal prop.}} \\
J_{\text{high altitude diurnal trap.}} \\
J_{\text{high altitude semidiurnal}}
\end{bmatrix} &= \begin{bmatrix}
0.548 \\
0.451 \\
0.37
\end{bmatrix} \times 0.5 \times (q_{UV} + q_{EUV}) \\
\exp[i(\sigma t + x/\alpha + \pi/2)] \\
\exp[i(\sigma t + x/\alpha + \pi/2)] \\
\exp[i(2(\sigma t + x/\alpha) + \pi/2)]
\end{align*}
$$

(7)

where phase is related to the hour of maximum by Eqs. (10) and (15) of Part II.

3. Atmospheric Models

The details of our atmospheric models are, for the most part, given in Part I. Thus our distributions of both eddy and molecular viscosity are given in Sec. 2c of Part I. Our distribution of ion drag is described in Sec. 2e of Part I; $z_c$ is taken to be 350 km. This model for ion drag would be appropriate at the equator if the geographical
and geomagnetic equators coincided. Of course, this is generally unrealistic; however, as shown in Part I, in the presence of molecular viscosity the effect of ion drag is small and its precise specification unimportant. The specification of Newtonian cooling is largely irrelevant to an investigation of the thermosphere. Thermal excitation has been described in Sec. 2 of this part and in Sec. 3 of Part II. The assumed distribution of molecular weight is given by Eq. (20) of Part I. What remains is the specification of the basic temperature distribution. We have retained the distribution of Eq. (18) of Part I up to about 75 km, while varying the exospheric temperature (i.e., the temperature of the isothermal region above about 200 km) and the temperature distribution between 75 and 200 km. In order to represent these variations we have had to modify Eq. (17) of Part I by increasing the limit for summation from 4 to 5. Table 1 gives the values of the temperature parameters for the five profiles examined in this paper. The profiles, themselves, are shown

<table>
<thead>
<tr>
<th>Profile</th>
<th>(c_i)</th>
<th>(z_i)</th>
<th>(\delta_i)</th>
<th>(c_i)</th>
<th>(z_i)</th>
<th>(\delta_i)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>-6.50</td>
<td>3.265</td>
<td>-5.140</td>
<td>6.809</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>2</td>
<td>same as Profile 1</td>
<td>16.00</td>
<td>50.000</td>
<td>82.000</td>
<td>180.000</td>
<td>—</td>
</tr>
<tr>
<td>3</td>
<td>same as Profile 1</td>
<td>4.00</td>
<td>7.50</td>
<td>9.000</td>
<td>20.000</td>
<td>—</td>
</tr>
<tr>
<td>4</td>
<td>same as Profile 1</td>
<td>9.877</td>
<td>0</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>5</td>
<td>same as Profile 1</td>
<td>16.030</td>
<td>0</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>6</td>
<td>same as Profile 1</td>
<td>5.060</td>
<td>7.723</td>
<td>0</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Units</td>
<td>(c_i): deg K/km</td>
<td>(z_i): km</td>
<td>(\delta_i): km</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
in Fig. 1, as are the distributions of \( x \) (the height in scale heights) for each profile. Profile 1 is exactly the profile employed in Parts I and II. Profiles 2 and 3 are almost identical to Profile 1 up to about 110 km above which they increase more rapidly to exospheric temperatures of 1000 °K and 1400 °K respectively. Profile 4 follows the U.S. Standard Atmosphere up to 115 km, increasing above this level to an exospheric temperature of 800 °K. Profile 5 increases with an approximately constant gradient from the mesopause to a 1400 °K exosphere. Profile 5 is considered only in Sec. 8. The distributions of \( x \) with height prove to be particularly important since mean density depends primarily on \( x \) and molecular diffusivities depend primarily on density. Profile 4 is considered the most realistic one and most of our calculations are for this profile. The question of mean temperature is discussed further in Sec. 8.

![Figure 1](image-url)  
Figure 1. Vertical distributions of mean temperatures, \( T_e \), used in our calculations. Also shown are the vertical distributions of height in scale heights, \( x \), corresponding to each of the temperature distributions. \( T_e = 800 \, ^\circ K \) corresponds to Profile 1, \( T_e = 1000 \, ^\circ K \) to Profile 2, \( T_e = 1400 \, ^\circ K \) to Profile 3, \( T_e = 800 \, ^\circ K \) (STAN. ATM.) to Profile 4, and \( T_e = 1400 \, ^\circ K \) (MOD) to Profile 5.
4. Solar Diurnal Response

The cases investigated for both the diurnal propagating and diurnal trapped modes all involved the sum of $O_2$ and EUV excitation. The four cases correspond to the use of profiles 1–4 for $T_0$. It has already been shown in Parts I and II that $O_3$ and H$_2$O diurnal excitation will not significantly affect the thermosphere—at least not above 200 km.

4a. Diurnal Propagating Mode

The amplitudes and phases for the temperature oscillations associated with the main diurnal propagating mode for each of the four temperature profiles are shown in Figs. 2 and 3. The amplitudes and phases for the horizontal velocity oscillations for temperature profile 4 are shown in Figs. 4–7. There is an evident tendency shown in Fig. 2 for the thermospheric response above 200 km to decrease as the thermosphere becomes hotter; however, a comparison of the responses for profiles 2 and 3 shows that the situation is more complicated. For the profiles considered here, the $x(z)$'s are virtually identical up to about 110 km. Hence, the height above which this mode propagates upward only with exponential amplitude decay is the same for each profile (ca. 108 km; viz., Part I). However, above 110 km the distributions of $x$ for each of our $T_0$ profiles begin to diverge. Thus, the height, at which the transition from propagation with attenuation to diffusion dominated constancy occurs, increases as the thermosphere gets warmer (from about 180 km for profile 4 to 300 km for profile 3). Oscillations excited above 108 km but below the transition level will be subject to considerable attenuation before entering the upper thermosphere, while oscillations excited near or above the transition will suffer little amplitude attenuation on entering the upper thermosphere. Now, our EUV excitation is centered near 150 km. Hence, we expect the response of the upper thermosphere to decrease as the transition level moves above 150 km. However, the situation is complicated by the fact that oscillations excited below the transition undergo rapid variations in phase with height before reaching the transition.

† EUV excitation proved much more important. As a result, there seems to be little advantage in considering $O_2$ excitation separately.
Figure 2. Vertical distributions of the amplitude of the temperature oscillation for the diurnal propagating mode. Results are shown for various assumed distributions of $T_0$: Profile 1 ———, Profile 2 - - - , Profile 3 ———, and Profile 4 . . . . . See Fig. 1 for the various profiles.

Figure 3. Same as Fig. 2 for the phase of the temperature oscillation for the diurnal propagating mode.
Figure 4. Vertical distributions of the amplitude of the westerly velocity oscillation for the diurnal propagating, diurnal trapped, and semidiurnal modes. Profile 4 has been assumed for $T_0$.

Figure 5. Same as Fig. 4 for the phase of the westerly velocity oscillation.
Figure 6. Same as Fig. 4 for the amplitude of the southerly velocity oscillation.

Figure 7. Same as Fig. 4 for the phase of the southerly velocity oscillation.
Thus, for transition levels above the bulk of the excitation we expect interference between oscillations excited at different levels. This is occurring for profiles 2 and 3.

Of the profiles considered here, it is for profile 4 that our diurnal excitation produces the largest response. Concentrating on profile 4's results we see that a temperature oscillation of amplitude 74°K is associated with the main diurnal propagating mode at the equator (values for other latitudes may be deduced from the discussion of Sec. 2 of Part II). Maximum temperature is predicted to occur at 1300 hours local time. Associated with this temperature oscillation is an oscillation in zonal wind of amplitude 10 m/s at the equator and an oscillation in southerly velocity of 120 m/s appropriate to a latitude of 18°. Reference to Sec. 2 of Part II shows that larger zonal wind oscillations are expected at latitudes away from the equator, while smaller southerly velocity oscillations are expected away from 18°.

4B. DIURNAL TRAPPED MODE

Results for this mode are shown in Figs. 8 and 9 as well as in Figs. 4–7. In Fig. 8 we see that for this mode the reduction of upper thermospheric response with increasing thermospheric temperature is clear and uncomplicated. The inviscid, adiabatic diurnal trapped mode is, as its name implies, evanescent; i.e., it decays exponentially with height away from levels of excitation without variation of phase. As viscosity and thermal conductivity grow in importance there is a transition to diffusion dominated constancy with height. This transition occurs again in the neighborhood of between 180 km for profile 4 and 300 km for profile 3. As with the diurnal propagating mode, the response of the upper thermosphere to a given excitation decreases as the distance that the transition region is above the excitation increases. However, unlike the diurnal propagating mode, the trapped mode is not associated with phase variations (except in the transition region; viz. Part I) and hence, it is not subject to particularly significant interference effects. It is for this reason also, as mentioned in Sec. 7 of Part I, that the diurnal trapped mode is, in fact, more effective than the diurnal propagating mode in carrying a disturbance from the lower to the upper thermosphere. A comparison of Figs. 2 and 8 quickly shows that even with less excitation
Figure 8. Same as Fig. 2 for the amplitude of the temperature oscillation for the diurnal trapped mode.

Figure 9. Same as Fig. 2 for the phase of the temperature oscillation for the diurnal trapped mode.
at the equator (viz. Eq. 7) the upper thermospheric response for the diurnal trapped mode is greater than that for the diurnal propagating mode. For profile 4, which is again associated with the largest diurnal response, we predict a temperature oscillation at the equator of amplitude 133°K (according to the E.G.M. formalism we expect greater amplitudes for this mode at greater latitudes; viz. Sec. 2 of Part II), with maximum temperatures predicted for 15.00 hrs local time. Associated with this temperature oscillation is a westerly velocity oscillation with an amplitude of 35 m/s at the equator and a southerly velocity oscillation with an amplitude of 113 m/s at about ±30° latitude. Again, we refer the reader to Sec. 2 of Part II for estimates of the oscillation at other latitudes.

40. TOTAL DIURNAL OSCILLATION

Finally, we must point out that the total diurnal variation is made up from the sum of the contributions in each diurnal mode. In this study we have considered only two modes, assuming that other modes will be of negligible importance. At the equator the sum of these two modes leads, for profile 4, to a diurnal temperature oscillation with an amplitude of about 200°K and maximum temperature at about 14.00 hours L.T. For profiles 1 and 2, amplitudes are reduced but phases are similar.

The amplitudes are, of course, proportional to the EUV excitation which as mentioned in Sec. 2, is subject to both uncertainty and variability. All that can be said is that our choice for the overhead sun value of $\overline{F}(\infty) = 1.05$ ergs/cm²/sec is that it is intermediate among values commonly employed.

5. Solar Semidiurnal Response

Six cases are investigated for the main semidiurnal E.G.M. The cases are described in Table 2. The amplitudes and phases of the thermospheric temperature oscillations as functions of height for each of the above cases are shown in Figs. 10 and 11. The amplitudes and phases of the westerly and southerly velocity oscillation for case (d) are shown in Figs. (4–7). The semidiurnal oscillations excited by O₃ and H₂O are described in detail in Part II. Their propagation into the thermosphere is described in Part I. These
Figure 10. Vertical distributions of the amplitude of the semi-diurnal temperature oscillation for cases (a)–(f) (see Table 1 for descriptions). (a)——, (b) --- ----, (c) ———, (d) . . . . . , (e) ——— ———, (f) ——— ———.

Figure 11. Vertical distributions of the phase of the semi-diurnal temperature oscillation for cases (a) ——— , (b) ——— , (c) ——— , (d) . . . . . , (e) ——— ——— , and (f) ——— ———.
Table 2 Description of cases studied for the semidiurnal mode.

<table>
<thead>
<tr>
<th>Case</th>
<th>( T_0 ) Profile</th>
<th>( H_2O )</th>
<th>( O_3 )</th>
<th>( O_2 )</th>
<th>EUV</th>
</tr>
</thead>
<tbody>
<tr>
<td>a</td>
<td>1</td>
<td>×</td>
<td>×</td>
<td>×</td>
<td>×</td>
</tr>
<tr>
<td>b</td>
<td>2</td>
<td>×</td>
<td>×</td>
<td>×</td>
<td>×</td>
</tr>
<tr>
<td>c</td>
<td>3</td>
<td>×</td>
<td>×</td>
<td>×</td>
<td>×</td>
</tr>
<tr>
<td>d</td>
<td>4</td>
<td>×</td>
<td>×</td>
<td>×</td>
<td>×</td>
</tr>
<tr>
<td>e</td>
<td>4</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>×</td>
</tr>
<tr>
<td>f</td>
<td>4</td>
<td>—</td>
<td>—</td>
<td>×</td>
<td>×</td>
</tr>
</tbody>
</table>

sources must again be considered here because the main semidiurnal mode alone, among the main tidal modes, appears capable of penetrating the thermosphere from below. As we can see from Fig. 2, the oscillations produced by UV and EUV excitation within the thermosphere are smaller than those excited below the thermosphere. Also, the excitation produced by UV absorption is smaller than that produced by EUV absorption, though less pronouncedly than for diurnal modes. From Fig. 3 we see that the oscillations excited in situ for temperature profile 4 are out of phase with those excited below. Hence, the amplitudes predicted for case (d) are smaller than what would have been predicted in the absence of in situ excitation.

A comparison of cases (a), (b) and (c) indicates a tendency for the semidiurnal response of the thermosphere to also decrease as the thermosphere gets hotter. Though not shown in our figures, the amplitude of the thermospheric response to \( H_2O \) and \( O_3 \) excitation is quite insensitive to thermospheric temperature. (The phase of the response, however, does depend on the thermospheric temperature distribution. The reason for this may be seen in Fig. 1 where we see that the height at which a given value of \( x \), or relatedly the mean atmospheric density, occurs moves up as the atmosphere gets hotter. Now density determines molecular diffusion and hence the height at which the transition from an inviscid, adiabatic, upward propagating wave to a diffusive solution independent of height will depend on the mean temperature distribution. The phase of the inviscid wave increases with height. Thus the greater the height at which the transition occurs, the greater the phase of the diffusive oscillation.) The amplitude of the response to in situ excitation, on the other hand,
decreases substantially as the thermosphere gets hotter.† For temperature profiles 1, 2 and 3, the responses for in situ excitation is reflected in the total response (viz Fig. 2). Actually, the largest response to in situ excitation occurs for profile 4. However, for this profile the response to in situ excitation is out of phase with the response to excitation from below and the overall response for case (d) is less than for cases (a) and (b). However, since excitation from below is of dominant importance the amplitude of the thermospheric response differs little among cases (a), (b), (c) and (d). In general we predict a semidiurnal oscillation in temperature at the equator with amplitude between about 190° and 280°K. These amplitudes obtain at altitudes above about 150 km. The amplitude decreases below this height to about 10°K at 90 km. In general, we expect amplitudes to decrease away from the equator (viz. Sec. 2 of Part II). As we see in Fig. 3, the phase of the response is largely determined by the distribution of mean temperature.

Such large temperature oscillations must be accompanied by large horizontal velocity oscillations. The predicted distributions of horizontal velocity for case (d) are shown in Figs. (4–7). Amplitudes of about 300 m/s for the southerly velocity and 150 m/s for the westerly velocity are predicted. No measurements of the time dependent behavior of wind at these altitudes are available. However some isolated rocket soundings up to 200 km have been made which reveal winds up to 250 m/s with virtually no shear above about 165 km (Bedinger, 1970).

As mentioned in Part II, such large temperature and wind amplitudes clearly indicate that non-linear processes will be important. However, as the calculations of Part II (Sec. 7) showed, such processes while important are not dominant for the semidiurnal tide; they can hardly be expected to reduce amplitudes by more than about 30%, since for further reductions the non-linear terms will, indeed, be relatively small. It is a trivial consequence of the nature of non-linearity that if any other process should reduce amplitudes by a similar amount then nonlinear processes will simply not come into play. Finally, it must be noted that in the E.G.M. formalism, the amplitudes shown in Figs. 4 and 6 do not apply to all latitudes.

† The reasons for this were discussed in section 4 on diurnal oscillations; also see section 8.
From Fig. 1 of Part II we see that smaller amplitudes may be expected at most latitudes—especially for the southerly velocity component.

Just as the diurnal amplitudes are proportional to the EUV excitation, so the semidiurnal amplitudes are approximately proportional to the excitation due to insolation absorption by ozone and water vapor. However, the latter sources are not (as far as we know) subject to solar cycle variability, and they appear to be relatively well known. While the lower atmospheric sources of excitation are subject to some uncertainty, the generally good agreement between tidal observations between the ground and 100 km and the theoretical predictions for both diurnal and semidiurnal tides excited by the sources described in Part II (Chapman and Lindzen, 1970) suggests that the uncertainty is not great. As an annual average the excitation is not expected to be more than 20% in error. Probably more important as concerns the penetration of the thermosphere by the semidiurnal tide is the variability and uncertainty of the mean lapse rate of the mesosphere between 50 and 80 km. In this region the main semidiurnal mode is evanescent, and the greater the lapse rate the less the semidiurnal tide is able to penetrate into the upper atmosphere. Our present choice for the mean lapse rate in this region is larger than most estimates for the tropics, but it is conceivable that during at least part of the year the lapse rate might still be larger—even in the tropics—in which case there might be as much as a 30% reduction of semidiurnal amplitudes in the thermosphere (Lindzen, 1968).

Taken at their most conservative, however, the above considerations still imply that the semidiurnal oscillation in the thermosphere must have an amplitude well in excess of 100°K.

6. Calculation of Density Variation in the Thermosphere

In Sec. 7 we will present results for the total daily variation in the thermosphere (i.e. the variations resulting from the sum of the main semidiurnal, the main propagating diurnal and the main trapped diurnal modes). Insofar as most data refer to density (as derived from satellite drag) we will certainly want to present calculated results for this field. As was mentioned in Sec. 4 of Part I, even if
\( \delta \rho/\rho_0 \) is less than 1 below the altitude where \( u', v' \) and \( \delta T \) become independent of height (i.e., the region of diffusion dominance), it will continue to grow, linearly, above this altitude and will eventually exceed one. Clearly the linearized calculation of density becomes inadequate sufficiently high in the thermosphere. This, however, does not mean that the problem as a whole has become nonlinear. If our solution is valid below the level where diffusion dominates the horizontal momentum and heat balances, the solutions for \( u', v' \) and \( \delta T \) will remain valid in the upper thermosphere. This is simply because, when diffusion dominates, the solutions are independent of density. Therefore, all that is called for is a recalculation of density using the hydrostatic relation and the perfect gas law. Let

\[
T = T_0(z) + \delta T_{\text{tidal}},
\]

where

\[
\delta T_{\text{tidal}} = \delta T_{\text{semidiurnal}} + \delta T_{\text{diurnal prop.}} + \delta T_{\text{diurnal trapped}}
\]

Then

\[
p = p(0) e^{-x}
\]

\[
\rho = \frac{p}{RT}
\]

where

\[
x = \int_0^z \frac{g}{RT} dz.
\]

These relations are the same as those for the basic density and pressure fields—except that \( T_0 \) is replaced by \( T \). When \( \delta p/\rho_0 \) and \( \delta \rho/\rho_0 \) are small then (10) and (11) yield approximately the same results as are obtained from the fully linearized equations.

7. Total Daily Variation of the Upper Thermosphere

In this section we restrict our discussion to fields for which extensive data analyses exist: namely density and upper thermospheric temperature as derived from density. Our calculated daily variations (sum of diurnal and semidiurnal contributions) of upper thermospheric temperature for basic profiles 1–4 are shown in Fig. 12.

† The \( \delta T \)'s must be expressed as real functions of time using the phase connection introduced in section 3 of Part II.
Figure 12. Daily variation of exospheric temperature for different distributions of $T_0$.

The general nature of the daily variation of temperature and, in particular, the hour at which maximum temperature occurs depends on our assumed profile for $T_0$. The main maximum occurs at 14.00 hrs. for profile 1, 13.00 hrs. for profile 2, 22.00 hrs. for profile 3,† and 15.7 hrs. for profile 4. Note, the hour of maximum for the total variation differs from the hour of maximum for the diurnal component.

† Profile 3 for which the diurnal oscillation is substantially suppressed is almost certainly incompatible with our choice of EUV excitation; see Sec. 8.
alone. Most data analyses suggest a maximum at 14.00 hrs (Harris and Priester, 1968; Jacchia and Slowey, 1967). Present data analyses also suggest ranges of daily variation varying from about 200 °K to 600 °K. We find ranges between 415 °K for profile 3 and 750 °K for profile 4. The difference between the results derived from satellite drag and theoretical results is almost entirely due to the prediction of a strong semidiurnal component in the theoretical daily variation. Harris and Priester (1968) suggest a minimum temperature at 04.00 hrs.† Since this is not 12 hrs away from the estimated maximum, some semidiurnal component is suggested by their results, but not a very prominent contribution. "Observed" temperatures are, of course, deduced from "measured" densities. Thus the above differences should be evident in our predicted density variations. Indeed; as we see from Eqs. (10) and (11) density depends exponentially on temperature (or, more precisely, on 1/T) and the exponent is proportional to altitude; hence, at high altitudes the differences as displayed in density should be greatly magnified. Similarly the density variations associated with the various assumed mean temperature profiles will also differ markedly—one from the other. However, each will be characterized by a strong semidiurnal component. For purposes of illustration we will concentrate on the results for profile 4. These are shown in Fig. 13.

Satellite drag data suggests maximum density at 14.00 hrs, minimum density at 04.00 hrs, and a range of variation at 600 km of about an order of magnitude. For profile 4 (viz. Fig. 13) we predict at 600 km a maximum density at about 16.00 hrs and a second, secondary maximum at about 05.00 hrs with densities about an order of magnitude less than at the primary maximum. The rough agreement between the two predicted maxima and the maximum and minimum derived from satellite drag data is readily traceable to the fair agreement between our calculated diurnal oscillation and that derived from satellite drag data. However, we also predict minima in density at 10.00 hrs and at 23.00 hrs, the latter with a density 5 orders of magnitude less than the primary maximum. These minima are due to the predicted presence of a strong semidiurnal oscillation. Our density predictions must have modest errors due to the errors in

† It is interesting to note that we predict, for profile 4, a secondary maximum at 04.00 hrs. We shall return to this point later in this section.
Figure 13. Daily variation of density at selected altitudes. Profile 4 has been assumed for $T_0$.

our specification of the distribution of mean molecular weight (viz. Sec. 2a of Part I) and due to the fact that we have not taken account of variations with height of $g$, the gravitational acceleration. However, the large variation of density is primarily due to the large variation of thermospheric temperature we predict—again due to the addition of a large semidiurnal oscillation to the diurnal oscillation. For profile 4 we predict daily variations in temperature from 460 °K to 1210 °K ($T_0 = 800 °K$ in the upper thermosphere for profile 4).

If our predictions are even roughly correct, then they will have important ramifications for all of upper atmosphere physics and hence for satellite technology. They suggest, for example, the possibility of selecting orbits at any given altitude which will involve a great
reduction of drag. However, the crucial question remains as to how satellite drag data analyses could have missed a strong semidiurnal oscillation—assuming for the moment that it is present.

In computing satellite drag it is assumed that density decreases with height with a constant scale height, independent of time (or equivalently, longitude). While it is recognized that this assumption is incorrect, it is also presumed that almost all drag is exerted near satellite perigee so that it doesn’t matter what the density distribution is away from perigee. (A review of satellite drag calculations may be found in Craig, 1965.) Behind this presumption are the following assumptions: (1) it takes on the order of 12 hrs (or, equivalently, 180° of longitude) for air density at a given level to go from maximum to minimum†; (2) for typically eccentric satellite orbits, satellite altitude will be more than 50–100 km above perigee altitude once one gets 45° or more away from perigee; and (3) atmospheric density at a given level is less than the density at any longitude at a level 50–100 km lower than the first level.

It is not within the scope of this paper to present a critical analyses of satellite drag analyses. However, it is clear that if a density variation like that shown in Fig. 13 were to exist, then assumptions (1) and (3) above would be strongly violated. For example, in an equatorial plane, there are over 120° of longitude (not necessarily contiguous) for which there would exist higher densities at 600 (598) km than exist at some longitudes at 400 (399) km. Let us, for purposes of argument, assume that Fig. 13 represents the actual behavior of the atmosphere. If a satellite orbit were such that perigee occurred at 23.00 hrs at 400 km altitude, then most of the drag would occur away from perigee at greater altitudes. However, present satellite drag analyses would attribute all the drag to the neighborhood of perigee and would therefore assign a spuriously large density to the particular altitude and time of perigee. In this manner, we expect present satellite drag analyses will smooth out very pronounced density minima like those shown in Fig. 13. This may account for why there appears to be fair correspondence between

† This assumption is basically necessary because even under ideal circumstances satellite drag density measurements constitute a weighted mean over at least 45° of longitude (or about 3 hours of local time). This alone will tend to selectively suppress semidiurnal oscillations.
our predicted primary and secondary maxima and "observed" maxima and minima. In such a situation, the drag analysis would also misrepresent the mean density as being between the two maxima and would therefore suggest a $T_0$ in excess of 1000 °K rather than 800 °K.

The purpose of the above discussion was merely to show that variations of thermospheric density like those we have been predicting would be severely distorted by satellite drag analysis. I hardly wish to suggest that the calculations presented are sufficient to show that the density variations shown in Fig. 13 are "correct." The sensitivity of the results to the assumed distribution of $T_0$ alone precludes this; and the critical assessment, in Sec. 8, of the calculations will show other uncertainties. Nevertheless, after all the uncertainties have been considered, the fact seems to remain that we still expect a strong semidiurnal oscillation in the thermosphere. Its amplitude may not be as large as we have calculated, but it should still exceed 100 °K. The addition of such an oscillation to the diurnal oscillation would lead to a density variation which, in a similar manner to the above example, could not be accurately described by present satellite drag analyses. Improved methods of thermospheric probing are certainly called for. Incoherent backscatter techniques or improvements of satellite drag analysis are possibilities. However, even existing satellite drag analysis programs could be used to calculate the drag that would result from density variations with strong semidiurnal components, analyze the drag for density, and compare the analyzed density with both the original density and the results from the analysis of real drag data.

8. Critical Assessment

Most of the approximations involved in our calculations have been discussed in detail in Parts I and II as well as in Secs. 4 and 5. In particular it was shown that uncertainties in ion drag are unlikely to have significant effects. As already shown, uncertainties in $T_0$ are more important and we shall discuss both this matter, and our uncertainty concerning thermospheric excitation at the end of this section. Our assumed distribution of $M$, the molecular weight, is probably in error. However, since the main variations in $M$ occur
above the beginning of the region where diffusion is dominant we expect errors in $M$ to have negligible effects on oscillations in temperature and horizontal velocity. Errors in our distribution of $M$ (as well as our ignoring variations in $g$, the gravitational acceleration) will affect predicted density variations in the thermosphere. However, it is easy to show that the errors thus produced will be smaller than the predicted range of daily variation. Errors may also arise from our neglect of the effects of mean zonal winds. Apart from the fact that the nature of such winds in the thermosphere is unknown, we know that such winds must be about 400 m/s at the equator for their effects to be comparable with those of inertia (Chapman and Lindzen, 1970). Within the region of diffusion dominance even such winds will be relatively unimportant.

Our most important limitations seem to be the assumption of linearity, and the use of the E.G.M. formalism. Both matters have been discussed in Part II. The E.G.M. formalism does not reflect the fact that the latitude and altitude dependence of tidal modes are non-separable when viscous forces and coriolis forces are of comparable importance. In this region (corresponding to the region over which the transition to a diffusion dominated solution occurs) the latitude structure of a given mode will change with height.

For the semidiurnal mode the coriolis forces are never dominant and, hence, the E.G.M. formalism should be approximately adequate. As shown in Sec. 7 of Part II (also see Sec. 5 of this paper), non-linearity will be important but not dominant for this mode (this result remains unchanged when thermospheric excitations are included). As already pointed out, if non-linear effects should reduce amplitudes by more than 30%, we will be left with an essentially linearizable problem.

For the diurnal modes, non-linearity does not seem to be a problem within the thermosphere. However, since coriolis forces are important for the latitude structure of these modes, the use of the E.G.M. formalism within the thermosphere is more problematic. Unfortunately, we are not able to present, to calculate the thermospheric response without the E.G.M. formalism. However, we can estimate the errors involved in using this formalism by investigating the effects of assuming different latitude (i.e., $y$) structures on the response of the thermosphere to in situ excitation. What we shall do is to
assume different values for the north–south wavenumber, $m$ (which may be imaginary; viz. Sec. 2i of Part I), and assume, for each choice, excitations of the same vertical structure as our O$_2$ and EUV sources and identical amplitudes and phases at the equator. This will permit us to determine, in a gross way, the effect of north–south structure on thermospheric response. Part of this task, has already been done, since the diurnal trapped and propagating E.G.M.'s have different $m$'s. To complete the task we will also investigate an artificial mode for which $m = 0$. For such a mode, the inviscid, adiabatic solutions involve growth of amplitude at a rate somewhat less than $e^{x/2}$ and no change of phase with height. As a result, such a mode is almost optimally effective in carrying a disturbance into the upper thermosphere. Of course, below the transition to diffusion dominance, the use of the E.G.M. formalism should be adequate and the use of the artificial ("ersatz") mode totally inappropriate. However, above the transition region we do not know what the meridional structure of our oscillation will be (though we expect it to be smoother than the sum of the E.G.M.'s) and a comparison of the thermospheric response for each of these modes should give us a feel for the errors involved in using the E.G.M formalism. For each mode we have assumed the same excitation as was used earlier in this paper for the diurnal propagating mode. The amplitudes and phases of the resulting temperature oscillations over the equator are shown in Table 3. At 600 km the amplitude reponse for the ersatz mode is 50% greater than that for the trapped E.G.M. and 300% greater than that for the propagating E.G.M. However, the phases for all modes are such that maximum temperature occurs between about 13.00 and 15.00 hrs local time. The differences between results for the various modes are probably greater than the errors involved in the E.G.M. formalism for the following reasons:

(i) the differences, for the most part, are already evident at 170 km at which height the E.G.M. formalism is probably adequate. Above 200 km all the modes behave quite similarly;‡

(ii) within the E.G.M. formalism the excitation is apportioned between the propagating and trapped modes; hence the differences

‡ Note that above 200 km the amplitude in each mode increases by about 40° while the phase changes by no more than 0.5 hours.
TABLE 3 Amplitudes and phases for the temperature oscillations at various heights associated with the diurnal propagating, the diurnal trapped and an artificial diurnal mode. Identical excitations (at the equator) have been used for each mode. Profile 4 has been assumed for $T_o$.

<table>
<thead>
<tr>
<th>$Z$ (km)</th>
<th>Diurnal Prop.</th>
<th>Ph ($\delta T$) (deg K)</th>
<th>Ersatz Diurnal</th>
<th>Diurnal Trapped</th>
<th>Ph ($\delta T$) (deg K)</th>
<th>Ph ($\delta T$) (hr. of max.)</th>
</tr>
</thead>
<tbody>
<tr>
<td>600.0</td>
<td>72.4</td>
<td>12.93</td>
<td>229.5</td>
<td>13.93</td>
<td>158.1</td>
<td>14.8</td>
</tr>
<tr>
<td>400.0</td>
<td>72.5</td>
<td>12.93</td>
<td>229.6</td>
<td>13.93</td>
<td>158.2</td>
<td>14.8</td>
</tr>
<tr>
<td>300.0</td>
<td>71.4</td>
<td>13.00</td>
<td>222.7</td>
<td>13.93</td>
<td>156.9</td>
<td>14.8</td>
</tr>
<tr>
<td>250.0</td>
<td>64.8</td>
<td>13.07</td>
<td>219.0</td>
<td>13.93</td>
<td>148.9</td>
<td>14.93</td>
</tr>
<tr>
<td>200.0</td>
<td>38.8</td>
<td>13.47</td>
<td>181.0</td>
<td>13.93</td>
<td>115.6</td>
<td>15.53</td>
</tr>
<tr>
<td>170.0</td>
<td>11.3</td>
<td>15.33</td>
<td>130.1</td>
<td>14.73</td>
<td>71.4</td>
<td>16.87</td>
</tr>
<tr>
<td>150.0</td>
<td>5.5</td>
<td>22.00</td>
<td>81.7</td>
<td>15.27</td>
<td>36.9</td>
<td>19.33</td>
</tr>
</tbody>
</table>

between the ersatz mode and E.G.M. results will not be as great as the differences between the ersatz mode and the propagating mode alone; and

(iii) the ersatz mode seems likely to represent an upper bound to the responsiveness of the thermosphere.

In view of the above, we tentatively conclude that the use of the E.G.M. formalism for diurnal oscillations in the thermosphere involves errors of no more than 40% in amplitude and one hour in phase at the equator. Within the upper thermosphere, however, we have no reason to believe that the detailed latitude structure associated with the E.G.M. formalism applies.

Finally, some words are in order on uncertainties in the main external inputs to our calculation; namely the distribution of EUV excitation and the distribution of $T_o$. We see from Figs. 2, 3, 8, 9, 10, and 11 that for a distribution of excitation fixed in space the thermospheric response to excitation is strongly dependent on the distribution of $T_o$. To a certain extent we have suppressed this effect by having most of the increase in temperature between the mesopause and the exosphere occur above 115 km. When we allow increases in exospheric temperature to be manifested in a uniformly increased temperature gradient between the mesopause and 200 km the effects of increased exospheric temperature are greatly enhanced. This is seen in Table 4 where results for profiles 1 and 5 are compared. In
<table>
<thead>
<tr>
<th>Altitude (km)</th>
<th>Diurnal Trapped Profile 1</th>
<th>Diurnal Propagating Profile 1</th>
<th>Semidiurnal Profile 5</th>
<th>Diurnal Propagating Profile 5</th>
</tr>
</thead>
<tbody>
<tr>
<td>150</td>
<td>11.3</td>
<td>22.33</td>
<td>2.17</td>
<td>4.03</td>
</tr>
<tr>
<td>200</td>
<td>56.6</td>
<td>16.33</td>
<td>13.61</td>
<td>22.0</td>
</tr>
<tr>
<td>300</td>
<td>86.1</td>
<td>15.53</td>
<td>35.24</td>
<td>27.7</td>
</tr>
<tr>
<td>400</td>
<td>87.2</td>
<td>15.47</td>
<td>35.17</td>
<td>27.7</td>
</tr>
<tr>
<td>600</td>
<td>87.1</td>
<td>15.47</td>
<td>35.17</td>
<td>27.7</td>
</tr>
</tbody>
</table>
general we have no reliable, accurate temperature measurements between the mesopause and 200 km. Above 200 km our estimates are based on satellite drag data. In view of the discussion in Sec. 7, we regard such estimates as uncertain. Nevertheless, there are physical reasons for assuming profile 4 to be relatively plausible—for an 800° exosphere: high exospheric temperatures are believed due to EUV absorption and other heating centered near 150 km, and convection seems likely to prevent very large positive gradients below the turbopause. Increases in exospheric temperature above 800°K are likely to be associated with increased gradients of temperature primarily above the turbopause. Even such changes will produce changes in the phase of semidiurnal oscillations in the thermosphere, as well as changes in the amplitude and phase of the diurnal response to in situ excitation. However, in the real atmosphere we may expect increases in exospheric temperature to be accompanied by less drastic alternations in the diurnal oscillation for the following reasons: increases in exospheric temperature are usually due to increases in EUV flux and hence in diurnal excitation; changes in thermospheric temperature will alter the distribution of EUV heating which is related (though not simply) to the distribution of density in the thermosphere. We felt that any attempt at more sophisticated modelling of excitation was unwarranted in view of our ignorance of absorption cross-sections and even incident solar flux. Our calculations for the diurnal oscillation simply show that for a plausible choice of $T_0$, currently acceptable distributions of excitation can lead to large thermospheric oscillations (ca. 200°K in amplitude for temperature oscillations). At our present level of ignorance, it is also useful to recognize that the response of thermosphere to in situ excitation depends not only on the magnitude and distribution of excitation, but also on the mean temperature distribution. One might suppose that one might use the mean value of $q_{EUV}$ over the day to compute the $T_0$ compatible with the oscillatory excitation—as was done in the non-hydrodynamic calculations of Harris and Priester (1965). However, as pointed out in Sec. 2, $T_0$ depends on other quantities besides $q_{UV}$, $q_{EUV}$ and molecular conductivity. Such factors as the rate of energy dissipation of the semi-diurnal tide, non LTE infrared cooling by atomic oxygen, and the little understood "eddy" diffusion in the lower thermosphere above
the mesopause are of importance. "Eddy" diffusion and atomic oxygen cooling can, moreover, be of importance for $T_0$ without being comparably important for the tidal variations. Thus we are not yet in a position to improve the accuracy of our calculations by simultaneously calculating $T_0$ and the daily variation.

Finally, I should emphasize that our uncertainties concerning the E.G.M. formalism, excitation by EUV, and the distribution of $T_0$ are far more important for the amplitude of the diurnal oscillation than they are for the amplitude of the semidiurnal oscillation. Moreover, the neglect of non-linear processes which is likely to be of importance for the semidiurnal oscillation, can only be of distinctly limited importance. Yet we are able to obtain reasonable agreement between calculations for the diurnal oscillation and satellite drag analyses; it is our semidiurnal oscillation which causes most of our disagreement with present data analyses.

9. Concluding Remarks

As we see from Sec. 8, the application of our calculations to the earth's thermosphere is subject to considerable uncertainties. None of these uncertainties, however, is large enough to fundamentally alter this paper's main conclusion: that the daily variation of the thermosphere consists not only in a diurnal oscillation excited by EUV radiation, but also in a comparably strong semidiurnal oscillation excited primarily by insolation as absorbed by ozone and water vapor in the lower atmosphere. Hence, the differences between our calculations and the currently accepted "observational" picture as discussed in Sec. 7 seem likely to survive corrections to our calculations.

A knowledge of the structure of the thermosphere is of considerable importance for a variety of problems including the calculation of satellite lifetimes. This structure is dependent on solar activity which determines the flux of EUV radiation (Craig, 1956). If one adopts the naive view that the thermospheric temperature is determined solely by the diffusive response to EUV heating then the dependence of both mean thermospheric temperature and its daily variation should be quite simple. The attempt to find such a simple relation has not been entirely successful (Jacchia and Slowey, 1967).
The present calculations imply a much more complicated dependence on solar activity primarily because a major part of the thermospheric daily variation should consist in a semidiurnal oscillation excited primarily below the mesopause. Solar activity influences this oscillation primarily through its effect on $T_0$, the mean temperature distribution. However, the amplitude of this oscillation is relatively independent of the distribution of $T_0$ within the thermosphere; only its phase is sensitive to $T_0$ in the thermosphere.

As we have discussed in Sec. 7, it will be difficult with present analyses of satellite drag data to check these possibilities.

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