

- This document has been reproduced from the best copy furnished by the organizational source. It is being released in the interest of making available as much information as possible.
- This document may contain data, which exceeds the sheet parameters. It was furnished in this condition by the organizational source and is the best copy available.
- This document may contain tone-on-tone or color graphs, charts and/or pictures, which have been reproduced in black and white.
- This document is paginated as submitted by the original source.
- Portions of this document are not fully legible due to the historical nature of some of the material. However, it is the best reproduction available from the original submission.

NASA TM X- 65694

A DIFFUSION MODEL FOR THE PHASE DELAY BETWEEN THERMOSPHERIC DENSITY AND TEMPERATURE

H. G. MAYR
H. VOLLAND



AUGUST 1971



GODDARD SPACE FLIGHT CENTER
GREENBELT, MARYLAND

FACILITY FORM 602	<u>71-36719</u>	(ACCESSION NUMBER)	(THRU)
	<u>28</u>	(PAGES)	<u>53</u>
	<u>TMX-65694</u>	(NASA CR OR TMX OP. AD NUMBER)	<u>13</u>
			(CATEGORY)

**A DIFFUSION MODEL FOR THE PHASE DELAY BETWEEN
THERMOSPHERIC DENSITY AND TEMPERATURE**

by

H. G. Mayr

**Thermosphere and Exosphere Branch,
Goddard Space Flight Center,
Greenbelt, Maryland, USA**

and

H. Volland

**Astronomical Institutes of the
University of Bonn, Bonn, Germany**

A DIFFUSION MODEL FOR THE PHASE DELAY BETWEEN
THERMOSPHERIC DENSITY AND TEMPERATURE

by

H. G. Mayr

Thermosphere and Exosphere Branch

and

H. Volland

Astronomical Institutes of the

University of Bonn

ABSTRACT

Considering a two dimensional time dependent model in which the thermosphere dynamics are excited by the UV heat input within the thermosphere, it is shown that the wind circulation-drag-diffusion induced variations in the diurnal component of atomic oxygen dominate over its temperature induced variations up to 200 km. The assumption of diffusive equilibrium is therefore in general not valid for O within the lower thermosphere. The effect of the diurnal wind circulation tends to remove O from the day to night side of the earth with the consequence that (a) the diurnal variations in the $[O]/[N_2]$ and $[O]/[O_2]$ ratios are damped by about 20% thus contributing to the maintenance of the nighttime F_2 region and (b) the maximum in the diurnal variation of O is shifted by one to two hours away from the temperature maximum toward noon thus explaining the temperature-density time lag at thermospheric heights above 200 km where O becomes the major constituent. The mass redistribution due to circulation, which accounts for significant density-temperature time delays within the N_2 region of the lower

thermosphere, is essentially extended into the upper thermosphere through the diffusion (drag) process which affects most significantly the atmospheric constituents with lower masses.

INTRODUCTION

Radar backscatter measurements have shown that the ion temperature—which has been assumed to be identical with the gas temperature—peaks at about 16³⁰ LT (e.g. Mahajan (1969)). Considering that the thermospheric density as deduced from satellite drag observations peaks at 14⁰⁰ LT (Jacchia (1964)) a phase difference of more than two hours became thus apparent.

Stubbe and Chandra (1971) showed that the dissipation of thermospheric wind energy into the ion component could significantly increase the ion temperature over the gas temperature, a mechanism which is particularly effective during night time when other external heat sources are absent and the energy loss rates reach a minimum. For this reason the magnitude and the local time dependence of the gas temperature are not known with certainty and the ion temperature measurements can only be considered as an indication for a density-temperature phase discrepancy. However any such phase difference would be significant since the two dimensional models of Dickinson et al. (1968) and Volland and Mayr (1970) find density and temperature essentially in phase with maxima near 14⁰⁰ LT.

Volland (1970), in analyzing the ion temperature measurements, concluded that (a) the diurnal component peaks at about 15²⁰ LT which is only about half an hour after the corresponding density component in the Jacchia model and (b) the much weaker semidiurnal component in the temperature peaks also at 15²⁰ LT and follows the corresponding density peak with a phase difference of more than two hours.

Chandra and Stubbe (1970) treated the thermosphere in a one dimensional model and found in agreement with the classical one dimensional model of Harris

and Priester (1962), that temperature and density peak at 17⁰⁰ LT. By introducing variable boundary conditions at 120 km Chandra and Stubbe then shifted the density peak in their model toward 14⁰⁰ LT and thus reproduced some of the observations below 300 km.

It has been pointed out (e.g. Dickinson et al. (1968), Volland and Mayr (1969)) that it is most important to treat the thermosphere in two or even three dimensional models since horizontal winds are shown to be significantly involved in the redistribution of mass and energy thus affecting amplitudes and phases in the diurnal variations considerably. We must therefore base our discussion on the results of the two dimensional models which show temperature and density to peak only two to three hours after the maximum in the diurnal heat input, the implication being that we are actually faced with two problems: the possibilities of the temperature maximum occurring in the late afternoon and of a significant phase difference between density and temperature.

THE ENERGY MODEL

It was shown by Volland and Mayr (1970) that in the upper thermosphere where the phase discrepancy is observed, the diurnal variations are primarily excited by the UV heat input within the thermosphere. This is further emphasized by preliminary results from a three dimensional model of Volland and Mayr (1971a) in which the energy coupling from the lower atmosphere is shown to be damped even more rapidly within thermospheric heights. It has been suggested by Volland and Mayr (1971b) that the semi-diurnal tide, which dominates in the lower thermosphere (Lindzen (1967)) is rapidly attenuated above 160 km and therefore the diurnal variations should be expected to prevail within the upper thermosphere in agreement with satellite observations (Spencer (1971)).

For these reasons we feel justified to adopt as basis a relatively simple two dimensional model in which only the diurnal component, excited by the solar EUV heat input within the thermosphere, is considered.

The solar heat input has been assumed to peak at the minimum of the solar zenith angle at 12⁰⁰ LT (e. g. Volland and Mayr (1970)). However, this assumption is strictly speaking not valid since the heating rates depend on the temporal variations in the temperatures and densities of the neutral and charged components of the thermosphere, none of which is symmetrical with respect to noon time. The most significant heat inputs result from dissociation of the molecular species O₂, N₂ and NO and from dissociative recombination involving the densities of molecular ions and electrons.

At present we cannot provide a quantitative description of the diurnal variations in the heating rates. However, it seems reasonable to assume that the maximum in the solar heat input occurs between 12⁰⁰ LT and 14³⁰ LT considering that the atmospheric and plasma densities peak at 14³⁰ LT and 16⁰⁰ LT respectively.

With the EUV heat input peaking at 13⁰⁰ LT the temperature maximum in the diurnal component would then occur 3.5 hours later (Volland and Mayr 1970) that is at 16³⁰ LT which would be in agreement with the radar backscatter observations.

The problem then remains to explain the relatively large phase difference between temperature and density. It will be shown in the following that wind induced diffusion processes tend to deplete atomic oxygen during daytime, thus shifting its phase sufficiently backwards to explain the phase in the diurnal density variations above 200km (where most of the satellite drag data are obtained).

The bulk motion of N_2 up to 250 km where this species dominates constitutes the wind field that affects the O concentration. As a major constituent, N_2 is not significantly influenced by diffusion. Furthermore, the energy sources that essentially determine these dynamics are located within the N_2 region up to 250 km. For these reasons, the wind field can be computed in a first approximation without considering diffusion processes.

Volland and Mayr (1970) have treated the diurnal dynamics of the thermosphere in a two dimensional model. We shall assume here that their results are valid for the temperature and for the density-and the wind-fields of N_2 , except in the phase which we shall modify in accordance with our assumption that the peak in the energy input does not occur at 12⁰⁰ LT but at 13⁰⁰ LT.

THE DIFFUSION MODEL

For atomic oxygen the continuity equations of mass and momentum conservation are (at the equator)

$$2q [O_2] - [O] [N_2] (2\alpha_1 [O] + \alpha_2 [O_2]) - \frac{\partial}{\partial r} ([O] V_{0r}) - \frac{1}{r} \frac{\partial}{\partial \lambda} ([O] V_{0\lambda}) = \frac{\partial}{\partial t} [O] \quad (1)$$

$$[O] (V_{0r} - V_r) = -D \left(\frac{\partial [O]}{\partial r} + \frac{[O]}{T} \frac{\partial T}{\partial r} + \frac{mg}{kT} [O] \right) - K \left(\frac{\partial [O]}{\partial r} + \frac{[O]}{T} \frac{\partial T}{\partial r} + \frac{\bar{m}g}{kT} [O] \right) \quad (2)$$

$$(V_{0\lambda} - V_\lambda) = 0 \quad (3)$$

where

k = Boltzmann constant

q = photo production rate due to dissociation of O_2

α_1 = rate of three body recombination ($O + O + N_2 \rightarrow O_2 + N_2$)

α_2 = rate of three body recombination ($O + O_2 + N_2 \rightarrow O_3 + N_2$)

$[O]$, $[N_2]$, $[O_2]$ = number densities for O, O_2 and N_2

$V_{0r}, V_{0\lambda}$ = transport velocities for O in the radial (r) and longitudinal direction

V_r, V_λ = wind components for the two directions

K = Eddy diffusion coefficient ($6 \times 10^{-6} \text{ cm}^2 \text{ sec}^{-1}$)

D = Molecular diffusion coefficient for O through N_2 and O_2
($0.26 (T/T_s)^{1.75} (p/p_s)^{-1}$, an experimental result arrived at by Walker (1961)). T_s and p_s are standard temperature and pressure, p is the sum of the partial pressure of N_2 and O_2

t = universal time

m, \bar{m} = mass of atomic oxygen, mean molecular mass

g = gravitational acceleration

In the longitudinal direction the drag term $[O] (V_{0\lambda} - V_\lambda)$ dominates the lateral momentum transfer thus reducing to the simplified equation (3). The inertia terms $\partial/\partial t (V_{0r}, V_{0\lambda})$ have been neglected in Equations (2) and (3) since they are negligible when compared with the accelerations due to the pressure gradients.

Although we have formally retained the photo production and loss terms for O as well as the Eddy diffusion term, these processes are not of great significance for the diurnal variations in atomic oxygen since variations with this period do not penetrate noticeably down to altitudes below 120 km.

In order to solve the Equations (1) through (3) a number of simplifying assumptions are made:

- 1.) The variables are separated into time t , longitude λ and altitude r components in the form

$$[O] = [O]_0(r) + [O]_1(r) e^{i\lambda} e^{i\omega(t-\phi_0)} \quad (4)$$

for atomic oxygen and for the diffusion flux

$$F = [O] (V_{or} - V_r) = F_0(r) + F_1(r) e^{i\lambda} e^{i\omega(t-\phi_r)} \quad (5)$$

where ω is the angular frequency corresponding to the period of one day, and ϕ is the phase

- 2.) From the two dimensional model of Volland and Mayr (1970) the wind and temperature fields were adopted. This model overestimates the energy coupling with the lower atmosphere and consequently the amplitude of the UV component is underestimated. However, this is of no significance since we shall not discuss the absolute values of the amplitudes. Corresponding to our assumption that the maximum in the diurnal heating rate occurs at 13⁰⁰ LT instead of 12⁰⁰ LT all the parameters in their model are phase shifted by +1 hour.

Thus we assume

$$V_r = V_r(r) e^{j\lambda} e^{j\omega(t-\phi_r - 1 \text{ hours})} \quad (6)$$

$$V_\lambda = V_\lambda(r) e^{j\lambda} e^{j\omega(t-\phi_\lambda - 1 \text{ hours})} \quad (7)$$

and

$$T = T_0(r) + T_1(r) e^{j\lambda} e^{j\omega(t-\phi_T - 1 \text{ hours})} \quad (8)$$

- 3.) Assuming that higher order terms in the frequency and longitude expansions can be neglected, perturbation theory is applied which leads in a straight forward manner to a set of first order differential equations

$$\frac{\partial}{\partial r} F_0 + \alpha_1 [O]_0^2 [N_2] + \alpha_2 [O]_0 [O_2] [N_2] - 2q [O_2] = 0 \quad (9)$$

$$F_0 + (D + K) \left(\frac{\partial [O]_0}{\partial r} + \frac{[O]_0}{T_0} \frac{\partial T_0}{\partial r} + \frac{[O]_0}{kT_0} \frac{g(mD + \bar{m}K)}{D + K} \right) = 0 \quad (10)$$

$$\frac{\partial f}{\partial r} + \frac{1}{[O]_0} \frac{\partial [O]_0}{\partial r} f + (2\alpha_1 [O]_0 + \alpha_2) \rho - j\omega\rho = -\frac{1}{[O]_0} \frac{\partial}{\partial r} ([O]_0 V_r) - \frac{j}{r} [O]_0 V_\lambda \quad (11)$$

$$\begin{aligned} \frac{\partial \rho}{\partial r} + \frac{1}{(D + K)} f + \left(\frac{1}{[O]_0} \frac{\partial [O]_0}{\partial r} + \frac{1}{T_0} \frac{\partial T_0}{\partial r} \frac{g}{kT_0} \frac{(mD + \bar{m}K)}{(D + K)} \right) \rho \\ = - \left(\frac{1}{T_0} \frac{\partial T}{\partial r} - \frac{1}{T_0^2} \frac{\partial T_0}{\partial r} T - \frac{g}{kT_0^2} \left(\frac{mD + \bar{m}K}{D + K} \right) T \right) \end{aligned} \quad (12)$$

with

$$f = \frac{F_1}{F_0} e^{j\lambda} e^{j\omega(t - \phi_F)} \quad (13)$$

and

$$\rho = \frac{[O]_1}{[O]_0} e^{j\lambda} e^{j\omega(t - \phi_0)} \quad (14)$$

The equations (9) and (10) describe the steady state distribution of atomic oxygen unaffected by wind and temperature variations. As they do not depend on other variables they can be solved independently. Such a solution has been described e. g. in Mayr and Volland (1970) and it is here adopted as input, with a Jacchia (1965) model for an exospheric temperature of 1000° K.

The equations (11) and (12), which are complex and therefore represent four equations, are then solved by means of a trial and error scheme with the following boundary conditions:

a) At heights above 250 km O becomes the major constituent and therefore the transport velocity for O becomes the wind velocity; we assume therefore

$$f = 0 \text{ at } 250 \text{ km}$$

b) The diffusion time increases to four days at 90 km thus the temporal variations must be small at this level; we assume therefore

$$\rho = 0 \text{ at } 90 \text{ km}$$

This assumption is somewhat arbitrary and therefore the solution is meaningless at the lower boundary level. However, at heights above 100 km the solution is very insensitive to this artificial constraint and can therefore be considered as unique.

The equations (11) and (12) have the important characteristic that the complex variables f and ρ depend linearly on the temperature T and wind fields V_r and V_λ . This implies that the effects induced by temperature and wind variations can be separated thus providing a means of identifying their relative significance.

DISCUSSION

In Figure 1 the input parameters for the temperature and wind fields are shown. They were taken from the model component of Volland and Mayr (1970) which considers the internal EUV heat input as the generation mechanism. However, owing to our assumption that the heat input peaks at 13⁰⁰ LT instead at 12⁰⁰ LT a phase shift of an hour was introduced into all the input parameters.

By solving the diffusion equations (11) and (12) the relative variations in the concentration of atomic oxygen are derived separately for the temperature and wind fields and for the sum of both. Since atomic oxygen becomes a major constituent above 250 km we assumed that it would be in diffusive (hydrostatic) equilibrium above that altitude and accordingly the solution was extended up to

400 km. The results are shown in phase and amplitude as functions of height in Figure 2. The subscripts "wind" and "T" identify the wind and temperature induced variations and the superscript, 2D, indicates the two dimensional model. The heavy lines represent the combined total effect from both these contributions.

It is apparent from Figure 2a that up to 200 km the wind induced variations in O exceed the temperature effect. Only at altitudes above 300 km the thermal expansion in atomic oxygen starts to dominate in the diurnal variations of this constituent.

Even more significant, however, is the difference in the phase of both components. Figure 2b reveals that the density maximum due to the diurnal temperature variation occurs at about 17⁰⁰ LT above 200 km, close to the local time for the temperature maximum (the temperature phase is shown in dashed line). In contrast, the density maximum for the wind component occurs nearly 6 hours earlier at 11³⁰ LT above 200 km. This phase difference results primarily from the interaction of the horizontal wind field which tends to deplete atomic oxygen on the day side and to supply it toward the night side of the earth. In this redistribution, vertical diffusion, induced by vertical winds, tends to supply atomic oxygen from below during day and to remove it toward lower altitudes during night time, a process that partially damps the effect from the horizontal wind component.

The sum of both the temperature and wind contributions, which produce the total density variation in atomic oxygen, shows then a phase shift between

temperature and atomic oxygen concentration which is in the direction of the observed phase discrepancy.

In Figure 3 the time difference between the maxima of temperature and oxygen concentration are shown explicitly. The two dimensional model is again indicated with a superscript 2D. For comparison we show as thin line the phase delay between temperature and density in the model of Volland and Mayr (1970). Dealing here with the total atmospheric density, we compare only the phases above 250 km where O becomes the major constituent. From this comparison it is evident that the diffusion process introduces in the density of the upper thermosphere a significant phase shift toward noon.

Below 250 km the phase in the total density of the diffusive equilibrium model corresponds to that of the concentration of N_2 , the dominant constituent below that height. It is evident from Figure 3 that in N_2 the phase difference increases significantly down to altitudes of about 130 km where the density peaks in the morning hour. The reason for this behaviour is similar to that for the phase shift in O. It is the mass transport due to horizontal winds from the day to the night side of the earth which constitutes an effective sink thus damping the amplitude and shifting the phase toward earlier hours. This similarity is further emphasized in the height distribution of the phases in the diurnal variations of N_2 and O. Both show a peak at some height but N_2 at 125 km and O at 170 km. The phases decrease higher up toward the phase of the temperature. However, while the phase difference in N_2 becomes negligible above 250 km the phase difference in atomic oxygen remains significant up to 400 km.

The picture that emerges from this is therefore the following:

At lower altitudes below about 200 km the atmospheric density, with N_2 being the major constituent, is in hydrostatic equilibrium. In this region mass transport due to global circulation redistributes the density by removing it from the day side and supplying it to the night side of the earth. This process constitutes a sink at times when the thermal expansion would tend to enhance the concentration and it constitutes a source at times when a thermal contraction occurs. As a consequence the phase of the N_2 concentration is shifted away from the temperature toward earlier hours in particular in the lower thermosphere where the thermal expansion and contraction are insignificant, and it approaches the phase of the temperature in the upper thermosphere above about 200 km where the temperature effects dominate. Thus a temperature-density phase discrepancy should be understandable in the lower thermosphere simply from the dynamic properties apparent from multi- (at least two-) dimensional thermosphere models that describe the diurnal variations in N_2 , excited by the UV heat source within the thermosphere.

The diurnal variations in atomic oxygen are similarly affected by the global wind circulation and by the processes of thermal expansion and contraction. The wind field redistributes the oxygen concentration removing O from the day to the night side of the earth while thermal expansion or contraction increase or decrease the density during day or night respectively. The main differences relative to the dynamics in N_2 , however, are that it is the drag interaction with the major constituent N_2 that drives the oxygen transport thus, owing to the mass difference between both species, diffusion processes are set up. Atomic oxygen is therefore not in diffusive equilibrium, particularly at lower altitudes. Furthermore, owing to the lower mass of atomic oxygen it is much less sensitive to temperature

effects than N_2 . The balance between wind and temperature induced variations in O is therefore more significantly weighted toward the wind effect and consequently the density-temperature phase difference extends up to much higher altitudes in O than it does in N_2 . The phase difference thereby extends up into the region where atomic oxygen is the major constituent and where satellite drag density data seem to indicate this kind of a phase discrepancy when compared with temperature data.

We have so far assumed that the thermosphere can be described in a two dimensional model. Thus, we have completely neglected the meridional wind field which is associated with the diurnal variations in the thermosphere. From the work of Kohl and King (1967), and Geisler (1967) it is, however, known that the meridional wind velocities are comparable or even exceed those in the longitudinal direction. For this reason we have most certainly underestimated the wind induced variations in the concentration of atomic oxygen.

In estimating the effects of the meridional wind velocity we assume a meridional wind field of the form

$$V_{\theta}(r, \theta, t) = -V_{\theta}(r) \cos \theta \cos(\omega t - \phi_{\theta}) \quad (15)$$

where $\theta =$ colatitude.

It is the term

$$-\frac{1}{r \sin \theta} \frac{\partial}{\partial \theta} (\sin \theta V_{\theta})$$

that enters into the continuity equation. Evaluating this expression for a velocity field of the form (15) leads then to

$$\frac{V_{\theta}(r)}{r \sin \theta} (\cos^2 \theta - \sin^2 \theta) \cos(\omega t - \phi_{\theta})$$

which becomes

$$-\frac{V_{\theta}(r)}{r} \cos(\omega t - \phi_{\theta}) \quad (16)$$

at the equator.

Considering that longitudinal as well as meridional winds tend to reduce or enhance the density during times when the density is enhanced or reduced respectively it seems reasonable to assume that the effects (divergences) of both wind components are in phase implying $\phi_{\theta} \sim \phi_{\lambda} - 6$ hours. Under the conservative estimate that the magnitudes of the meridional wind velocities are equal to those of the longitudinal wind component the contribution (16) in the continuity equation (11) becomes then equal to the term associated with the longitudinal wind V_{λ} on the right hand side of (11).

It is obvious that with the introduction of a meridional circulation also the vertical wind velocity has to be changed. Here we make the plausible assumption that the magnitude of the vertical velocity field is increased proportional to the increase in the divergence of the horizontal wind velocities which is assumed to be a factor of two.

In Figure 2 we show (marked with superscripts "3D") the wind induced variations in atomic oxygen when we increase the wind field by a factor of two to additionally consider the estimated contribution from the meridional component. Combining it with the temperature effect - which is assumed unchanged - one finds a significant increase of the amplitude below 250 km where the wind effects dominate and a relatively small increase at higher altitudes where the temperature remains the controlling factor.

A very significant effect of the "three dimensional correction" becomes apparent in the phase of atomic oxygen which is further shifted toward noon thus increasing the temperature-density phase difference to nearly 1.5 hours at altitudes as high as 350 km (see Figure 3).

At higher altitudes the phase difference between the oxygen concentration and temperature will of course gradually disappear. However above 600 km H_e becomes the major constituent and H starts to dominate at altitudes above 900 km, and the transport effects discussed in this paper for atomic oxygen are even more significant for the lighter constituents. This is particularly true considering the decreasing significance of the effects of thermal expansion and contraction as the atomic mass decreases. Preliminary calculations for helium have verified this. Exospheric transport and planetary escape further complicate the dynamics and contribute to upset the diffusive equilibrium conditions for these constituents. So it has been observed (Brinton and Mayr (1971) that the hydrogen concentration at 350 km peaks during night (thus exhibiting a density temperature phase difference in the order of 12 hours) consistent with the theoretical predictions of Kockarts and Nicolet (1962), McAfee (1967) and Patterson (1966). Temperature density phase differences would therefore be quite understandable and should be expected especially in the thermosphere at higher altitudes.

SUMMARY AND CONCLUSION

It is recognized that two and three dimensional models of the diurnal variations in the thermosphere predict the day time maxima in the exospheric density and temperature to be essentially in phase and to occur about 2 hours after the maximum in the diurnal heat input. To reproduce the temperature maximum, which - inferred from radar backscatter observations - seems to occur at 16-17⁰⁰ LT,

it is proposed to shift the maximum in the diurnal heat input from 12⁰⁰ to about 13⁰⁰ LT; a phase shift of this magnitude being reasonable and probably conservative considering that a.) the solar heat input is proportional to the atmospheric density (in the optically thin region) which peaks near 14⁰⁰ LT and b.) the thermospheric heat input originating from dissociative recombination is proportional to the electron density which peaks near 16⁰⁰ LT.

Assuming that the thermosphere dynamics of the diurnal component is generated primarily by the internal heat input within the thermosphere, the two dimensional model of Volland and Mayr (1970) is then adopted to describe the diurnal variations of the temperature, and the density and wind fields for N₂, a constituent which can be assumed in diffusive equilibrium below 250 km.

Assuming that the diffusion velocities become zero at higher altitudes and that, due to the long diffusion time the density variations vanish at lower altitudes, the momentum and time dependent continuity equations for O are solved. With consideration of the temperature field and the N₂-O drag interaction from the windfield, the diffusion effects in the diurnal variations of atomic oxygen are discussed.

While at higher altitudes the temperature effects prevail, it is shown that the wind induced diffusion effects dominate the diurnal variations in O in the thermosphere below 200 km thus producing deviations from diffusive equilibrium that cannot be neglected.

It is shown that, through the interaction of horizontal winds, atomic oxygen is removed from the day side and supplied toward the night side of the earth with the vertical diffusion partially supplying O from below during day and supplying it toward lower altitudes during night. In this rather complicated redistribution the horizontal transport prevails at altitudes above 140 km thus shifting the phase of the wind induced variations in O toward the late morning hours (10^{30} LT).

Combining this diffusion component - which has so far been neglected - with the temperature effects, amplitude and phase in the diurnal variations of atomic oxygen are significantly changed. Two of the most important consequences are summarized below:

- 1) The phase (time of maximum) of O is shifted toward earlier hours thus producing a phase difference between temperature and density which is consistent with the phase discrepancy apparent from satellite drag data of the density and radar backscatter observations of the exospheric temperature.
- 2) Atomic oxygen and molecular nitrogen are shown to be not in phase, with the maximum in O occurring a few hours before that in N_2 at F_2 region heights. This contributes to a reduction in the amplitude of the $[O]/[N_2]$ ratio by about 20% and shifts the maximum of this ratio into the morning hours, effects both of which could significantly contribute to the maintenance of the night time F_2 region. (We gratefully acknowledge a discussion with Dr. J. W. King in which he pointed out to us the significance of these composition changes for the ionosphere.)

Although the model presented here can explain in a self consistent form the density-temperature phase discrepancy that appears to exist in the

thermosphere, our model should not be considered as unique. Apart from the deficiency that the three dimensional (r, λ, θ) nature of the thermosphere dynamics was not considered rigorously, our model was based on the assumption that the diurnal variations of the thermosphere are primarily excited by the UV heat source within the thermosphere. This is a very important qualification since one can show (as has been shown by Mayr and Volland (1970) in the case of the storm time behavior in the composition) that significant energy coupling with the lower atmosphere could change amplitude and phase in the diurnal variations of atomic oxygen such as to reduce or even reverse the density-temperature phase difference. Furthermore, we have not considered the semidiurnal component in the temperature and density variations which, although small could perhaps significantly determine the time difference between temperature and density maxima (Volland, 1971).

REFERENCES

1. Brinton, H. C., and H. G. Mayr, "Temporal Variations of Thermospheric Hydrogen Derived from In Situ Measurements," J. Geophys. Res., 76, 25, 6198-6201, 1971
2. Chandra, S. and P. Stubbe "The Diurnal Phase Anomaly in the Upper Atmospheric Density and Temperature" Planet. Space Sci., 18, 1021-1033, 1970
3. Dickinson, R. E., Lagos, C. P. and R. E. Newell "Dynamics of the Neutral Thermosphere for Small Rossby Number Motions" Journ. Geophys. Res., 73, 4299-4313, 1968
4. Geisler, J. E., "A Numerical Study of the Wind System in the Middle Thermosphere," Journ. Atm. Terr. Phys., 29, 1429-1482, 1967
5. Harris, I., and W. Priester, "Time Dependent Structure of the Upper Atmosphere," Journ. Atm. Sci., 19, 286-301, 1962
6. Jacchia, L. G., "Static Diffusion Models of the Upper Atmosphere with Empirical Temperature Profiles," Smithonian Contr. to Astrophys., 8, pp. 215, 1965
7. Kockarts, G., and M. Nicolet, "Le Problème Aéronomique de Helium et de L'hydrogène Neutres," Ann Geophys., 18, 269, 1962
8. Kohl, H. and J. W. King, "Atmospheric Winds Between 100 and 700 km and their Effects on the Ionosphere," Journ. Atm. Terr. Phys., 29, 1045-1062, 1967

9. Lindzen, R. S., "Thermally Driven Diurnal Tide in the Atmosphere," Quart. Journ. Ray. Meteorolog. Soc., 93, 18-42, 1967
10. Mahajan, K. K., "Diurnal Variation of the Ion Temperature," Journ. Atm. Terr. Phys., 31, 93-101, 1969
11. Mayr, H. G. and H. Volland, "Magnetic Storm Effects in the Neutral Composition," NASA Document X-621-71-143, 1971, EOS, 51, 1970, accepted for publication in Planet. Space Sci.
12. Mayr, H. G. and H. Volland, "Semi-annual Variations in the Neutral Composition," NASA Document X-621-70-478, 1971, EOS, 42, 1971 (Submitted to Ann. Geophys.)
13. McAfee, J. R., "Lateral Flow in the Exosphere," Planet Space Sci., 15, 599, 1967
14. Patterson, T. N. I., "The Diurnal Variation of the Atomic Hydrogen Concentration at the Base of The Exosphere," Planet. Space Sci., 14, 425, 1966.
15. Spencer, N. W., "The Neutral Atmosphere - Consistency of Experimental Data," EOS, 52, 1971
16. Stubbe, P. and S. Chandra, "Ionospheric Warming by Neutral Winds," Planet. Space Sci., 19, 1971
17. Volland, H. and H. G. Mayr "The Significance of the Wave Picture in the Theory of Diurnal Tides Within the Thermosphere," NASA-Documnt X-621-68-466, 1968, EOS, 50, 1969

18. Volland, H., "Phase Delay Between Neutral Temperature and Neutral Density at Thermospheric Heights," J. Geophys. Res., 75, 28, 5618-5620, 1970
19. Volland, H., and H. G. Mayr, "A Theory of the Diurnal Variations in the Thermosphere," Ann. Geophys., 26, 907-919, 1970
20. Volland, H. and H. G. Mayr, "A Numerical Study of a Three Dimensional Thermospheric Density and Wind Model" (in preparation 1971 a)
21. Volland, H. and H. G. Mayr, "The Degeneration of the Hough Functions Within the Thermosphere," Presented at COSPAR, paper C. 6. 2., 1971 b, to be published in Space Research XII
22. Walker, R. E., "Measurement of 0-0 Diffusion Coefficient," J. Chem. Phys., 31, 2197-2197, 1961

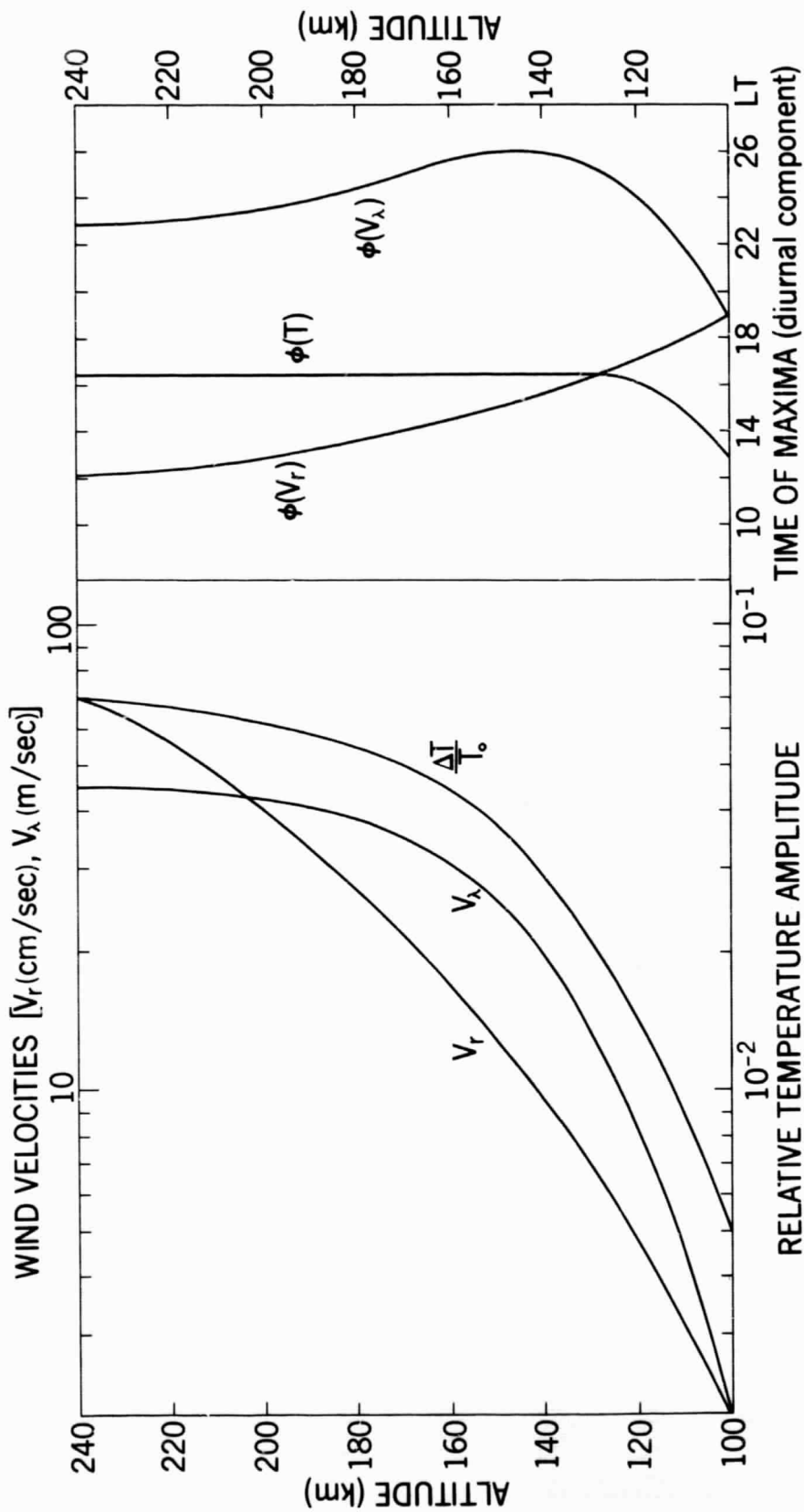


Figure 1. Amplitude (a) and phase (b) of the temperature ($\Delta T/T_0$) and wind fields (V_r, V_x), adopted from the two dimensional model component of Volland and Mayr (1970) in which only the UV heat input within the thermosphere was considered. Note that the phase has been shifted by one hour.

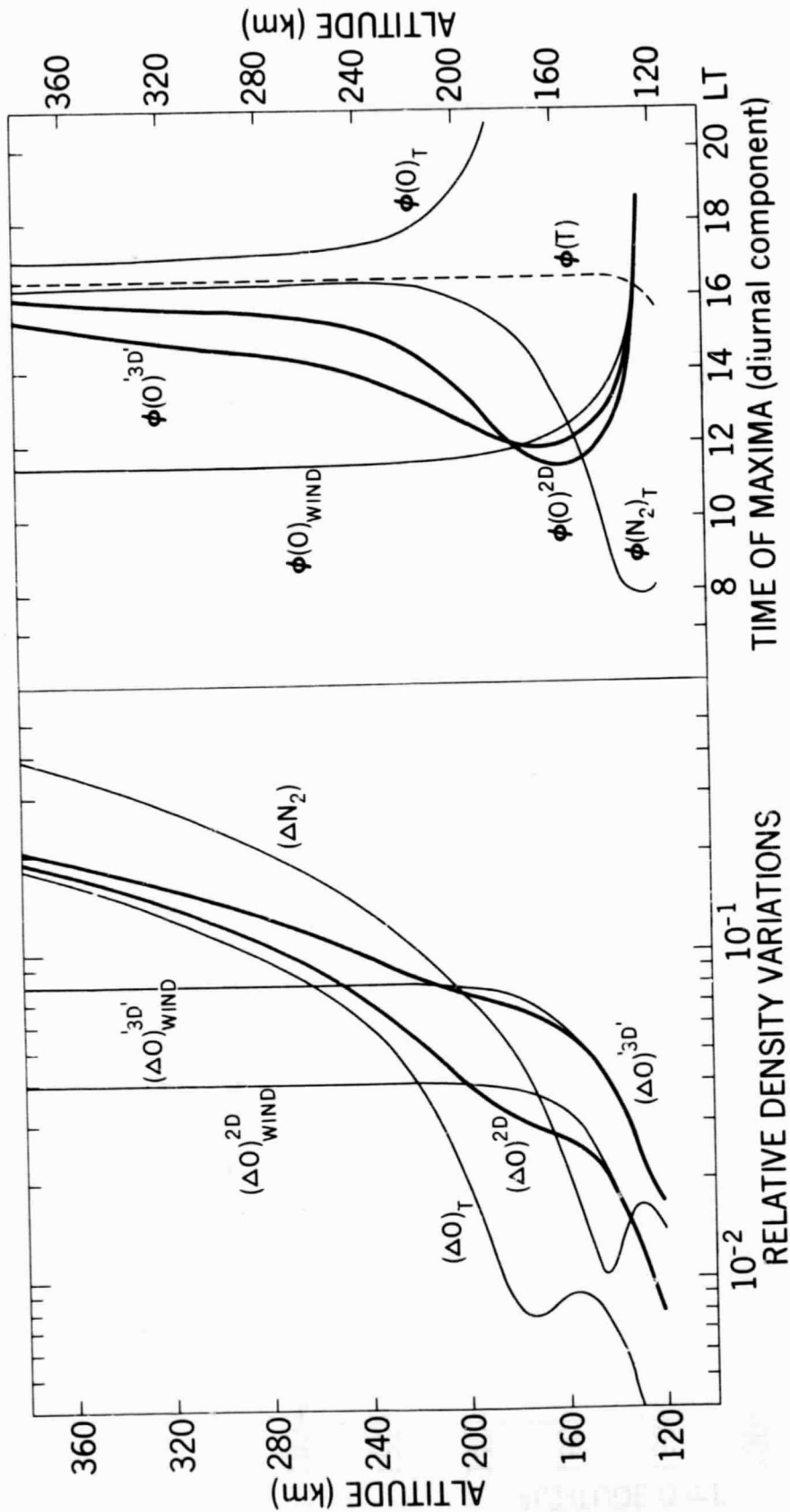


Figure 2. Amplitude (a) and phase (b) of the various density components: $(\Delta O)_T$ and $\phi(O)_T$ = amplitude and phase of the temperature induced variations, $(\Delta O)_{WIND}$ and $\phi(O)_{WIND}$ = amplitude and phase of the wind induced variations (ΔO) and $\phi(O)$ = sum of both temperature and wind components. The superscripts 2D and "3D" indicate the results from the two dimensional and three dimensional (estimated) models. For comparison are shown the amplitude (ΔN_2) , and phase $\phi(N_2)_T$, for the N_2 density and the phase for the temperature (dashed line) $\phi(T)$, from the model of Volland and Mayr (1970).

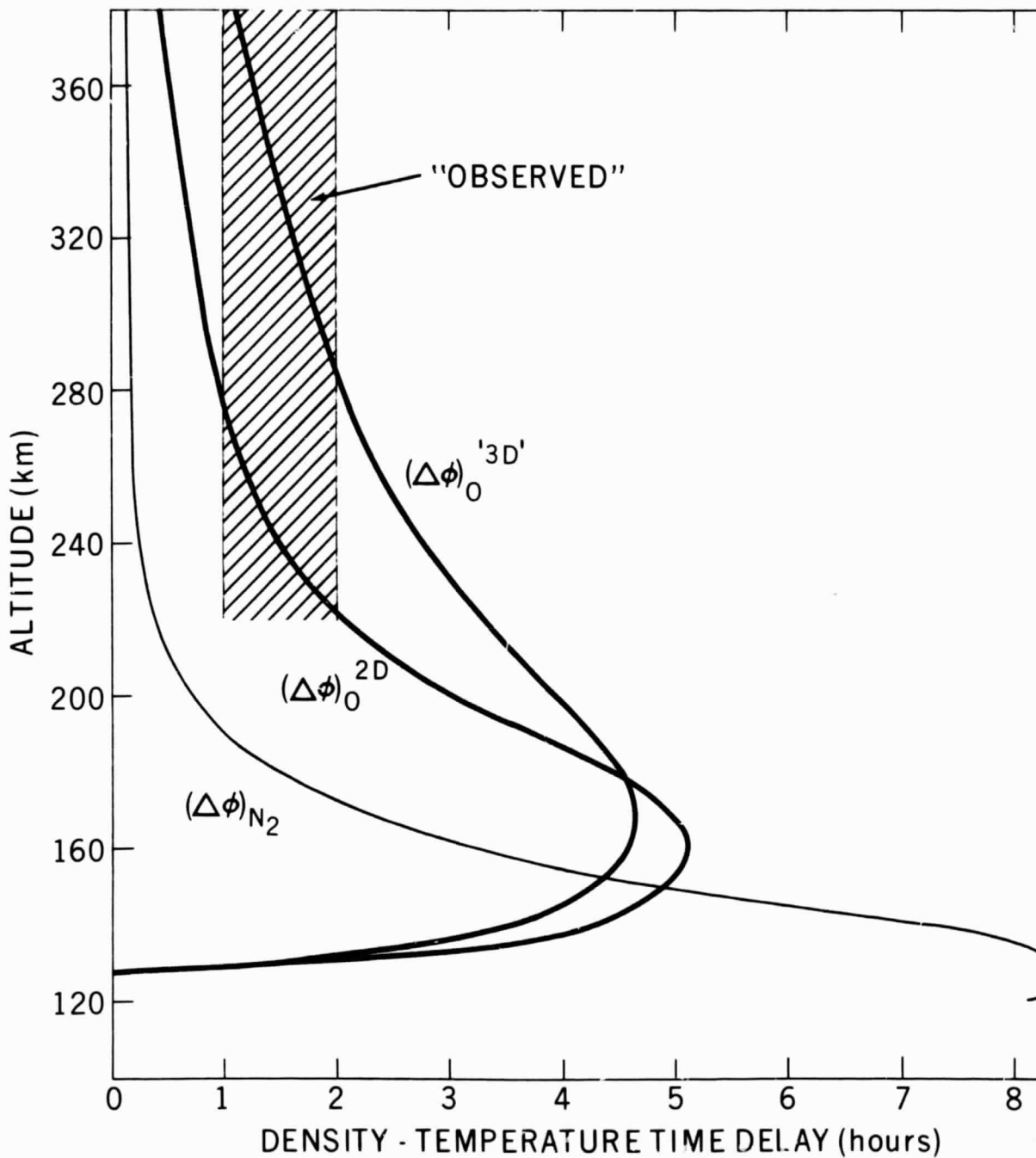


Figure 3. The density - temperature time delay. Note that above 250 km, where O is the major constituent, the time difference is between one and two hours, in agreement with "observational" evidences.