1. Publication Nº	2. Version	3. Date	5. Distribution
INPE-3823-PRE/899		*March, 1986*	☐ Internal ☒ External
4. Origin *DGA* — Program *DIO/IONO*			☐ Restricted

6. Key words - selected by the author(s)
THERMOSPHERIC MODELS
NEUTRAL ATMOSPHERE BEHAVIOR

7. U.D.C.: *523.4-8*

8. Title — *INPE-3823-PRE/899*

THERMOSPHERIC MODELS: A SHORT REVIEW ON THE BASIC ASSUMPTIONS

10. Nº of pages: *34*

11. Last page: *32*

12. Revised by

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Director General

14. Abstract/Notes

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15. Remarks *This work is being submitted to Reviews Geophysics and Space Physics.*

THERMOSPHERIC MODELS: A SHORT REVIEW ON THE BASIC ASSUMPTIONS

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ABSTRACT

Thermospheric models vary in a wide range of approaches: theoretical, analytical, semiempirical, numerical etc. The models should be used only if the basic assumptions, on which they are based, are valid. This paper deals with a critical review of thermospheric models keeping in view that they are designed to predict the actual behavior of the neutral atmosphere parameters above 90 km. Emphasis is put on the lower thermosphere to avoid interhemispheric transport effect, not included in most of the models.

1 - INTRODUCTION

In the past twenty years considerable effort has been made to establish upper atmosphere models. The idea is to build up a reference so that regular and irregular behavior of the neutral atmosphere may have a well-defined meaning. Tidal studies (e.g. Chapman and Lindzen, 1970; Volland and Mayr, 1970; Lindzen, 1970) contributed to give a definite meaning to static and dynamic regular behavior of the neutral atmosphere. It must be emphasized that static conditions do not exist in the upper atmosphere. Therefore, theoretical or static models constitute an idealized reference, which must be obtained when the dynamics is removed. The problem would be very simple if the system of conservation equations could be solved completely. This has not been feasible so far. The reasons are manifold and probably the large number of indetermined parameters, together with multiple interactions, will still restrict the solutions for a long time.

It is generally accepted that the conservation equations (of mass, momentum and energy), together with an equation of state, constitute a reasonable mathematical concept of the problem. The equations are those of a fluid and the neutral gas treated as such. The various approaches result from the different simplifications imposed on these equations to obtain a solution. They will be presented and discussed in this work. We will restrict our comparison to the lower thermosphere range which is the less studied region. Even the sophisticated approaches (e.g. Dickinson et al., 1981) are not guaranteed to work in this region.

2 - THE BASIC EQUATIONS

The three basic equations to describe the behavior of the neutral atmosphere are: the continuity equation and the equations for conservation of momentum and energy.

The continuity equation is expressed as:

$$\frac{\partial\rho}{\partial t} + \nabla \cdot (\rho \underline{u}) = 0 , \qquad (1)$$

where ρ is the total density, t is the time and $\underline{u}$ stands for the wind velocity. It assumes that no sources or sinks of neutral particles need to be considered. Ionized particles which enter or leave the upper atmosphere do not significantly alter the balance expressed by (1) and may be neglected[1].

The conservation of momentum is written as:

$$\frac{d(\rho \underline{u})}{dt} = \Delta\underline{f} , \qquad (2)$$

where $\Delta\underline{f}$ expresses the balance of external forces per unit volume applied to the volume element under consideration. These forces are usually:

a) the Coriolis force given by:

$$2\rho \, \underline{\Omega} \times \underline{u}$$

b) the ion drag given by:

$$\rho \, \nu_{ni}(\underline{v}_i - \underline{u})$$

---

[1] This assumption is not strictly valid when interhemispheric transport affects significantly the value of ρ like in the lower thermosphere at midlatitudes (Waldteufel, 1970).

c) the volume weight given by:

$\rho \underline{g}$

d) pressure acting on the given volume:

∇p

e) viscosity forces

$\mu \nabla^2 \underline{u}$

where $\underline{\Omega}$ is the earth's sidereal angular velocity, ν_{ni} is the neutral-ion collision frequency, $\underline{v}_i$ is the ion velocity, $\underline{g}$ the acceleration of gravity, p the pressure and μ the coefficient of viscosity. One must note that only one ion mass is being considered. No account is made for perturbations produced by interhemispheric flux of particles.

The conservation of energy is expressed by:

$$\frac{dW}{dt} = \Delta Q \ , \tag{3}$$

where ΔQ indicates the balance of energy and W stands for total energy in the volume considered. The various components of ΔQ are:

a) transported energy:

$(W+p) \ \nabla \cdot \underline{u}$

b) work done:

$\underline{u} \cdot \Delta \underline{f}$

c) rates of thermal energy gain and loss:

$$(Q_n - L_n)$$

d) thermal energy transported:

$$\nabla \cdot \underline{\phi}$$

where $\underline{\phi}$ is the thermal flux.

The energy equation presented here also does not account for any flux of particles.

To complement these equations, one has the equation of state:

$$p = \rho RT = \rho R_0 T/M , \qquad (4)$$

where R_0 is the universal gas constant and M is the mean molecular mass in a.m.u.

3 - HYDROSTATIC EQUILIBRIUM

The first unidimensional attempt to model the neutral atmosphere consists in considering it as a static ($\underline{u} = \underline{0}$), perfect gas in equilibrium ($\partial/\partial t = 0$). The basic equations reduce to:

$$\rho = \rho (z) , \qquad (5)$$

$$\nabla p = \rho \underline{g} , \qquad (6)$$

$$\overline{W} = (Q_n - L_n) - \nabla \cdot \underline{\phi} , \qquad (7)$$

$$p = \rho RT , \qquad (8)$$

where the bar denotes average values.

The system (5), (6) and (8) admits the solution:

$$p = p_0 \exp\left(-g R^{-1} \int_{z_0}^{z} T^{-1} ds\right) , \qquad (9)$$

where T is determined from equation (7), z_0 is a reference altitude, p_0 the pressure at z_0 and M and g were taken constant. The flux ϕ is given by:

$$\phi = -K \frac{\partial T}{\partial z} , \qquad (10)$$

where K is the coefficient of molecular heat conductivity. The average energy is the internal energy $\rho C_v T$ (C_v = specific heat at constant volume).

In principle the solution of (7), (9) and (10) should give the desired static reference. In practice, due to difficulties with the measured solar flux and complete knowledge of heat sources, we may not expect to obtain a reliable result with this approach.

4 - HARRIS AND PRIESTER MODEL

The first attempt to improve the static model, by introducing some dynamics, was the model of Harris and Priester (1962). They allowed for a time variation and motion along the vertical to accomodate the gas to the perturbation caused by the day-to-night alternation. This was the concept of a "breathing atmosphere".

The model preserves the solution (9), from the equations of motion and state, to describe the behavior of the pressure. This implies that the acceleration of the resulting motion is negligible compared to that of gravity. Moreover, the upper limit of integration in (9) changes with time as: $z + u_z \Delta t$.

To simplify the energy equation this model imposes the "following the cell" approach considering a test volume which moves vertically, all the while remaining at constant pressure. An additional assumption is that the kinectic energy, produced by the motion, does not significantly alter the total energy in the test volume. This is equivalent to saying that the internal energy, $\rho C_V T$, predominates. Equation (3) reduces then to:

$$\frac{\partial T}{\partial t} + u_z \frac{\partial T}{\partial z} + \gamma T \frac{\partial u_z}{\partial z} = (\rho C_V)^{-1} \left[(Q_n - L_n) - \frac{\partial \phi}{\partial z} \right] , \qquad (11)$$

where γ is the ratio between specific heats. The original version of Harris and Priester (1962) has a different coefficient for the accelaration term. The difference came from the simplification used by Eckart (1960) to determine the energy balance equation.

From the condition $dp/dt = 0$,

$$u_z = -(\partial p/\partial t) \,/\, (\partial p/\partial z) \,, \qquad (12)$$

where the partial time derivative is obtained from (9) and the partial space derivative comes from the equation of motion (6). It then follows:

$$u_z = T \int_0^z \frac{1}{T^2} \frac{\partial T}{\partial t} \, ds \,. \qquad (13)$$

From the continuity equation and the equation of state we get:

$$\frac{\partial u_z}{\partial z} = \frac{1}{T} \frac{dT}{dt} \qquad (14)$$

which, when substituted into (11) using also (13), yields:

$$(\gamma+1)\left\{\frac{\partial T}{\partial t} + \frac{\partial T}{\partial z} T \int_0^z \frac{1}{T^2} \frac{\partial T}{\partial t} dz\right\} = (\rho C_v)^{-1}\left[(Q_n - L_n) - \frac{\partial \phi}{\partial z}\right] . \qquad (15)$$

Numerical integration of (15) using experimental results for $\partial T/\partial t$ gives the desired temperature profile. The original version (Harris and Priester, 1962) also ran into troubles with appropriate energy balance.

5 - EMPIRICAL MODELS

In 1964 started a series of models by Jacchia (e.g. Jacchia, 1964, 1971, 1977). These models used essentially the hydrostatic equilibrium approach but substituted the solution of equation (7) by empirical temperature profiles. The purpose was to overcome the difficulties of the energy balance equation. Experimental data were considered in order to establish average temperature profiles for a variety of parameters: local time, latitude season, etc.

The empirical models considered the diurnal and seasonal variations to account for the effect of tidal and planetary atmospheric waves. Also, the temperature data were usually obtained by indirect means (see Hedin et al., 1977).

To some extent, the empirical models proved to be more reliable in the description of the thermospheric behavior and are usually preferred when the neutral atmospheric parameters are required. The inclusion of an adjustable latitudinal dependence is partially responsible for this success. The reason is that interhemispheric transport effects (Young et al., 1980; Chandler, 1983) participate somehow in the data fitting. A discussion of the empirical models more frequently used in recent literature can be found in Jacchia (1978).

6 - THE ANALYTICAL APPROACH

The analytical approach consists in assuming an analytical expression to describe the temperature behavior, instead of trying to solve the energy equation. In this aspect, it is similar to the empirical models. The parameters of the temperature profile are chosen to match actual temperature data. The difference between this and the empirical approach is that here we look for simple temperature expressions, so that all the parameters may be determined by analytical expressions.

The expression which describes the temperature behavior, in the thermosphere, was suggested by Bates (1959) to be:

$$T(z) = a - b \exp(-c.z) \tag{16}$$

and has been used since then. Here a, b and c are parameters to be established in order to adjust the model to the desired condition. The advantage of equation (16) is that the parameter a fits the upper limit condition (exospheric temperature), c adjusts the slope and b completes the matching for the lower limit condition. Therefore, (16) is a reasonable average description of the thermospheric temperature.

These models still consider expression (9) of the hydrostatic equilibrium as valid without restriction. From it and the equation of state, Walker (1965) determined an analytical solution for the density. Alcayde (1981) made one more effort in favor of analytical models by including the influence of the mesosphere and stratosphere.

The approach cannot compete with empirical models if more realistic temperature and densities are desired. There are several reasons for that; but the critical one is that c does not remain constant for the entire range of altitudes. These models are also frozen in time since they do not account for the diurnal, regular variation of temperature. This, however, is not a critical restriction since time variation may be easily included in the exospheric temperature. The inclusion of other parameters like latitude, season, etc. increases the complexity of the analytical models to the point that they coincide with the empirical models. This can easily be seen if we compare Alcayde (1981) model with the MSIS model (Hedin et al., 1977).

7 - HORIZONTAL TRANSPORT MODELS

The earliest models were characterized by a net vanishing horizontal flow. A new trend appeared when Lindzen (1966) and Lagos and Mahoney (1967) emphasized the importance of horizontal energy transport to account for the diurnal temperature and density variations.

The first attempts (Johnson and Gottlieb, 1970; Stubbe, 1970) considered just one horizontal component (meridional or zonal) and a net nondivergent contribution. This requires one more component of the vertical wind to balance with the horizontal contribution. It yields:

$$u_{z_1} = - \frac{1}{\rho(z)} \int_{z_0}^{z} \rho(s) \frac{\partial u_y}{\partial y} \, ds \qquad (17)$$

$$u_{z_2} = - \frac{\lambda z}{r} \, \overline{u}_x \, \mathrm{tg}\, \Psi \qquad (18)$$

whether one considers zonal flow of mass (Stubbe, 1970) or meridional transport of energy (Johnson and Gottlieb, 1970). Here the bar denotes daily time average, λ_z the vertical wavelength r denotes the radial distances from the center of the earth to the considered altitude and Ψ relates to the latitude Θ as:

$$\Psi = \Theta - 23.5 \cos \left[\frac{2\pi}{365}(t - \tau_e)\right] ,$$

where t is the day of the year and τ_e is the day of the vernal equinox. The angles are expressed in degrees.

Expression (17) accounts for day to night asymmetries in the distributions of the solar energy whereas expression (18) accounts for winter to summer differences.

The importance of horizontal winds comes from their contribution to the vertical velocity which produces adiabatic warming or cooling of the neutral gas. This is easily seen from the contribution of the term $\underline{u} \cdot \Delta \underline{f}$ in the energy equation (see Section 2). The barometric equation is still being considered as a valid approach for the local momentum equation.

The importance of adiabatic thermal variations was pointed out by Dickinson et al. (1968) and Dickinson and Geisler (1968). They compared the magnitude of these thermal variations with those produced by the solar heating.

8 - MODELS BASED ON TIDAL THEORY

The importance of atmospheric motions to the heat balance can be readily seen in the energy balance equation. Systematic experimental studies, of atmospheric motion, have been carried out for a long time (e.g. Greenhow and Neufeld, 1961; Spizzichino, 1969a, b, c; 1970a, b; Amayenc and Reddy, 1972; Zamlutti, 1975; Salah et al., 1975; Fontanari et al., 1983). It was concluded that tidal and planetary waves dominate the dynamics of the neutral atmosphere. Theoretical studies of long period waves have been reviewed in its classical form by Lindzen and Chapman (1969) and considered in more recent theoretical works by Forbes and Garrett (1979). The importance of these waves to the heat balance also received some attention (e.g. Hines, 1965; Dickinson and Geisler, 1968; Groves, 1982).

The basic approximations used in tidal theory (Lindzen and Chapman, 1969) are the validity of the fluid tidal approach, the preservance of hydrostatic equilibrium and the possibility of applying the perturbation theory, to account for the presence of the oscillatory components in the neutral atmosphere parameters. Volland and Mayr (1977) restricted the problem further by considering the Rayleigh friction to approach the viscous force, the Newtonian cooling to replace the heat flux and the ion drag term, the dominant one in the interaction between charged and neutral particles. These approximations had been considered in Volland and Mayr (1977) and will not be discussed here.

With the considerations of the last paragraph our system of equations generates two system of perturbed equations:

$$- i\omega \underline{u}_h + \nu^* \underline{u}_h + 2\underline{\Omega} \times \underline{u}_h + (\nabla_h \delta p) / \bar{\rho} = 0 \qquad (19)$$

$$bg\bar{\rho}\nabla_h \cdot \underline{u}_h - i\omega\delta p = 0 \qquad (20)$$

from the horizontal momentum equation and:

$$-i\omega\delta\rho + \nabla_z \cdot (\rho u_z) + \nabla_h \cdot (\rho \underline{u}_h) = 0 \qquad (21)$$

$$\nabla_z \delta p + \underline{g}\delta\rho = 0 \qquad (22)$$

$$C_v \left(1 - \frac{\nu_n \gamma}{i\omega} \right) [-i\omega\delta T + u_z \nabla_z \delta T] - \frac{\bar{p}}{\bar{\rho}^2} [- i\omega\delta\rho + \nabla_z \delta\rho] = \delta Q \qquad (23)$$

$$\delta p/\bar{p} = \delta\rho/\bar{\rho} + \delta T/\bar{T} \qquad (24)$$

from the vertical conservation equations (Volland and Mayr, 1977). Here, the subscripts z and h stand for vertical and horizontal components respectively, the prefixes δ denote perturbations and the bar represents basic state of the parameter (from hydrostatic equilibrium). Moreover, sinusoidal variations with time denoted by the prefixes $-i\omega$ were assumed. The other symbols are: ν_n the Newtonian cooling coefficient; $\nu^* = \nu_{ni} + \nu_i$, where ν_{ni} is the neutral -ion collision frequency; ν_i the Rayleigh friction coefficient; $\gamma = C_p/C_v$, where C_p and C_v are specific heats at constant pressure and volume respectively and b is a separation constant.

The solution of (19) and (20) which constitute a Laplace system of equations has the form:

$$\delta p = \tilde{p}_\alpha^\beta(z)\, H_\alpha^\beta(\Theta)\, \exp[i(\beta\phi-\omega t)] \,, \tag{25}$$

Volland and Mayr (1977). Here Θ stands for colatitude, ϕ is the longitude, H_α^β represents the Hough function and $\tilde{p}_\alpha^\beta$ is the amplitude of the oscillation. The indices α and β represent the number of complete periods in the meridional and zonal directions respectively. The velocity components are expressed as:

$$u_x = \left\{-\frac{1}{a}\frac{\partial\Phi}{\partial\Theta} + \frac{i\beta}{a \text{ sen } \Theta}\Psi\right\} \tilde{p}_\alpha^\beta \exp[i(\beta\phi-\omega t)] \tag{26}$$

$$u_z = \left\{-\frac{i\beta}{a \text{ sen } \Theta}\Phi + \frac{1}{a}\frac{\partial\Psi}{\partial\Theta}\right\} \tilde{p}_\alpha^\beta \exp[i(\beta\phi-\omega t)] \tag{27}$$

(Volland and Mayr, 1977). Here a is the earth's radius, Φ and Ψ functions to be determined in terms of the Hough functions from the Laplace equations.

The solution of (21)-(24) yields the amplitude of perturbations in pressure, temperature, density and vertical winds along the vertical (Volland and Mayr, 1977).

We turn now to the actual effect of atmospheric oscillations on the heat balance. Hines (1965) estimated an energy input rate due to atmospheric oscillations of:

$$Q_a = \rho u_h^2 \; u_z \; (2 H_o) \; , \tag{28}$$

where H_o is the scale height of the vertical variation of the energy ρu_h^2.

If we consider a fixed latitude, longitude and instant of time, expressions (26) and (27) may be rewritten as:

$$u_x = c_1 \; \tilde{p}_\alpha{}^\beta(z) \; , \tag{29}$$

$$u_y = c_2 \; \tilde{p}_\alpha{}^\beta(z) \; , \tag{30}$$

where c_1 and c_2 are constants. It then follows that:

$$\rho u_h^2 \; \propto [p_\alpha{}^\beta(z)]^2 \; . \tag{31}$$

In order to estimate the vertical variation of the energy density in (31) we consider that:

$$\nabla_z \cdot (\rho \underline{u}_z) = \eta \; \nabla_h \cdot (\rho \underline{u}_h) \; , \tag{32}$$

where η measures the contribution of the horizontal divergence to the vertical motion of air. At a surface of constant pressure $\eta = -1$. It then follows from (20), (21) and (32) that:

$$\delta\rho = \frac{(1+\eta)}{bg} \, \delta p \; . \tag{33}$$

If we now consider (22) and (33) we get:

$$\delta p \;\alpha\; \tilde{p}_{\alpha}^{\beta} \;(z_0)\; \exp\left[-\frac{(1+n)z}{b}\right] . \tag{34}$$

Usually b has a complex value (Volland, 1974; Volland and Mayr, 1977) from which results an upward propagating part as well as an attenuation factor in the exponential component of (34). Therefore from (31) and (34) one can estimate the value of H_0 for (28). From Volland (1974) one gets for very low collision frequency a purely attenuated wave with $H_0 = 5(1+n)^{-1}$ km.

Dickinson and Geisler (1968), considering that the kinectic energy dissipation results from the work done, compute the temperature variation assuming the velocity pattern established by pressure gradients. To some extent this is a semiempirical approach. The results are expected to be equivalent to those of Hines (1965) method since the loss of kinetic energy is what actually produces the net atmosphere heating.

One last remark about the tidal oscillations contribution to thermospheric modeling is that, regardless of the sign of the vertical velocity, expression (28) refers to effective heating on the unperturbed atmosphere (Hines, 1965). Therefore, a basic state temperature can be determined by:

$$\bar{\rho}\, C_p\, \bar{T} \;=\; \langle (Q_n - L_n) + \rho u_h^2 \;|u_z|\; / \;2H_0 \rangle , \tag{35}$$

where $\langle \cdots \rangle$ stands for time average. The value computed with (35) should reproduce the values obtained with static empirical models but this, in fact, does not occur. The presence of more than one oscillatory frequency is pointed out as the cause of the discrepancies For a detailed study on mode interaction and its effects, the reader is referred to the series of papers by Forbes (Forbes 1982a,b; Forbes and Champion, 1982). Seasonal waves also contribute to the diurnal balance of energy by adiabatic transport.

9 - NUMERICAL MODELS

Numerical models are produced to study the time-dependent behavior of the thermosphere. Emphasis on time dependent models increased after the works by Dickinson et al.(1968) and Dickinson and Geisler (1968). This last work provides a very good insight on the physical balance of the parameters affecting the thermospheric dynamics.

More recent time dependent models use the complete set of hydrodynamic and thermodynamic equations which increase the complexity to a point where no analytical solution becomes possible (see Fuller-Rowell and Rees, 1980). Nevertheless the complexity of such an approach, the same problem found in the earliest (Harris and Priester, 1962) model persists (viz., the insufficiency of solar fluxes to yield actual temperatures). Some authors decided to enhance the solar fluxes by a factor ranging from 1.5 to 2.0 (e.g. Dickinson et al. 1975; Fuller-Rowell and Rees, 1980), whereas others opted in favor of giving up the thermodynamic equation by assuming an empirical temperature profile (e.g. Fontanari et al., 1982; Sutherland and Zinn, 1983). With this restriction numerical models have been quite successful to describe the actual time-dependence of neutral atmosphere parameters.

10 - CURRENT STATUS

Since the earliest days of atmospheric modeling, the models keep increasing in complexity, in order that self-consistent approaches (which include all the equations) could reproduce actual dynamic phenomena. A satisfactory stage was finally achieved with three-dimensional models like that of Dickinson et al. (1981). A question still persists: are these necessary? The recent work by Torr et al. (1980) brought a new light to the problem. While Stolarski et al.(1975) estimate the heating efficiency to be around 30%, Torr et al. (1980) suggested that this efficiency could actually go up to 50%, near the peak of heat transfer to the neutrals. A comment on the effect of these new heating efficiencies, in thermospheric modeling, was made by Dickinson et al. (1981).

Measurements of lower thermosphere parameters with increased accuracy are now available (e.g. Zamlutti, 1973; Zamlutti and Farley, 1975; Harper, 1977, 1978; Mathews, 1976; Wand, 1976; Evans, 1979; Oliver, 1980). These data and references therein complement the earlier incoherent scatter data from the French radar used in the MSIS model (see Hedin et al., 1977). It is then time for a critical analysis on the reliability of simple models and on the validity of the basic assumptions used in the various approaches.

11 - REVIEW ON THE BASIC ASSUMPTIONS

We start this section with the static modeling. Pratically all the thermospheric models consider the validity of the barometric equation, which must be recovered if diurnal and seasonal effects are removed. An approach to the unperturbed conditions holds for the diurnal average of the atmospheric parameters during equinoxes. It is then possible to test the barometric equation under these circumstances.

Considering the results of Zamlutti (1973), obtained from data collected at 105 km during November 7, 1972 at Arecibo (18^oN), we have:

$\bar{T} = 223^oK$

$\bar{\nu}_{in} = 970$ coll/sec

$T' = 10^oK/km$ $\qquad\qquad$ $\tilde{T} = 4^oK$

$\nu'_{in} = -200$ coll/km $\qquad\qquad$ $\tilde{\nu}_{in} = -124$ coll/sec .

Using these values in the equation of state we obtain:

$$\frac{p'}{\overline{p}} = \frac{\nu'_{in}}{\overline{\nu}_{in}} + \frac{T'}{\overline{T}} = -0.161 .$$

Computing $p'/\overline{p}$ with the barometric equation, using a mean molecular mass of 28 a.m.u., we get:

$$p'/\overline{p} = -0.148 .$$

Comparing the two results it may be observed that the discrepancy is only 8.5%, which lies within the expected error bars for these calculations. Therefore, to the extent of the available data the barometric equation can be assumed to hold, at the lower thermosphere.

The consistency of Zamlutti (1973) data with other results obtained at the same site is easily verified. From the results obtained by Wand (1976) we see that the values employed above agree within 10% accuracy.

Assuming that the analytical expression (16) can describe the average temperature variation with altitude, above 110 km, we computed an exospheric temperature range of 900^{o}K-950^{o}K, with the temperature results published by Zamlutti (1973, 1975) for November 7, 1972. The computations are subjected to large errors since a small number of points could be used. Nevertheless, a reference value of 1000^{o}K exospheric temperature was considered by Zamlutti (1973, 1975) for comparing his results with empirical models. It was realized that above 105 km the slope of the Jacchia (1970) model was less severe than that of actual data. Figure 1 presents a comparison of Zamlutti (1973) data and a few models. It can be observed that, in spite of the large errors involved, the Bates type expression (16) yields a fairly good matching with exospheric temperature of 900^{o}K. Temperature models, above 100 km, based on expression (16) are then quite reliable.

The data used to compare temperature models are consistent with that obtained by Wand (1976) within 10% accuracy. The results obtained here agree with those used before by Waldteufel and Cogger (1971) as far as the slope of the exponential function is concerned. A discrepancy is observed in the temperature at 120 km. Waldteufel and Cogger (1971) use a reference temperature of 350^{o}K which agrees with some experimental data utilized to compare the MSIS-83 model (see Forbes, 1985 and Hedin, 1983). Our value of 440^{o}K is in agreement with those of Wand (1976). Since the 120 km level has no physical significance there is no reason to believe that the temperature would remain constant at it.

Next we consider the aspects of dynamic modeling, by going one step further in complexity in the solution of the energy equation. We first investigate the validity of simple models, like that of Harris and Priester (1962). This model holds if the vertical velocity can be determined by the expression:

$$u_z = -(\partial p/\partial t) / (\partial p/\partial z) .$$

Using Zamlutti (1973) results and the equation of state it is possible to test this expression. Some caution is necessary while using the data. Below 105 km the incoherent scatter velocity measurements lie within the error bar limits and no comparison is possible. Above 115 km the collision frequency measurements are subject to large errors and the computations are not reliable. A compromise solution was considered, for the present comparison, by using the data for 112.1 km obtained on November 7, 1972 (Zamlutti, 1973). The computed velocity amplitude using Harris and Priester expression is u_z = - 0.15 m/s, whereas incoherent scatter measurements indicate $u_z \cong 5$ m/s. Therefore, an unidimensional approach to the energy equation is not sufficient.

The values of vertical velocities used here, from Zamlutti (1973), proved to be consistent with subsequent results of a series of works by Harper (see Harper et al., 1976; Harper, 1977, 1978).

Harper et al.(1976) also considered a few problems related with internal consistency of the data and physical consistency of them.

We continue further with expression (17) from Stubbe (1970), which considers the zonal flow of mass. To use the incoherent scatter data we will rewrite it as:

$$u_{z_1} = \frac{1}{\nu_{in}(z)} \int_{z_0}^{z} \nu_{in}(s)\, f\,(\phi)\, \frac{\pi}{21600}\, u_y(s)\, ds\ ,$$

where $f(\phi) = 2.16 \times 10^{-5}/\cos\phi\,(cm^{-1}sec)$ as described in Stubbe (1970), and the diurnal time dependence was omitted for simplicity. The values of u_y were taken from the values of u_x (meridional winds) by using the time shift of 3 hours between these velocities (see Harper et al., 1976). The amplitudes were considered the same (see Harper et al., 1976). Roughly one wavelength was considered along the vertical by integrating from 98 km to 125.7 km to compute u_{z_1} at 1300 hr (close to hour of maximum). This procedure reduces the problem of possible phase shifts along the vertical.

The computed value was 9.5 m/s whereas the measured value is 23 m/s. The result can be considered satisfactory because of the large number of approximations used and the large errors involved. However, considering expression (32), from tidal theory, one sees that a more reasonable approach is using twice the computed value, to account for zonal and meridional flow of mass. The result produced by expression (32) goes then in the right direction to closeness with measured values.

The physical consistency of the time variation of neutral atmosphere parameters, determined from incoherent scatter data, with that predicted by tidal theory was verified by Harper et al.(1976), Wand (1976) and Zamlutti (1983).

Summarizing, for equinoctial condition, we have the validity of the following simplifications:

a) hydrostatic equilibrium for unperturbed atmosphere;
b) a Bates type formula to described the heat balance;
c) the tidal theory to represent the regular diurnal perturbation produced by the day-night alternation.

During solstices the seasonal waves (see review by Evans et al., 1979) effects must be considered in the analysis. These effects can be estimated using expression (18). Assuming the same meridional velocity distribution of Johnson and Gottlieb (1970) we evaluated a vertical velocity of 0.1 m/s, due to seasonal meridional asymmetries for the altitude of 107.8 km, using the daily average meridional velocities obtained on December 4, 1972 and presented in Zamlutti (1973). The zero of meridional wind, which determines the height of the turbopause (see Volland and Mayr, 1972, b), lies at 100 km. Another zero, not predicted by linear tidal theory, is present at 110 km.

The daily average of the vertical wind published by Zamlutti (1973) presents the value of -3 m/s for the altitude of 107.8 km on December 4, 1973. This large value is inconsistent with theory and is probably biased by recovery effects (see Zamlutti, 1973; Zamlutti and Farley, 1975).

Another approach to seasonal waves is based on the same equations which have been established for tidal theory (Volland and Mayr, 1972, a and b). It considers the horizontal flow of mass produced by energy differences which is balanced with vertical flow of mass [see equation (32)]. This approach is essentially the one undertaken by Dickinson and Geisler (1968). In terms of incoherent scatter data expression (32) is written as:

$$\overline{u}_z = \frac{2}{r\,\overline{\nu_{in}}} \int_{z_o}^{z} \overline{u}_x \;\delta\overline{\nu_{in}}\; tg\,\psi\; dh \;,$$

where $\delta\overline{\nu_{in}}$ is the amplitude of the seasonal wave and the averages refer to winter values. Here only the meridional velocity was considered since the zonal winds were not measured. Using Zamlutti (1973) data we obtained a vertical velocity of 0.01 m/s, with the last expression, for the altitude of 107.8 km. This result is consistent with the thermodynamic equation outcome from the same set of data, which for the present conditions is produced by:

$$\overline{u}_z = - \left(\frac{7\pi}{2}\right) \left(\frac{H}{\overline{\nu_{in}\quad T}}\right) \left(\frac{\overline{\Delta\nu_{in}\quad T}}{\Delta t}\right),$$

where the Δ denotes forward finite difference and H is the vertical scale height. The agreement of these last results shows the internal consistency of the set of data employed. Unfortunately such small vertical velocities can not be accurately measured by Doppler type techniques.

The last result concerning internal consistency of data shows that the tidal theory can be used to explain the seasonal variations of neutral atmosphere parameters.

The validity of tidal theory to explain, in a satisfactory way, the heat balance is well documented in the literature (see for instances Forbes, 1982a, b; Forbes and Champion, 1982). This is easy to understand since the theory agrees well with data as far as the dynamics of the upper atmosphere is concerned, therefore, the terms depending on the velocity are well-determined in the energy equation and the accuracy will be restricted only by the production and loss terms.

12 - DISCUSSION

The use of low latitude and altitude data, in this work, was deliberate to avoid the strong effect of interhemispheric transport (Waldteufel, 1970; Young et al.,1980; Chandler et al.,1983). With this restriction, the analysis of upper atmosphere models is easier and, as discussed before, reduces to the study of the effects of the solar EUV energy on the neutral species.

An analysis of higher altitudes and latitudes must necessarily include flow situations which are not properly described by hydrostatic equilibrium equations. Although the neutrals stay in a near equilibrium they interact with the particles flowing through them. Therefore interhemispheric transport, as well as particles bombardment,must be considered in thermospheric modeling.

For the low latitudes the ion neutral interaction must be taken into account during solstices above 200 km (Chandler et al., 1983). This is necessary because of the presence of a low speed proton flow through the ionosphere-atmosphere system. For the middle-latitudes the problem increases since the proton flow penetrates down to 100 km (Waldteufel, 1970). This affects the lower thermosphere temperature due to the transported energy, the momentum through mechanical collisions and the mass because of the injection of atomic ions at E region altitudes.

So far interhemispheric transport effects were not properly considered on the neutral atmosphere. The models discussed here keep being used to determine the consequences of this transport on the ionized species. However, problems do appear to match the theoretical results with experimental data (see Chandler, 1982).

A first attempt to consider interhemispheric transport is found in Dickinson et al.(1981). Their approach uses the heating efficiencies of Torr et al. (1980). This last work, however, considers only the transient to establish chemical equilibrium. Transport of ionized particles (Young et al., 1980) must still be accounted for.

13 - CONCLUSIONS

Thermospheric modelling was considered focusing the aspects of the validity of the currently used approximations. Hydrostatic equilibrium was observed to hold during unperturbed situations (equinoctial conditions). Regular wavelike perturbations (tides, planetary waves) could be explained in terms of the classical tidal theory. Solstice conditions were examined only as referred to planetary oscillations. Interhemispheric transport, which constitutes an important perturbation during solstices, was not considered in this work because it was not implemented in thermospheric models so far.

ACKNOWLEDGEMENTS

The author is grateful to miss M.L. Ribeiro for her help in library work. Support for this work came from Instituto de Pesquisas Espaciais - INPE, Brazil.

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FIGURE CAPTION

Figure 1 - Temperature variation in the range 110 - 130 km. Empty circles refer to Jachia(1970) model, with exosphere temperature of 1000^{o}k. The dots come from Bates (1959) analytical expression with exospheric temperature of 1000^{o}k and lower limit temperature identical to that of experimental data. The triangles also came from Bates formula with $T_{\infty} = 900^{o}$k. The x are measured values from Nov. 7, 1972 (Zamlutti, 1973). Where triangles and x coincide the triangles were omitted.

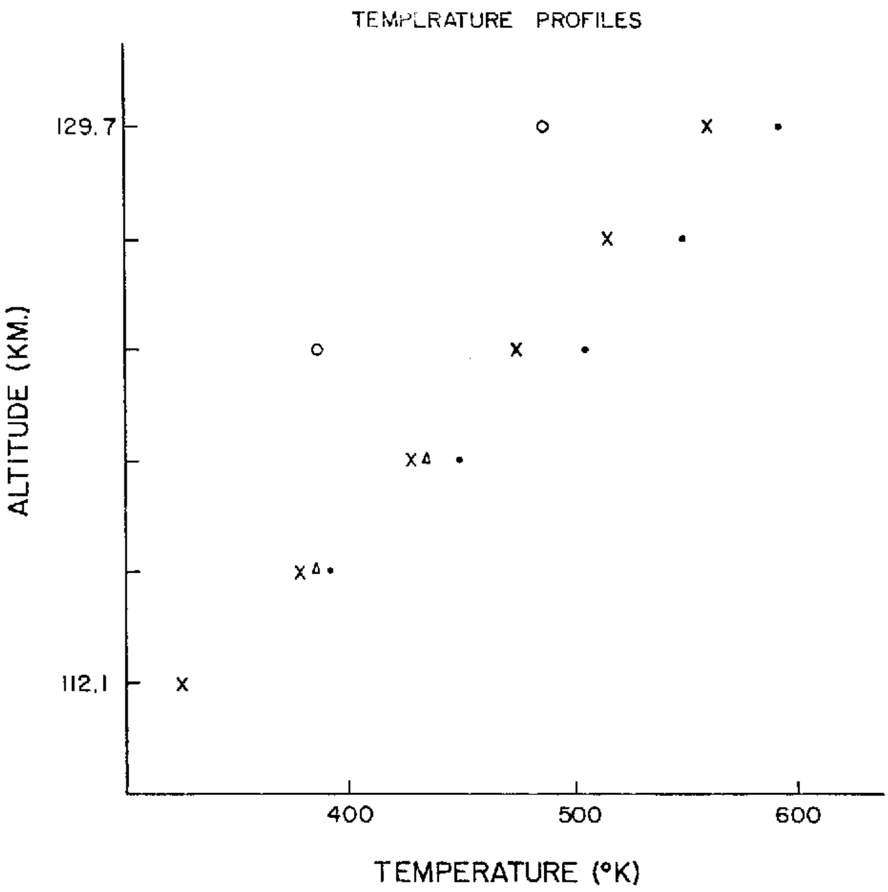
TEMPLRATURE PROFILES
129.7
112.1
ALTITUDE (KM.)
400
500
600
TEMPERATURE (°K)

Fig. 1