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THEORETICAL AND EXPERIMENTAL INVESTIGATIONS OF UPPER ATMOSPHERE DYNAMICS

By

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Theoretical and Experimental Investigations of Upper Atmosphere Dynamics

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CONTENTS

Abstract

Theoretical and experimental investigations of upper atmosphere dynamics have been pursued under this grant for almost two decades. This final report presents a brief overview of the significant contributions made to the understanding of the dynamics of the earth's upper atmosphere, including the addition of winds and diffusion to the semi-empirical Global Reference Atmospheric Model developed for the design phase of the Space Shuttle, reviews of turbulence in the lower thermosphere, the dynamics of the equatorial mesopause, and stratospheric warming effects on mesopause level dynamics, and the relevance of these studies to the proposed Middle Atmosphere Program (1982-85). The report ends with a chronological bibliography, including abstracts, of all papers published under this grant.

CHAPTER I

Overview

In the early sixties, the foremost technique for the measurement of wind profiles in the lower thermosphere was the rocket released chemical vapor trail. The early work performed under this grant consisted of the development of techniques for the accurate measurement of trail position with time (based on triangulation against the stellar background from two or more ground based camera sites), and lead to the determination of wind and wind shear profiles, gravity wave and turbulence spectra, and representative turbopause heights. In the mid and late sixties, better models explaining the turbulence observed evolved, supported by concurrent interpretation of wind shears in the 80 to 100 kilometer height range, measured by the multi-station meteor wind radar technique. The cascade of energy from the diurnal tide to gravity waves to turbulence was proposed at this time, and subsequently verified by others.

Work on turbulence in the lower thermosphere continued through the early seventies, culminating in the review presented here as Chapter III.

Also during the early seventies, the space shuttle reentry model atmoshpere, and its extension to a wind and diffusion model, were developed. This Global Reference Atmospheric Model (GRAM) is available on magnetic tape, and has been widely used both within and outside NASA. The model describes pressure, density, temperature and winds as monthly means and variances (details in Chapter II).

Funds from this grant have been used in support of the analysis and interpretation of winds from the Georgia Tech Radio Meteor Wind Facility.

This is the only continuously operating meteor wind radar in the world which produces height/time profiles of winds in the 80 to 100 km height range (a sample of the prevailing winds, zonal and meridional, for the period August '74 - August '75 is presented as Figure 6 in Chapter VII, paper b).

Mesospheric and lower thermospheric dynamics have been investigated (Chapter IV), and more recently, in cooperation with the French National Center for Telecommunications Studies (CNET), the dynamics of the equatorial mesopause (Chapter V).

Of particular interest to meteorologists and aeronomers in recent years has been the delineation of the coupling between various height ranges within the atmosphere. One obvious source of this coupling is to be found in random internal atmospheric gravity waves. Their contribution to vertical mixing in the stratosphere and mesosphere has been investigated, and is detailed in Chapter VI.

More recently, the effects of winter polar stratospheric warmings on the circulation at midlatitude mesopause levels has been studied, and preliminary results of this investigation are presented in Chapter VII.

This report would be incomplete without a reference to MAP - the Middle Atmosphere Program, scheduled for 1982-85, which is discussed in Chapter VIII.

Finally, and probably most importantly, a bibliography of all publications produced under this grant, together with abstracts, is appended.

CHAPTER II

The Global Reference Atmospheric Model

Although the Global Reference Atmospheric Model (GRAM) was developed as a thermodynamic variable model in support of the design phase of the Space Shuttle by a grant from NASA Huntsville, part of its subsequent evolution into a wind and diffusion model was supported under this grant.

GRAM is a computer program, available on magnetic tape, which combines the 4-D model of Spiegler and Fowler (1972) (0-25 km), a development of the Groves (1971) model (for the region between 30 and 90 km), and the Jacchia (1970) model (115-700 km). Smooth transitions between these models are accomplished with a fairing technique (90-115 km) and an interpolation scheme (25-30 km).

In addition to monthly mean values of pressure, density, temperature and winds, two types of perturbation are evaluated: quasi-biennial (QBO) and random. For further details of this model, see "A Global Reference Atmospheric Model for Surface to Orbital Altitudes" by C. G. Justus, R. G. Roper, Arthur Woodrum and O. E. Smith, Journal of Applied Meteorology, 15, 3-9, 1976.

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Groves, G. N. "Atmospheric Structure and its Variations in the Region from 25 to 120 km," AD-737-794, AFCRL-71-0410, Environmental Research Paper 368, July 1971.

Jacchia, L.G., "New Static Models of the Thermosphere and Exosphere with Empirical Temperature Profiles, Spec. Rept. 313, Smithsonian Astrosphysical Observatory, May 1970.

Spiegler, D.B. and M.G. Fowler, "Four Dimensional World Wide Atmospheric Model - Surface to 25 km Altitude," NASA Contract Report CR-2082, July, 1972.

CHAPTER III

Turbulence in the Lower Thermosphere

(Chapter 7 of "The Upper Atmosphere and Magnetosphere," published by the Geophysics Research Board of the National Academy of Sciences, Washington, D.C., 1977).

Turbulence in the Lower Thermosphere

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7.1 PROLOGUE

A static, or motionless, atmosphere can be adequately described on a macroscopic scale by three parameters: pressure, temperature, and composition. If the atmosphere were macroscopically motionless for a sufficiently long time (its molecules in purely random motion), one would find a decrease in mean molecular weight with altitude, the lighter species having separated from the heavier species by the process of molecular diffusion. Such a process is not observed in the earth's atmosphere until one reaches an altitude of from 100 to 110 km; up to this altitude the mean molecular weight remains approximately constant. This comes about because the earth's atmosphere is not motionless on a macroscopic scale but is continually mixed by atmospheric turbulence and large-scale circulation. The beginning of diffusive separation at 100–110 km does not mean that there are no winds to cause turbulence and mixing above these altitudes but rather that at such low densities molecular diffusion rates are much greater than the wind-induced mixing rates. This chapter concentrates on the region known as the lower thermosphere, between the mesopause at approximately 80 km, where the atmosphere is always mixed, and the altitude of 130 km, where diffusive separation is always observed.

Our knowledge of mixing processes in the upper atmosphere has been accumulated mostly in the years since World War II. While the lower thermosphere has been probed for decades using radio techniques, detailed knowledge of its structure has only come with the use of rocket soundings.

Before considering the techniques of measurement of mixing processes in the upper atmosphere and the interpretation of those measurements, a definition of "mixing process" is appropriate. On a global scale, the atmosphere can be said to be mixed by the transport of constituents from one location to another by large-scale wind systems. This mixing is important. Here, however, we will concern ourselves with more localized mixing, usually brought about in the free atmosphere by shears in the wind system and called atmospheric turbulence. A turbulent atmosphere is mixed at rates often hundreds or more times faster than its molecules can diffuse by means of their thermal motions, and thus turbulence can be very effective in maintaining homogeneity in an atmosphere composed of several molecular species; hence the term "homosphere" is often applied to describe the homogeneous atmosphere, with the heterosphere above. The level at which turbulent mixing ceases to be effective is sometimes referred to as the homopause, but here we will use the term turbopause.

In addition to mixing the atmosphere, locally intensive turbulence also causes local heating. Turbulence is dissipative, extracting energy from the total flow and transferring it by a cascading process to scales so small that the random motion of the molecules, which determines the temperature, is increased. Thus turbulence in the lower thermosphere is important because its intensity affects both the relative concentrations of constituents of the thermosphere and the heat budget in the lower thermosphere. The source of the turbulent energy resides ultimately in the lower atmosphere, and this energy is transferred to the thermosphere by the upward propagation of internal atmospheric gravity waves and tidal winds (see Chapters 8 and 9).

7.2 INTRODUCTION

In the parlance of ionospheric physicists, the lower thermosphere is known as the E region. The E region has been probed from the ground almost since the inception of radio. In particular, irregularities in the structure of the ionization that produces reflections of radio waves at this altitude have been observed for decades. Because these irregularities could be observed only through reflection from the continually changing ambient ionization imbedded in the neutral atmosphere, their interpretation in terms of the turbulent structure of the neutral atmosphere was highly speculative. However, in the early 1950's, a technique was devised that enabled the characteristics of the neutral atmosphere to be determined by direct observation of relatively well understood radio reflectors—the ionized trails formed by meteorites entering the earth's atmosphere.

Measurements of the radio-frequency Doppler shifts produced as meteor trails were blown along by the neutral winds between 80 and 100 km (the altitude range over which most meteorites burn up in the earth's atmosphere) were, and still are, used to provide knowledge of the neutral wind and its variation in both height and time. At the same time, considerable progress was being made by scientists working in the analysis and interpretation of atmospheric turbulence in the troposphere. The interpretation was assisted in no small measure by the progress made in turbulence theory, in particular by Batchelor's (1953) *Theory of Homogeneous Turbulence*. After this theory was applied successfully to the explanation of the tropospheric scattering of very-high-frequency radio waves, Booker and Cohen (1956) attempted to explain the fading observed on long-duration meteor echoes in terms of turbulence in the neutral atmosphere at E-region altitudes. From their data, they deduced that energy was extracted from the large-scale wind motions at meteor altitudes and dissipated at a rate ϵ = 25 W/kg. While the underlying theory was sound, their paper was attacked on the basis of their interpretation of the echo-fading process.

In the late 1950's, the chemical-release rocket technique was perfected and used to introduce a visible tracer, initially sodium, into the atmosphere over the altitude range 80 to 200 km. Such a release, made at twilight while the trail was illuminated by sunlight and the ground was in darkness, could be photographed from several camera sites on the ground, and a time series of exposures recorded simultaneously could be used for triangulation of the release, thus determining its motion with time. Winds could be determined for as long as the trail remained visible, sometimes for as long as 15 min. Since the early 1960's, trimethyl aluminum (TMA), which reacts with atomic oxygen in the ambient atmosphere to produce products that not only scatter sunlight but also glow in the dark (making nighttime as well as twilight measurements possible), has been used in addition to sodium. For daylight releases in particular, lithium has been used.

One outstanding feature of these trails is the fact that below an altitude of some 105 km the release is "obviously turbulent," whereas above that altitude the trail expands smoothly, as it should under the action of molecular diffusion alone. Considerable controversy still exists over the interpretation of data from rocket-released vapor trails as evidence for turbulence of the ambient atmosphere. Some interpretations of sodium vapor trails, for example, have led to anomalous results that are thought to be related to the energetics of the thermite burn used to produce the sodium vapor. However, interpretation of the breakup of vapor trails below what has become known as the turbopause is not the only evidence for the existence of turbulence at these altitudes. Theoretical studies of the atmospheric heat budget in the high atmosphere by Johnson and Gottlieb (1970), for example, require vertical diffusivities below the turbopause that are much larger than molecular. Calculations of the diffusivities responsible for the measured constituents above the turbopause also require similar values of diffusivity below the turbopause; composition measurements made using rocketborne mass spectrometers show the level of diffusive separation to be considerably higher than would be the case in a nonturbulent atmosphere. Further, a completely different class of measurements, based on the shearing of radio meteor trails by the winds in the 80- to 100-km region rather than the fading of individual meteor echoes, yields values consistent with those deduced from vapor-trail observations.

7.3 THE INTERPRETATION OF MEASUREMENTS

Is it possible for rocket-wake effects and chemical energy released by contaminants (from the thermite canister) to influence the subsequent dispersion? Certainly. Such effects have been documented. For energetic releases such as rocket exhausts or large quantities of explosive, the "release phase" at thermospheric altitudes lasts only for some 10 sec or less. Figures 7.1 and 7.2 show an excellent example of a release phase anomaly in a TMA trail, as photographed by the Smithsonian Institution's Baker-Nunn camera at Woomera, Australia (31° S). Figure 7.1 shows a portion of the trail 2 sec after release in the 100- to 110-km height range; vortex shedding on the right-hand side of the trail is clearly visible. In Figure 7.2, a frame taken 4 sec later, the effects of this motion have disappeared. The subsequent breakup of the trail, with the production of the characteristic "obviously turbulent" appearance, did not occur until some 30 sec later, as illustrated in the isodensitrace montages of Figure 7.3. These unique Baker-Nunn photographs are an example of the rewards of international cooperation. The Smithsonian Institution for many years operated a worldwide network of Baker-Nunn cameras for the photographic determination of satellite positions. Relationships between the satellite station staff and the Australian and British rocket experimenters at Woomera were such that the Baker-Nunn camera would be used to photograph the rocket releases when such use did not interfere with the primary mission of the observatory.

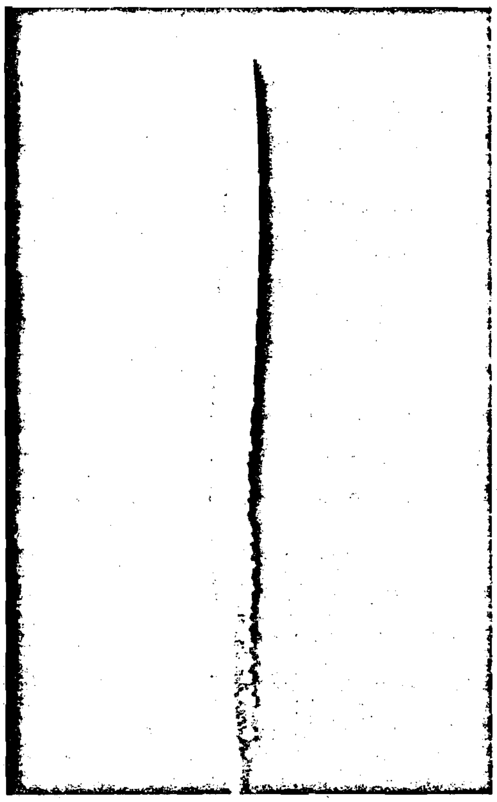

FIGURE 7.1 Portion of a trimethyl aluminum trail showing a release phase instability with vortex shedding on the right-hand edge 2 sec after release at 103 km.

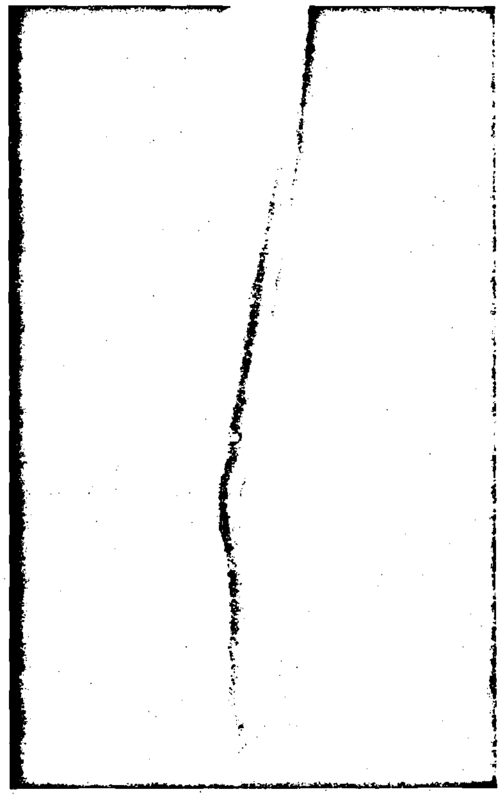

FIGURE 7.2 Same as for Figure 7.1, but 6 sec after release. Note that the release phase instability has been damped out.

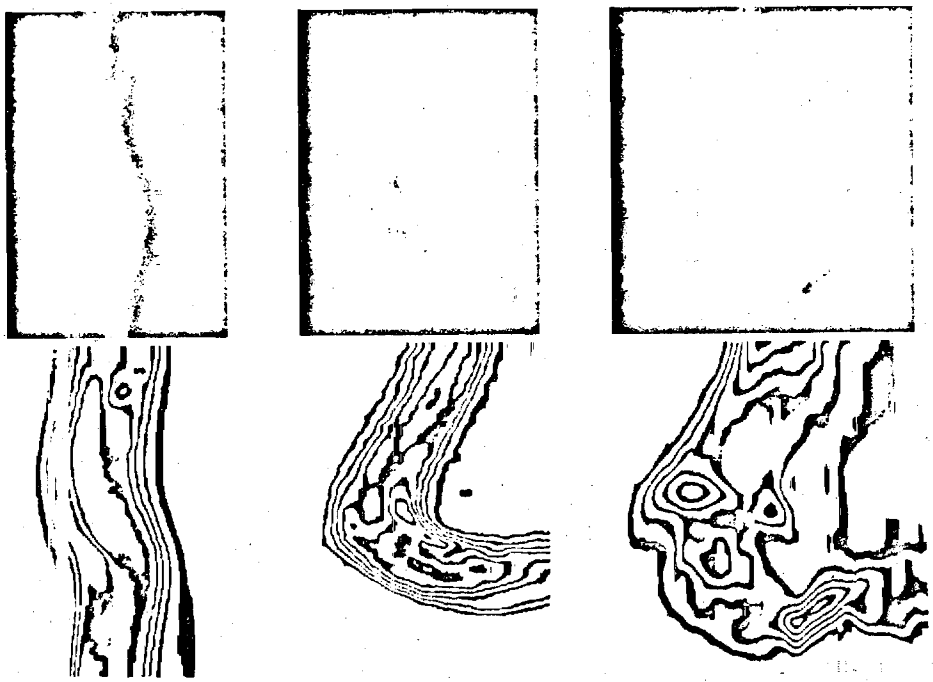

FIGURE 7.3 Time history of a portion of the trail of Figure 7.2 showing the initial laminar behavior and subsequent breakup induced by ambient atmospheric turbulence. Quantitative measurements of dispersion can be made from the isodensitrace contours. The kink is at an altitude of 102.7 km. Note that the trail is laminar at +30 sec and that the trail diameter at this time is greater than the scale of the eddies at +50 sec.

The abrupt cutoff in ambient turbulence (the turbopause), which almost always underlies a region of high wind shear, is illustrated in Figure 7.4. That the breakdown from "obviously laminar" to "obviously turbulent" represents a dramatic change, easily recognized by visual inspection of the film, is shown in Figure 7.5, in which the square of the effective radius of the trail is plotted against time after release at 105 km. The growth in the first few seconds after release is an order of magnitude faster than the subsequent growth up to the time of trail breakup, and this is regarded as representing the release phase, in which the energetics are definitely nonambient. The growth between 8 and 32 sec after release could be molecular diffusion of a cloud with an initial radius of 130 m, i.e., the release produced a cloud with, effectively, this radius at zero time. However, one can only say that the growth during this phase could be molecular—the 15 percent error in the determination of the effective radius from each film frame, and the fact that the trail cross section is only approximately Gaussian, precludes a measurement of the diffusion coefficient to better than a factor of 3. The spreading due to turbulence commences 33 sec after release and then proceeds to follow the dispersion relation predicted by the theory of homogeneous turbulence:

$$r_e^2 \propto t^3,$$

until, at 54 sec, the trail becomes too irregular for an estimate of radius based on a Gaussian distribution to be

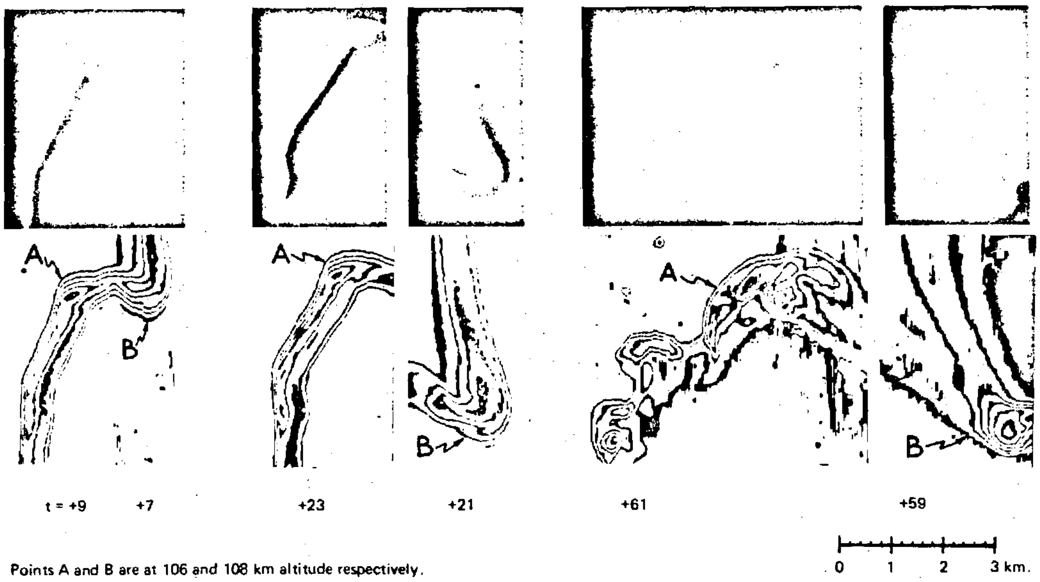

FIGURE 7.4 Montage of the behavior of the trail above and below the turbopause. Points A and B are at 106 and 108 km, respectively. Note the transition from laminar to turbulent at A between +23 and +61 sec. This has not occurred at B.

meaningful. However, Figure 7.5 does provide two parameters: the effective radius at transition and its time of occurrence after release, which should characterize the small-scale end of the turbulence spectrum. The concept is that until the trail spreads to a size equal to that of the smallest scale eddies, it is distorted but not spread by the eddies. Thus the scale size at transition is interpreted as indicating the size of the smallest eddies.

The most important parameter of any turbulence spectrum is ϵ, the rate of dissipation of turbulent energy. In order to calculate ϵ from any set of space–time correlations, it is necessary to use a model. The simplest is that of A. Kolmogoroff, as elaborated on by Batchelor. Kolmogoroff put forward the hypothesis that, in any turbulent flow field in which energy is extracted from the large-scale or mean motion and cascaded to smaller scales before eventual dissipation at scales where molecular viscosity becomes important, there exists a range of scales sufficiently removed from the large-scale anisotropic eddies, and yet not appreciably damped by molecular viscosity, that is both homogeneous and isotropic. The assumptions of homogeneity and isotropy inherent in this model are open to question in the case of lower thermospheric turbulence, but a considerable amount of work by several experimenters has shown these assumptions to be reasonable, at least for length scales less than 1 km.

Batchelor defines the basic length parameter of the viscous region (that is, the size of the small-scale eddies that are responsible for the conversion of the eddy kinetic energy to heat) as

$$\eta = (\nu^3/\epsilon)^{1/4},$$

where ν is the kinematic viscosity (a measure of molecular diffusivity), and η is the scale size in meters/radian (thus the characteristic wavelength of these turbulent eddies is

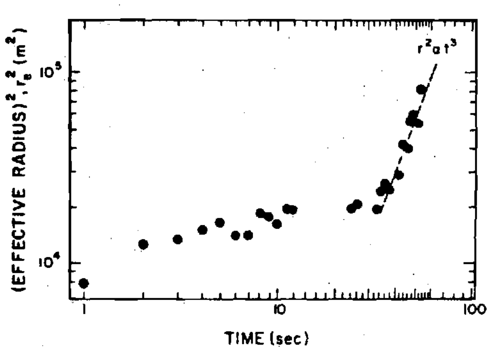

FIGURE 7.5 Variation of the square of the radius of the trail of Figure 7.1 with time after release at 105 km.

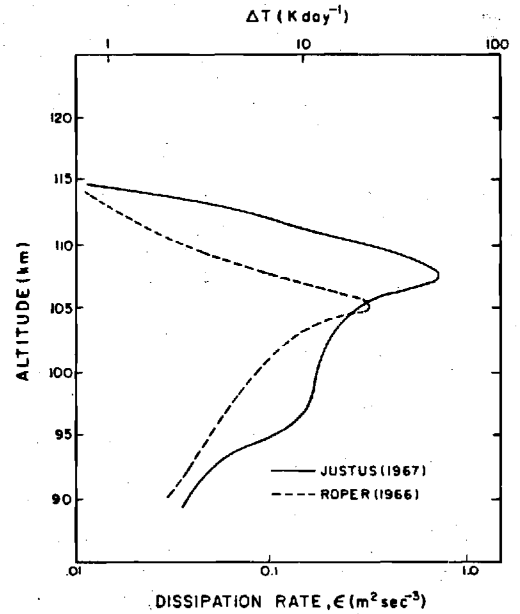

FIGURE 7.6 Profiles of the rate of dissipation of turbulent energy deduced from wind shears measured for 25 trails over Wallops Island (38° N) by Roper (1966b) and from the dispersion of a similar number of trails over Eglin Air Force Base (29° N) by Justus (1967). The upper scale gives a measure of the atmospheric heating resulting from turbulent dissipation.

$2\pi\eta$). Batchelor also defines a characteristic time constant t^* corresponding to this unit length scale such that

$$\eta = (\nu t^*)^{1/2}.$$

t^* is the lifetime of the smallest eddies, essentially the time required for their dissipation by molecular viscosity. Combining these two equations to eliminate η yields

$$t^* = (\nu/\epsilon)^{1/2}.$$

In terms of the length scale η,

$$\epsilon \propto \eta^{-4}.$$

In terms of the time scale t^*,

$$\epsilon \propto t^{*-2}.$$

Wind shears have been used to calculate the variation with height of the rate of dissipation of turbulent energy (Roper, 1966b). The kinematic viscosity ν may be determined from the viscosity and density published in the U.S. Standard Atmosphere Supplements for 1966. The function

$$\nu\ (\mathrm{m^2\ sec^{-1}}) = \exp\,[0.17\,(z - 80.0)],$$

with z in kilometers, fits the data at the tabulated altitudes. The ϵ profile determined in this way is shown in Figure 7.6, together with the profile inferred from measurements made by Justus (1967), calculated from the velocity fluctuations observed on 18 TMA trails. These values of ϵ and ν have been coupled to produce what can be regarded as average height-dependent characteristic length scales η_R (= $2\pi\eta$, in order to express the length scale in the more usual wavelength notation, meters/cycle). η_R and t^* are plotted, together with values calculated from Justus's ϵ profile, in Figures 7.7 and 7.8, respectively. Also plotted are the values determined from two Skylark-released TMA trails photographed by the Baker-Nunn camera at Woomera at dawn and dusk on May 31, 1968.

The similarity in form, and the order of agreement, between the predicted "average" trail time constants t_R^* and t_J^* (Roper's and Justus's values, respectively) and the measured values for each of the two releases is surprisingly good when one considers the approximations involved in (a) the model atmosphere viscosity, which is based on an average atmospheric model; (b) the variation, both diurnal and seasonal, in the turbulent dissipation rate, which has been averaged out in the construction of the t_R^* and t_J^* profiles; and (c) the fact that, while the

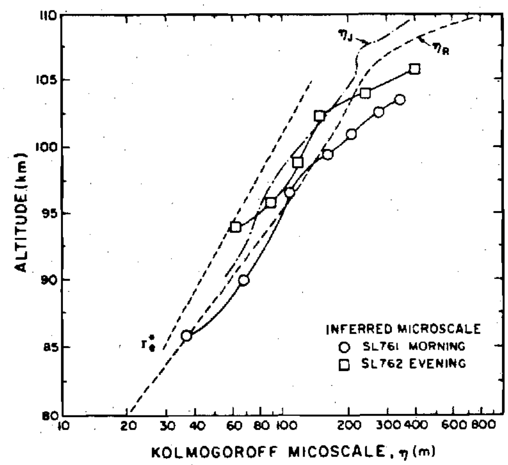

FIGURE 7.7 The variation with altitude of the Kolmogoroff microscales η_J and η_R inferred from the turbulent dissipation profiles of Figure 7.6. Also shown are the microscales inferred from the May 1968 trails. r_e^* is the Gaussian radius of the evening trail at the time of onset of trail breakup, t^*.

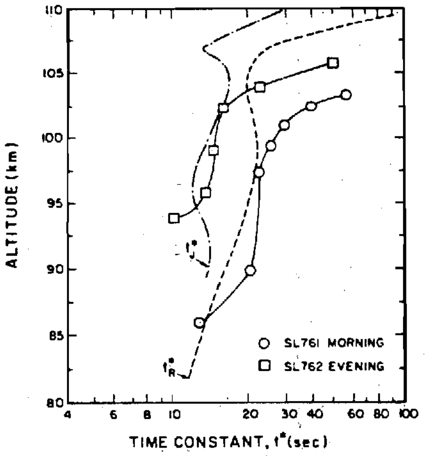

FIGURE 7.8 The variation with altitude of the time constant of the Kolmogoroff microscale for the ϵ profiles of Figure 7.6 and the May 1968 trails.

parameter t^* appears to be a measure of the time taken for trail breakup to occur, it is not clear why it should be.

The definition of η is a mathematical expedient characterizing the scale size for turbulence at which viscous dissipation becomes important. There is no obvious reason why either η or t^* should be physically measurable features of the motion. Nevertheless, the close similarity in the shapes of the t^* curves strongly indicates that the time delay in the onset of turbulence should be related to the time constant of the Kolmogoroff microscale. Furthermore, these observations suggest a reason why the turbopause, defined as the boundary between the regions that break up and those that remain laminar, should manifest itself so abruptly. Above 105 km, the time constant t^* for the onset of turbulence increases so rapidly with altitude that the trail is not, in general, observed for a sufficient length of time for visible breakup to occur.

Attempts have been made to explain the existence of the turbopause in terms of a critical value of some parameter, generally the Reynolds number R_e or the Richardson number R_i. These attempts have met with marginal success at best, partly because of the difficulty in defining the characteristic lengths that occur in these parameters and partly because it is not evident *a priori* what critical value the parameter should have at the turbopause. Without a detailed knowledge of the temperature gradient at the scale characteristic of the vertical mixing process (a few hundred meters or less), the Richardson number just cannot be specified. Johnson (1975) has considered the relative importance of buoyancy and dissipation in some detail, referring particularly to the work of J. D. Woods, who determined that there was hysteresis in the criterion—laminar flows become turbulent when $R_i \leq$ 0.25, while turbulent flows become laminar when $R_i \geq 1$. While the mean (undisturbed) atmospheric temperature profile is pertinent to the breakdown from laminar to turbulent flow, once turbulence is established its cessation will depend on the temperature gradient as modified by the turbulence. As yet, there is no technique available for measuring such temperature gradients in the lower thermosphere.

Another error that has often been made in attempting to explain turbulence in the lower thermosphere is the assumption that the turbopause corresponds to an altitude at which turbulence ceases abruptly. The results presented here, on the contrary, show that the turbopause is the altitude at which the time constant of the Kolmogoroff microscale of the turbulence increases very rapidly with altitude. This viewpoint resolves the paradox that regions above the turbopause, which were thought of as nonturbulent, have diffusion coefficients based on the measured laminar trail growth that are greater than molecular. We now see that turbulence does exist above the turbopause but that its efficacy in transport relative to molecular diffusion decreases with altitude. At an altitude of 130 km, the contribution of turbulence to diffusivity is insignificant, even though its absolute value may be as large as it is at the turbopause.

The rate of dissipation of turbulent energy may also be calculated from η, the length scale at which eddy diffusion becomes effective. However, since this length depends critically on the shape of the cloud (the assumed Gaussian variation across the cloud is rarely realized in practice below the turbopause) and because $\epsilon \propto \eta^{-4}$, Rees *et al.* (1972) chose to use the relatively precisely determined t^* values to calculate the variation of ϵ with height. This is shown for a pair of dawn and dusk releases above Woomera (30° S) on May 31, 1968, in Figure 7.9. Note that

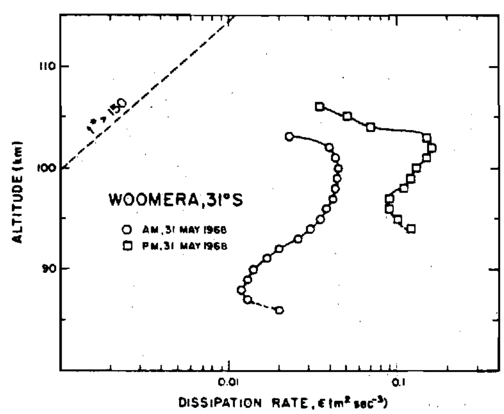

FIGURE 7.9 Estimates of the rate of dissipation of turbulent energy ϵ from the time of onset of turbulence for the release of May 1968.

the turbopause is higher in the evening than in the morning and that the higher turbopause is associated with a greater overall turbulent intensity. This substantiates the suggestion of a variation in midlatitude turbopause height made by Elford and Roper (1967), which was based on seasonal variations in turbulent intensity at 93 km as determined from the wind shear measured simultaneously on individual radio meteor trails.

The results from two further TMA releases made above Woomera at dawn on October 16 and dusk on October 17, 1969, are presented in Figure 7.10. For this pair of trails, the turbulent intensity is higher in the morning than in the evening—opposite to the May 1968 releases. This diurnal variation with season is the same as that measured for the large-scale turbulent velocity component from radio meteor winds at Adelaide (35° S). These October releases are of particular interest in that they show alternating laminar and turbulent regions similar to those previously reported by Blamont and Barat (1967). The various layers observed in these releases do not seem to be quite so simply related to the wind profile as those of Blamont and Barat. However, the regions of prolonged laminar behavior all seem to be located at altitudes where the wind shear is high.

In an attempt to explain why turbulence in the lower thermosphere should be stratified at times, and why in fact an ostensibly highly stable region of the earth's atmosphere should be turbulent at all, Lloyd *et al.* (1972) proposed a model in which random internal gravity waves produce turbulence accompanied by a considerable modification of the temperature profile. The creation of temperature inversions by turbulence is commonplace in the troposphere. It is proposed that a similar effect occurs in

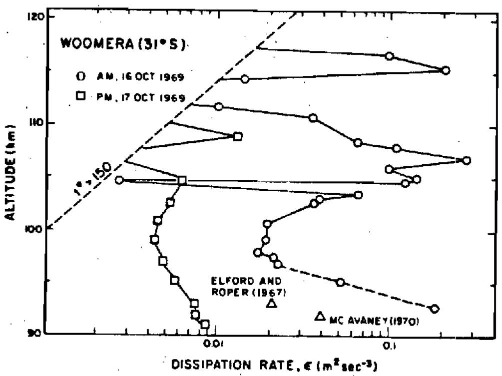

FIGURE 7.10 Same as for Figure 7.9 but for the releases of October 1969. Note the presence of turbulent "sheets" alternating with laminar layers. Diurnally averaged dissipation rates for the month of October 1961 (Elford and Roper, 1967) and October 1969 (McAvaney, 1970) as measured by the radio-meteor technique at Adelaide (35° S) are shown for comparison.

the more stable lower thermosphere, with part of the vertical component of the gravity-wave velocity as the source of the turbulent energy. If random, internal gravity waves propagate at a significant angle to the vertical, then their vertical velocity component will be able to contribute to the destabilization of the stably stratified lower thermosphere (Hodges, 1967). Similar destabilization mechanisms have been discussed for the oceans by Phillips (1971) and for the lower stratosphere by Roach (1970). Radio meteor studies have already established that gravity waves are the source of the turbulent energy in the lower thermosphere. This has been confirmed quite independently in radio-meteor studies by Spizzichino (1972) and by Blamont and Barat (1967) through observations of chemical releases.

The lower thermosphere is stabilized against vertical motion by the mean temperature gradient; above the mesopause near 80 km the mean temperature increases with height. By equating the work done in the vertical displacement of a parcel of gas to the fraction of the vertical component of the gravity-wave spectrum that is responsible for the measured turbulence spectrum, Lloyd *et al.* deduced the modification of the mean temperature profile that would be produced by the turbulence measured on the October 16 trail in the height range 104 to 110 km. An isodensitrace montage of this portion of the trail is shown in Figure 7.11. The choice of the turbulent layers within which the analysis can be applied is somewhat subjective, being based on regions where growth is "obviously" different from that above and below.

The solid lines of Figure 7.12 show the results of the application of the Lloyd *et al.* model to each of the layers delineated in Figure 7.11. The dashed lines are necessary for profile continuity and must represent laminar sheets. Because of the discontinuous nature of the determination of the modified profile and the forced fitting of the mid-point temperatures, the magnitudes of the positive and negative gradients are open to question. However, it is interesting to note that the existence of similar gradients in the lower stratosphere, a region of similar mean-temperature gradient, is well documented from aircraft observations, as can be seen from the project HICAT determination shown in Figure 7.13 from Mitchell and Prophet (1969). Unfortunately, the flight of an aircraft at constant altitude cannot reveal a height profile of turbulence, but at least in the encounter with CAT (clear air turbulence) at 21 km, the temperature profile has been modified in a manner consistent with the present model.

Several deficiencies exist in the model, since the finite time constants of the processes involved (the period of the destabilizing gravity wave, for example) have not been considered. The basic energy-budget equation can be made more general by the inclusion of terms describing energy sinks (e.g., heat conduction). One promising model being developed uses the reversible heating associated with propagating gravity waves (Hines, 1965) as the initial destabilizing energy. Even with this criticism, the above semiempirical approach allows deduction of

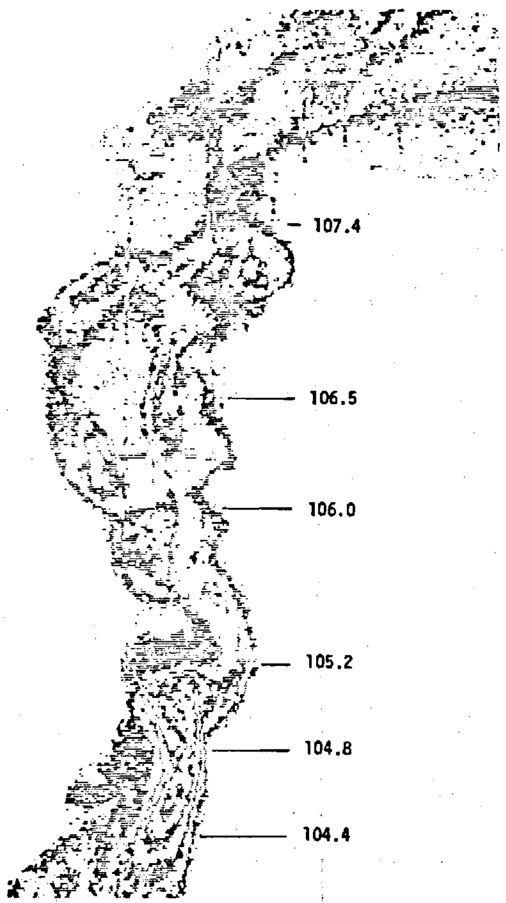

FIGURE 7.11 Isodensitrace of portion of the morning trail of October 16, 1969, 50 sec after release, showing the division into laminar and turbulent layers. Altitudes are shown in kilometers.

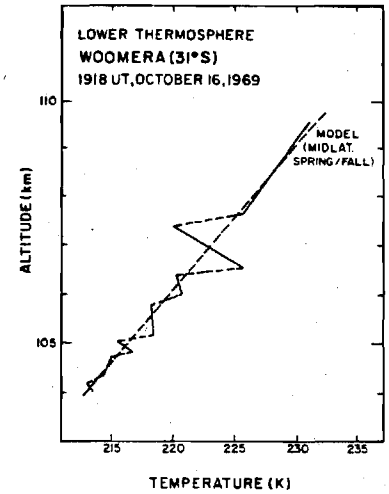

FIGURE 7.12 Theoretical morning temperature structure deduced from the profile of Figure 7.11.

many reasonable properties of the atmosphere. In particular, the model counters the objections raised to the existence of turbulence in what ostensibly is a highly stable region; the presence of turbulence itself tends to destabilization by modification of the temperature profile.

Up to this point, major emphasis has been placed on the fundamental parameter ϵ, the rate of dissipation of turbulent energy. There is an equally important although not so easily defined parameter, K_z, the coefficient of turbulent eddy diffusion in the vertical, which is the transport parameter incorporated in all one-dimensional models of the chemistry and constituents of the lower thermosphere. For some time, there was considerable discrepancy between the diffusion coefficients calculated from the growth of rocket-released tracers and those inferred from measurements of diffusive separation and atmospheric heat-budget calculations. With the discovery that the turbulence responsible for the enhanced diffusivity of chemical releases in the lower thermosphere was highly

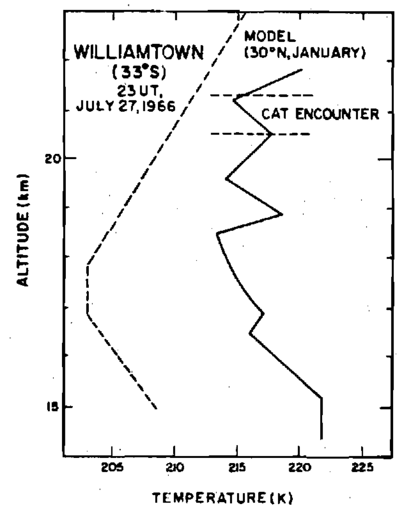

FIGURE 7.13 An example of temperature structure in the lower stratosphere (Mitchell and Prophet, 1969).

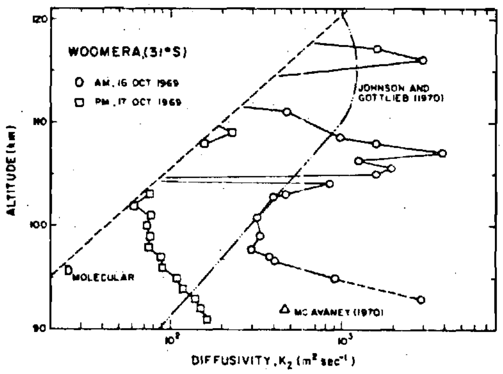

FIGURE 7.14 The turbulent vertical eddy-diffusion coefficient profiles deduced from the structure on the October 1969 releases. Δ indicates the coefficient deduced from simultaneous radio-meteor observations at Adelaide (35° S), approximately 450 km southeast of Woomera.

anisotropic with horizontal scales ten times those of the vertical, this discrepancy was readily explained. Since the time constant t^* for the observed onset of turbulence as used by Rees *et al.* (1972) is characteristic of the small-scale, isotropic turbulence spectrum, Lloyd *et al.* (1972) used the ϵ values thus determined to estimate the vertical diffusion coefficient. By an extension of their temperature profile modification model, they equated the vertical destabilizing influence of the turbulence to the stabilizing influence of the mean-temperature profile to obtain a vertical diffusion coefficient

$$K_z = \frac{\beta\epsilon}{\dfrac{g}{T_0}\left(\dfrac{dT_0}{dz} + \Gamma\right)},$$

where g is the acceleration due to gravity, T_0 is the temperature at altitude z, dT_0/dz is the undisturbed mean temperature gradient, and Γ is the adiabatic temperature lapse rate, 9 K/km. The above relationship is based on the fact that as long as the vertical temperature gradient remains greater than the adiabatic, there is an inhibition of vertical transport by vertical motion. It is interesting to compare this relationship with that determined independently by Lilly *et al.* (1973) for the lower stratosphere. The constant β above was determined from turbulence theory to be 10. Lilly arrived at the value of 1/3! This discrepancy is not so serious as it appears at first sight, since Lilly's temperatures and temperature gradients were the measured values, leading to a considerably smaller denominator than that produced by the use of the undisturbed mean temperature and gradient values by Lloyd *et al.* The variation of K_z with height for the October 1969 releases is shown in Figure 7.14. The theoretically deduced upper limit for eddy diffusion, based on the heat flux model of Johnson and Gottlieb (1970), is shown for comparison. Note that the maximum value of eddy diffusivity estimated by Johnson and Gottlieb is the integrated global maximum averaged over all seasons and can be exceeded by a particular measurement if there is any temporal, latitudinal, diurnal, or seasonal variation. Note also that all the turbulent intensity and diffusivity profiles presented here reflect the apparent sharp cutoff in turbulence as observed on chemical trails. With the time constant in onset of turbulence increasing so rapidly at and above the turbopause, most trails are not observable long enough for breakup to occur. However, observed diffusivities between the turbopause and approximately 130 km are usually greater than molecular, in agreement with the Johnson and Gottlieb calculations.

The only two sets of data so far, produced that are amenable to analysis in terms of month-by-month variation of ϵ or K_z were determined from meteor trail shear measurements made at Adelaide (35° S) by Roper (1966a) and McAvaney (1970). The variation of the monthly mean K_z at 93 km is shown in Figure 7.15. Also plotted in this figure are the results deduced from a far more rigorous treatment of the 1961 ϵ data by Zimmerman; some modifications of absolute values is evident, but the overall variation, with equinoctial maxima, remains. There is a real difference between the 1961 and the 1969 data that is

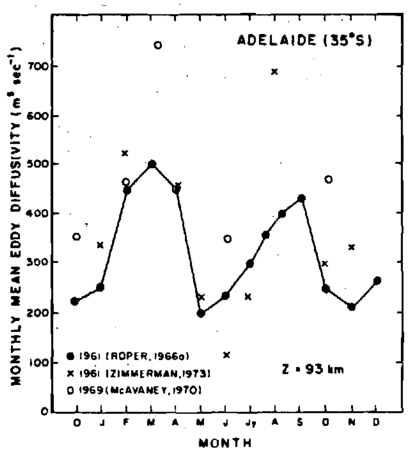

FIGURE 7.15 The southern hemisphere midlatitude variation of vertical eddy diffusivity deduced from two years of radio-meteor wind-shear observations. Zimmerman's values result from a more rigorous analysis of the same 1961 data.

not one of location, interpretation, or technique. The higher values measured in 1969 may be a consequence of the higher solar activity that year.

7.4 SOURCES OF TURBULENCE ENERGY

The fact that random internal gravity waves are the source of the turbulence energy in the lower thermosphere has already been mentioned. While this subject is covered in more detail in Chapter 8, one further correlation is pertinent to this discussion. In looking for possible sources of turbulence in the meteor region, comparisons were made with the magnitudes of the prevailing and tidal winds and shears as measured simultaneously by the radio-meteor method. While no apparent relationship existed between the turbulence parameters and the prevailing and semidiurnal winds, the seasonal variation of the amplitude of the diurnal oscillation was highly correlated with the turbulent intensity. The monthly means of the amplitude of the diurnal tide for several years of observation are shown in Figure 7.16. Note that equinoctial maxima in the diurnal tidal wind amplitude are not observed in all years and, therefore, that the measured variation in turbulent intensity may not be characteristic of all years.

The global-scale diurnal tidal wind does not produce turbulence directly but by a cascade process in which the tidal wind becomes unstable and generates *in situ* a spectrum of random internal gravity waves. This instability in the diurnal tidal wind may come about either because its amplitude becomes large or because of nonlinear interactions with gravity waves generated below and propagating upward through the lower thermosphere. Such gravity waves propagating from below can themselves be a direct source of turbulent energy. Thus the dominant feature of turbulence in the lower thermosphere, even if it is present at all times, will be the intermittency of its intensity. Little is known of the role played by large-scale motions such as planetary waves in the stability of this region of the atmosphere. In fact, since practically all measurements of turbulent intensity have been made at middle latitudes, the properties of the global turbopause and the influence of the turbopause on, for example, the hemispherical asymmetries in minor constituents measured globally at satellite altitudes can hardly be estimated at this juncture. Our knowledge of the temperature structure of the turbopause region at scales less than 1 km, which is crucial to the understanding of turbulence and diffusivity, is woefully inadequate. Present measurement techniques are quite incapable of producing such detail at these altitudes.

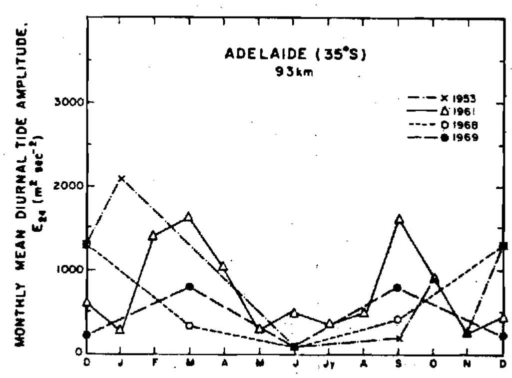

FIGURE 7.16 The seasonal variation of the southern hemisphere midlatitude diurnal tidal wind energy per unit mass determined from radio-meteor data. Equinoctial maxima are not a feature of all years.

In addition to the modifications to the temperature profile already discussed, the dissipation of turbulent energy produces heating of the ambient atmosphere. For the mixed atmosphere of mean molecular weight 29, a rate of turbulent dissipation ϵ of 1 W/kg produces heating at a rate of 85 K per day. The top scale of Figure 7.6 gives the heating rates appropriate to the inferred turbulent dissipation rates. In the light of the considerable variability indicated by the measurements presented here, even more emphasis must be given to the rate of dynamical heating of the lower thermosphere, as proposed by Hines (1965). The dissipation of wind energy at these altitudes may, at times, give rise to local heating rates in excess of those due to the solar input, which has usually been considered to be the major source of heating in this region. The relative importance of turbulence in mixing and dynamical heating has been summarized by Johnson (1975).

7.5 CONCLUSIONS

While the turbulence in the lower thermosphere is isotropic to scales of a few hundred meters, the transition from turbulence to the nondissipative scales of the gravity-wave spectrum is gradual, and therefore horizontal diffusivity is greater than vertical diffusivity. Based on diffusivity criteria alone, one would say that the turbulence is anisotropic, with a vertical scale of the order of 1 km and a horizontal scale of a few kilometers. The turbulent intensity is intermittent in both space and time, with large diurnal and significant seasonal and possible solar-cycle variations. For this reason alone, it is essential that simultaneously measured *in situ* values of atmospheric parameters be used in any attempt at meaningful comparisons. Since practically all measurements of turbopause altitude and turbulent intensity have been made at middle latitudes, and those low and high latitude measurements that have been made have not been coordinated with simultaneous observations elsewhere, variations with latitude are largely unknown. However, it has been established by rocket-grenade and falling-sphere measurements that the winter polar mesopause is warmer than that at the summer pole and that in the absence of solar input a significant atmospheric dynamical heating source is responsible. As has been shown, the turbulent dissipation of wind energy in the 90- to 125-km height range is significant and should therefore be included in any model of the thermospheric heat budget.

While knowledge of the vertical diffusivity in the lower thermosphere is vital to the understanding and modeling of thermospheric constituents, the role played by the global turbopause in the dynamics of the thermosphere has yet to be determined. High-resolution photography, with a frame rate of at least one every 2 sec, is able to produce data on turbulent intensities and vertical diffusivities from rocket vapor trails, but such measurements are highly localized in time and space. A recent development, the use of a high-flying aircraft as a camera platform, overcomes two ground-based camera problems: atmospheric haze and clouds are avoided, and photographs can be taken in the daytime. Daytime releases of lithium can also be observed from the ground using narrow band filters and electronic scanning systems (recording on video tape), but interpretation of the dispersion of the highly energetic release in terms of turbulence parameters is difficult. Vertical diffusivities and turbopause altitudes can be inferred from rocketborne mass-spectrometer measurements, regardless of hour of day or cloud cover but with location limitations similar to the vapor-trail method (both require reasonable range facilities). Multistation meteor wind radars can provide a continuous measurement of the turbulent intensity just below the turbopause, but only a few stations, all in midlatitudes, are currently capable of this type of operation, and none operates continuously. Incoherent-scatter radars are able to measure the temperature structure in the neighborhood of the turbopause with about 2-km height resolution, but these measurements are subject to constraints similar to, but even more severe than, those of the radio-meteor technique. A suggestion by Bencze (1970) that an ionosonde can be used to measure turbopause altitude warrants further investigation, since a global network of ionosonde stations is already in existence.

A proper understanding of the nature of the turbulence in the lower thermosphere requires a knowledge of the temperature profile with better than 1-km (preferably 100-m) resolution—a resolution that is not technically feasible at this time. Nevertheless, global variations in turbopause altitude and intensity can be determined using well-established techniques but only with international cooperation. Simultaneous global observations could be coordinated through the Middle Atmosphere Program (MAP) of the Special Committee on Solar Terrestrial Physics (SCOSTEP) and the International Mesospheric and Ionospheric Structure Parameter Interaction program (MISPI) proposed by the Soviet Union. Particular emphasis is being placed by these programs on the encouragement of the Arctic, equatorial, southern hemisphere, and Antarctic observations so badly needed for global coverage.

ACKNOWLEDGMENT

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CHAPTER IV

Mesospheric and Lower Thermospheric Dynamics

by

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INTRODUCTION

The region between 60 and 85 km altitude (the mesosphere) has often been referred to as the "ignorosphere," in that it is above the heights sounded by conventional meteorological rockets, and below the E region, which has been monitored for decades using radio techniques.

The advent of the rocket released vapor trail in the late 1950's did little to increase our knowledge of the mesosphere, in that these sodium vapor and trimethyl aluminum releases were concentrated in the 85 to 240 km altitude range, with the exception of a few titanium tetrachloride releases in the mesosphere.

Since the early 1950's, a considerable amount of data on neutral atmosphere motions in the height range 80 to 100 km has come from the meteor radar technique. However, with the notable exception of those stations operated by the Hydrometeorological Service of the U.S.S.R. (which, unfortunately do not routinely measure height, but consider all echoes as coming from a mean echo height near 95 km), the continuous operation needed to infer the synoptic meteorology of the region has been rare. The only attempt at such operation producing height/time profiles of winds between 80 and 100 km has been carred out at Atlanta (34°N, 84°W), and winds from August 1974 through December 1977 have been published by Roper (1978). While these results are valuable, being from a single station they provide only a crude insight into the synoptic meteorology of the meteor region.

PREVAILING WINDS

While considerable variability exists in the meteor winds measured by various groups, global circulation systems with seasonal variability, reproducing year after year, have been found. Analysis of several years northern hemisphere radio meteor wind data by Minina, et al., 1977, and Illichev and Portnyagin, 1977, shows that the pressure field at meteor heights in winter is similar to the baric profile of the underlying atmosphere. During the summer, however, the lower thermosphere differs considerably from the baric profile of the mesosphere, stratosphere, and troposphere. In winter, at high and mid latitudes a cyclonic vortex with its center near the pole is observed; in summer, at latitudes greater than 65°, anticyclonic motion prevails. Cyclonic motions prevail through the year at mid-latitudes, with anticyclonic circulations in the subtropics. On the average, the spring reversal of the circulation in the meteor zone occurs before that in the stratosphere--thus continuous monitoring of mesopause circulation can be used as a predictive tool in inferring the circulation of the upper stratosphere.

To date, the meridional circulation, which is of comparable strength to the zonal at meteor heights, has received little detailed attention. However, there does seem to exist a global system of motion from summer to winter pole, which agrees with the warm winter/cold summer mesopause inferred from high latitude rocket grenade temperature soundings.

One outstanding feature of the prevailing wind motions as deduced at mid-latitudes (see, for example, Elford, 1974) is the fact that the 85 km (mesopause) level is an obvious boundary between the mesospheric circulation below, and the thermospheric circulation above.

MEAN VERTICAL MOTIONS

One of the most puzzling features of winds measured in the 80 to 100 km height range is the apparently large (meters per second) mean motion which is inferred from both incoherent scatter and radio meteor wind observations. While the large magnitude of these winds is still in question, the sign of these motions, at least as inferred from the quasi-continuous observations over Atlanta, is consistent with the general circulation (Dolas, 1979).

TIDAL WINDS

The most outstanding feature of the motions at meteor height is the prevalence of solar tidal oscillations. On the average, more wind energy resides in tidal motions than in the prevailing circulation in the lower thermosphere.

While the irregularity of the diurnal tide, particularly in the northern hemisphere, has hampered the investigation of this tide as a global phenomenon, recent work by Mathews (1976) using the Arecibo Observatory incoherent scatter facility, and by the French with their portable meteor wind radar at Ramey, Puerto Rico (results as yet unpublished) has shown the presence of a stable, large amplitude (~50 m/sec) diurnal tide at the equinoxes.

In the southern hemisphere (Adelaide, 35°S, 139°E), the diurnal tide has a larger average amplitude than is found at northern hemisphere midlatitudes, but both phase and amplitude are highly variable in both hemispheres.

The semidiurnal tide, by comparison, exhibits relatively constant phase, but, again, great variability in amplitude. This variability can be explained by variations in the stratospheric/mesospheric wind fields through which the tidal energy is propagating to reach the mesopause. The variability has time scales related to the planetary wave time scales in the lower atmosphere, and further understanding of these variations can only come from long period (perferably continuous) observation by a network of stations widespread in both latitude and longitude.

STRATOSPHERIC WARMINGS

It has been known for some time that the sudden warmings of the polar stratosphere which occur during some northern hemisphere winters affect the circulation at mesopause levels at high latitudes. More recent measurements at mid-latitudes (Atlanta, 34°N, 84°W) have shown a direct cause/effect relationship, with major stratwarms causing zonal wind reversals. There are indications from the limited amount of data available from the southern hemisphere that northern hemisphere winter polar stratwarms may be a global phenomenon at the mesopause level.

Stratospheric warmings also affect tidal amplitudes and phases, but with a lack of continuous height/time profiles available globally, reports of such effects have been for the most part descriptive of particular locations only, and somewhat confusing in their interpretation.

MODELS

Two significant semiempirical models of stratospheric/mesospheric/lower thermospheric winds have emerged in the last decade. Groves (1970) took data from 1000 rocket launches with ejected sensors, 127 rocket grenade experiments, and 230 experiments with rocket released vapor trails and clouds, and constructed monthly mean zonal and meridional circulation patterns. This model, together with monthly mean tropospheric data, upper air global temperature maps, radio meteor winds, and the Jacchia thermospheric model have been incorporated in the Global Reference Atmospheric Model of Justus, et al. 1976. This empirical model generates latitude, longitude, and altitude dependent monthly mean values for pressure, density, temperature and winds from surface to orbital altitudes.

Theoretical models of atmospheric tides have recently been reviewed by Forbes and Garrett (1979). While there is a measure of agreement between the latest theoretical models and observational data, since the tides are a global phenomenon, questions as to standing versus travelling waves, and latitudinal variation will not be resolved until a global network of synoptic measurement capability is established.

An additional problem in resolving theory and observation centers around the variability encountered in all meteorological phenomena. The presence of other motions besides those of primary interest, in particular, the aliasing of tidal observational data by long period (planetary) waves, and the interaction between the tides and the shorter period random internal atmospheric gravity waves, makes a widespread, continuously operating monitoring network a necessity. An ideal radar for deployment is that described by Aso, et al. (1979).

COMPATABILITY OF MEASUREMENT TECHNIQUES

While early attempts to compare winds measured by different techniques (e.g., rocket released vapor trails, radio meteor winds, airglow drifts, and partial reflections radio observations) were singularly unsuccessful, more recent comparisons, in which allowances were made for sampling differences, in particular integration intervals in both height and time, have resolved these inconsistencies (see, for example, Felgate, et al., 1975; Geller, et al., 1976; Vincent, et al., 1977; and Hernandez and Roper, 1979). As yet unpublished comparisons between the French meteor radar located at Ramey, Puerto Rico, and Arecibo incoherent scatter facility, have shown excellent agreement. One outstanding achievement at Arecibo has been the extension of the lower limit of reliable wind determination by the incoherent scatter radar down to as low as 55 km.

The radio meteor wind technique is capable of continuously monitoring the dynamics of the 80 to 100 km altitude range, with a lower limit of observable variation of a few hours, and a height resolution of some 2 km. Specialized highpowered radars, such as that operated by the University of Illinois, can monitor variability of time constant less than one hour, but the logistics of highpowered operation preclude continuous sampling beyond intervals of a few days. The diurnal variability of meteor influx renders the error of wind determination greater at dusk than at dawn, but most facilities achieve a sufficiently high echo rate at dusk to make these results still statistically significant.

The partial reflections drift experiment is capable of sampling, with a time resolution of minutes, and a height resolution of 3 km, a height range from 60 kilometers to 110 km in daylight hours, but returns are reliably received only from 90 km up at night. The incoherent scatter technique suffers similarly from the diurnal variation of mesospheric ionization.

Rocket vapor techniques, while yielding excellent instantaneous snapshots of the stratosphere/mesosphere/thermosphere (height range governed by type of vapor released, and time of day) are prohibitively expensive if synoptic data is required. Their dependence, together with airglow observations, on cloud cover for observation is also a significant drawback.

Winds can be inferred in the stratosphere, mesosphere, and lower thermosphere by application of the thermal wind equation to satellite temperature soundings. Unfortunately, the large tidal winds at upper mesospheric heights and above, preclude the use of quasi-geostrophic approximations--the inference of viable winds from satellite temperatures at these altitudes is highly doubtful.

A further groundbased technique worthy of mention is the VHF radar technique (Rottger, et al., 1978). While not yet in routine operation, these VHF radars have demonstrated the capability of measuring winds and turbulence in the troposphere, stratosphere, and upper mesosphere.

Also, the VLF observations of Schminder and Kurschner (1979) warrant further investigation, since they appear to provide continuous, reliable observations of mesopause dynamics.

AVAILABILITY OF DATA

Beginning with the formation of GRMWSPM, the Global Radio Meteor Wind Studies Project of IAGA, in 1970, followed by the URSI/IAGA Cooperative Tidal Observations Program (CTOP), which added incoherent scatter radar observations, attempts have been made to coordinate observations (see Roper (1979)). Some of these results were published in the August, 1978, issue of the "Journal of Atmospheric and Terrestrial Physics."

While data is available from individual observers, as yet formats and analysis techniques have not been standardized, and no central depository exists. The Soviet Union is moving toward having all their radio meteor wind data available through the World Data Center in Moscow. The mechanics of such archiving, both in the Soviet Union and elsewhere, are to be discussed at a meeting of IAGA Division V, Working Group 2 (Meteor Observatories) at the IUGG Assembly in Canberra, Australia, in December 1979.

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CHAPTER V

DYNAMICS OF THE EQUATORIAL MESOPAUSE

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ABSTRACT

Until recently, our view of the neutral atmosphere dynamics of the equatorial mesopause has been restricted to "snapshots" resulting from either one time campaigns, or intermittent observations. Rockets flown from Brazil, the "big gun" launches from Barbados, meteor wind radar data from Somalia and Jamaica, and VHF radar observations from Peru, while producing much information on the nature of motions with periods of up to a day, provided only meager details of the synoptic seasonal meteorology of the region. More recent observations have resulted from the extension to lower altitudes of the incoherent scatter technique at Arecibo, the installation of a partial reflection drift station at Townsville, Australia, and the establishment by the French of a meteor wind radar at Ramey, Puerto Rico. The incoherent scatter and partial reflection facility operate in the "campaign" mode while the French meteor radar is operating continuously.

Because of the limited amount of synoptic data available, this review concentrates on results produced by those techniques most likely to be used for synoptic observations in the future. However, since even these promising techniques may not be able to be used in other than a campaign mode (for reasons ultimately financial), a rationale for their use in winter months (on the basis that northern hemisphere winter polar stratospheric warmings are probably a global phenomenon at mesopause altitudes) is presented.

INTRODUCTION

Many observational techniques have been used to measure winds at mesopause heights over the tropics - chemical trails from projectiles launched by the "big gun" on Barbados (Murphy, et al., 1966), rocket grenades over Natal, Brazil (Groves, 1974), radio meteor measurements over Somalia (Babajamov, et al., 1970), and Jamaica (Alleyne, et al., 1974, Scholefield and Alleyne, 1975), VHF radar observations (Woodman and Guillen, 1974), and incoherent scatter observations (Mathews, 1976; Harper, et al., 1980). More recently, a partial reflection drift station has been set up at Townsville, Australia by the Department of Physics of the University of Adelaide (Dr. R. A. Vincent), and a meteor radar by the French National Center for Telecommunications Studies (Dr. M. Glass) at Ramey, Puerto Rico. The various techniques have been reviewed most recently by Woodman (1977) and Evans (1978).

Woodman places particular emphasis on equatorial measurements and the need for more observational data. His paper at this meeting concentrates on small scale structure (gravity waves and turbulence), and so this present work will confine itself to tidal and planetary waves and the prevailing winds.

OBSERVATIONS

To date, all observations of winds at the mesopause level over the equator have been of the "snapshot" variety, with a few days of observation, very occasionally repeated seasonally. (The same can also be said for other latitudes, with the exception of the meteor radar network in the U.S.S.R. operated by the Hydrometeorological Service, which operates continuously, but unfortunately does not measure wind structure with height.) These "snapshots" have revealed considerable variability in the winds, but with indications of predominant easterlies at the equator, in contrast to the predominantly westerly flow observed at mid latitudes.

RADIO METEOR WINDS

An example of one of these observations, obtained by a collaborative effort between the French National Center for Telecommunications Studies, and the Georgia Institute of Technology, using the French meteor radar located at Ramey, Puerto Rico (18°N), is presented in Figure 1. The zonal wind component of Figure 1a is predominantly easterly (wind vector directed toward the west), for the first four days of observation, but a strong westerly intrusion then develops from above. The intrusion weakens and ascends toward the end of the observational period. While this phenomenon can be associated with the simultaneous spring reversal of the zonal stratspheric circulation, it may also be evidence for the mean flow acceleration by tidal interaction proposed by Miyahara (1979) - as can be seen from Figures 1b and c, both the

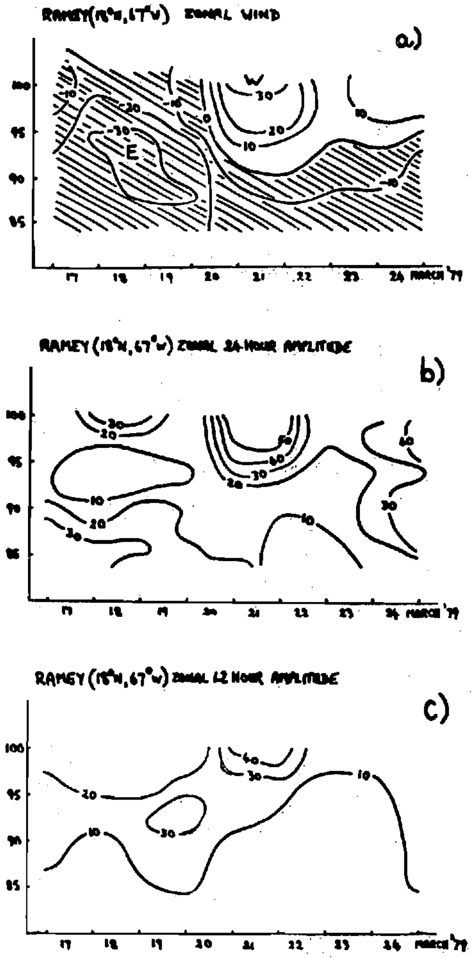

Figure 1. Radio meteor winds from 85 to 105 km for the period March 17-24, 1979, over Puerto Rico. a) zonal wind; b) 24 hour tidal amplitude; c) 12 hour tidal amplitude.

diurnal and semidiurnal tidal amplitudes have large amplitudes coinciding with the reversal.

One phenomenon which has received considerable attention at middle latitudes is the 2 day wave, which has been observed in week to 10 day meteor radar winds in both the northern and southern hemisphere. By using a low pass filter to remove periods shorter than 30 hours from 8 days of meteor wind radar data obtained at Ramey in August-September 1977, Glass (private communication, 1978) produced the results shown in Figure 2. The presence of a wave with a quasi-2 day period is well illustrated.

INCOHERENT SCATTER WINDS

Some recent results from Harper, et al. (1980), taken using the Arecibo incoherent scatter radar from September 1 to 14, 1977, are shown in Figure 3. The 14 noontime profiles illustrate the variability encountered in winds at the mesopause level. Note that these are meridional winds and that speeds of ±100 m/sec occur - zonal and meridional wind speeds are comparable at altitudes above the mesopause, in contrast to the predominantly zonal flow in the stratosphere and mesosphere below.

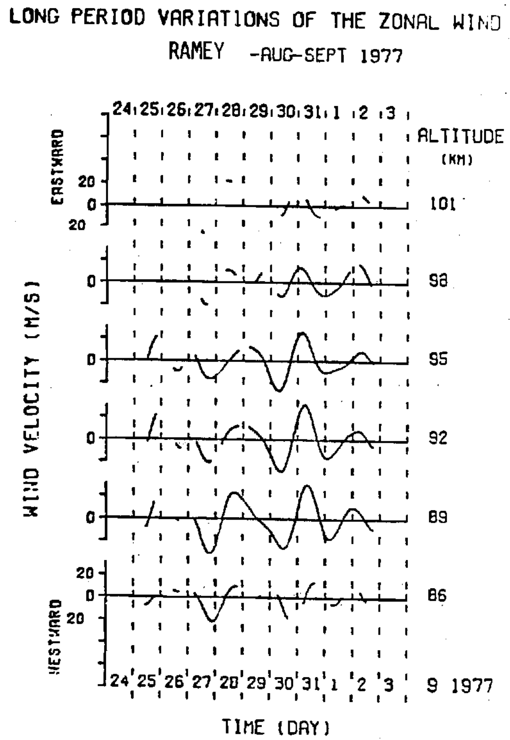

Figure 2. Radio meteor winds - long period variations.

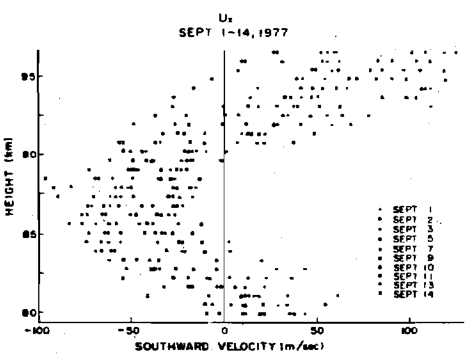

Figure 3. Meridional winds from the Arecibo incoherent scatter radar (from Harper, et al., 1980).

METEOR/INCOHERENT SCATTER WIND COMPARISONS

Mathews, et al. (1980) presents several comparisons between meteor winds from the

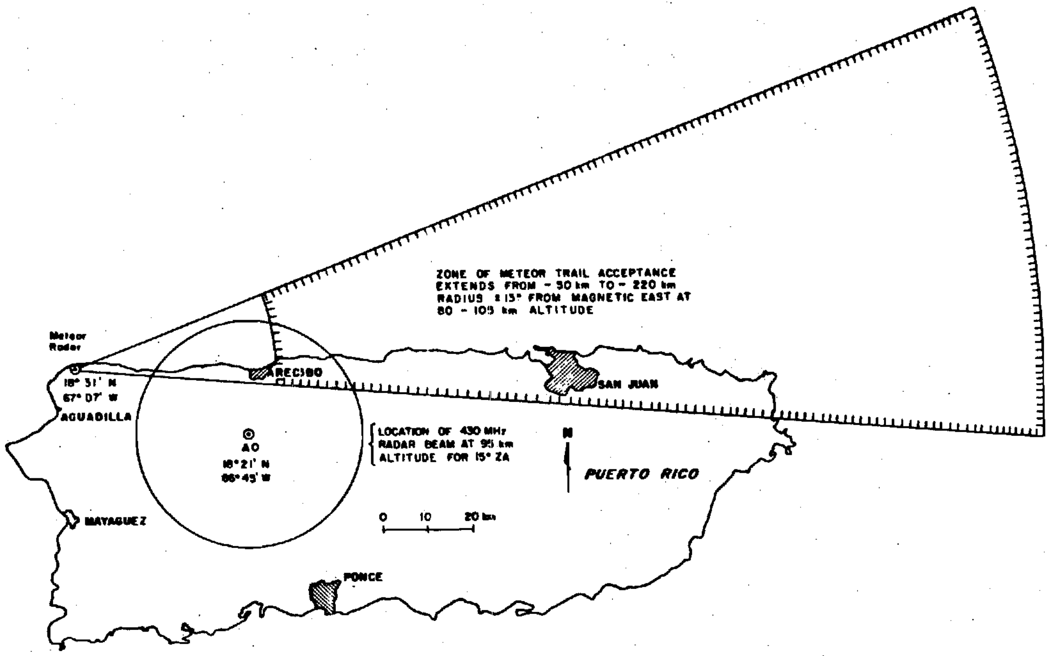

Figure 4. Location and sampling regions of Ramey meteor and Arecibo incoherent scatter radars (from Mathews, et al., 1980).

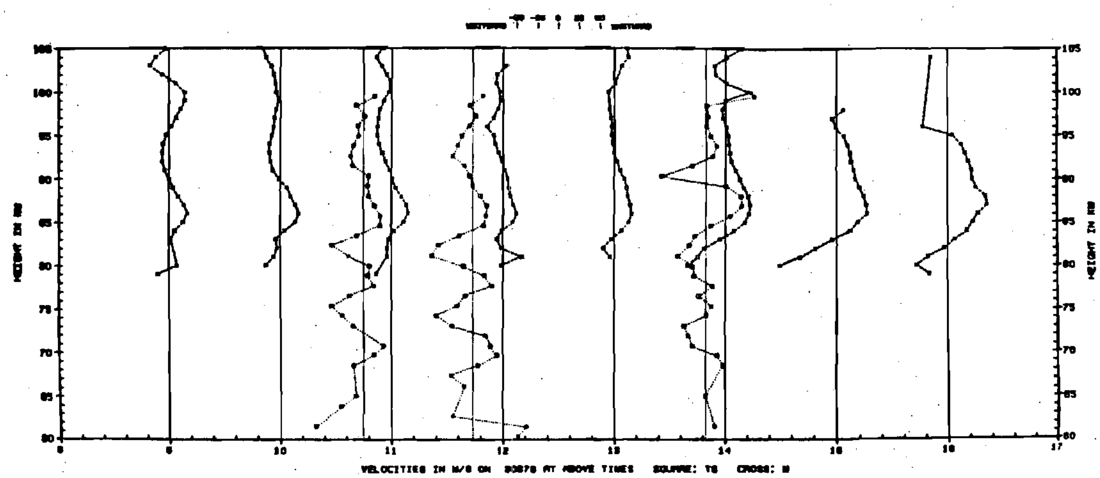

Figure 5. Comparison between Arecibo incoherent scatter radar winds (dashed lines) and Ramey meteor radar winds (full lines), August 3, 1978 (from Mathews, et al., 1980).

French radar at Ramey and winds from the incoherent scatter radar at Arecibo. The locations and "zones of echo acceptance" of the two instruments are shown in Figure 4. One comparison, from 0900 to 1600 hours on August 3, 1978 (Figure 5), shows quite good agreement between the two techniques, with the incoherent scatter winds showing more structure because of better height resolution. Between 1000 and 1400 hours, the incoherent scatter winds are significant down to 60 km - a real breakthrough when one remembers that just a few years ago 110 km was the lower limit of reliable wind measurement using this technique. However, there are no incoherent scatter winds below 100 km at night, whereas the radio meteor technique produces winds over the 80 to 100 km height range both day and night.

PARTIAL REFLECTION DRIFTS

This relatively inexpensive technique is capable of measuring winds between 60 and 100 km by day, and 90 to 100 km by night. The first partial reflection drifts experiment operating on a routine basis was installed at

Townsville, Australia (20°S) by Dr. R. A. Vincent of the University of Adelaide, Australia. This equipment is capable of continuous operation, but for financial reasons is run in the campaign mode. Results from this experiment are eagerly awaited.

VHF SCATTER RADARS

The most recently developed tool capable of continuous monitoring of mesopause dynamics is the VHF Scatter Radar. These radars have the capability, shared by the partial reflection drift technique, of measuring periodicities in wind profiles down to the Brunt Vaissala, and therefore represent powerful tools for gravity wave and turbulence studies, in addition to planetary waves, mean winds and tides. Details of this technique are to be presented at this meeting by Ben Balsley.

AIRGLOW DRIFTS AND TEMPERATURES

While unable to provide continuous measurements (nighttime only, no clouds), airglow measurements can supplement our knowledge of winds (see, for example, Hernandez and Roper, 1979) and provide temperature data of use in modeling mesopause dynamics.

POLAR STRATWARM EFFECTS AT MESOPAUSE HEIGHTS

The necessity for continuous monitoring of mesopause dynamics is illustrated in Figure 6 (from Roper, 1978). Without

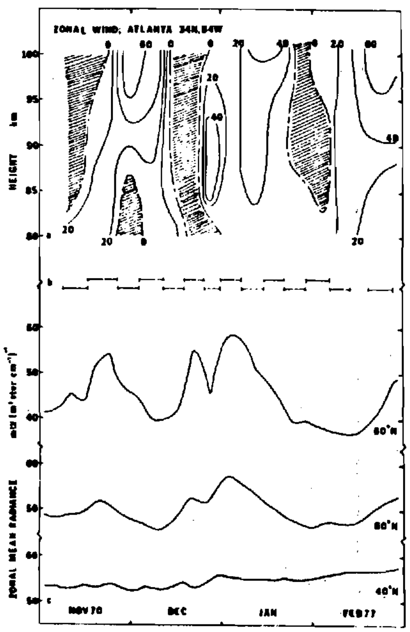

Figure 6. Comparison between zonally averaged satellite radiance data and the zonal wind at meteor heights over Atlanta (from Roper, 1978).

continuous data, the association between the polar winter stratospheric warming and the reversal in the zonal wind at mesopause, altitudes would not be nearly as convincing. While this data is for mid-latitudes (Atlanta, 34°N), the effect is so pronounced that it would not be surprising if it extends as far south as Puerto Rico (18°N).

CONCLUSIONS

No mention has been made here of satellite remote sensing to monitor meospause dynamics. Because winds at these altitudes are ageostrophic, conversion of temperature fields to winds is well nigh impossible. For many years to come, and certainly through the Middle Atmosphere Program period, 1982-85, ground based radio techniques will be our primary source of data on mesopause dynamics at all latitudes.

ACKNOWLEDGEMENTS

Radio meteor winds from Ramey have been measured by the French National Center for Telecommunications Studies, and, under a grant from the Atmospheric Research Section (Aeronomy) of the National Science Foundation, by the Georgia Institute of Technology. Preparation of this paper has been supported by the Planetary Atmospheres Branch of the National Aeronautics and Space Administration. The Arecibo Observatory is funded by the National Science Foundation through a grant to Cornell University.

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CHAPTER VI

UPPER ATMOSPHERIC MIXING BY GRAVITY WAVES

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Abstract

Magnitudes and horizontal and vertical scales of gravity waves have been measured and used in a formula developed by Hines to estimate resulting eddy diffusion coefficients. Values between 90 and 110 km agree well with those determined from energy flux measurements from chemical cloud releases. Values between 25 and 45 km altitude are in agreement with high eddy diffusion coefficients in the upper stratosphere recently determined by Wofsy. Comparison with other eddy diffusion estimates is also given.

1. Introduction

A recent study of upper atmospheric data in the region above 25 km has been conducted for the purpose of resolving small scale variations which may be associated with internal atmospheric gravity waves. The method used, known as the daily difference technique, was developed for resolving small scale irregular variations from limited amounts of data (i.e. of insufficient number for direct Fourier analysis). The form of the method used in this study is a slight extension of the method originally developed by Woodrum and Justus[1,2]. The magnitudes and horizontal and vertical scales of the small scale perturbations, assumed to be produced by gravity waves, have been measured by this technique. The details of the daily difference method and the results of these magnitude and wavelength measurements have been presented by Justus and Woodrum[3].

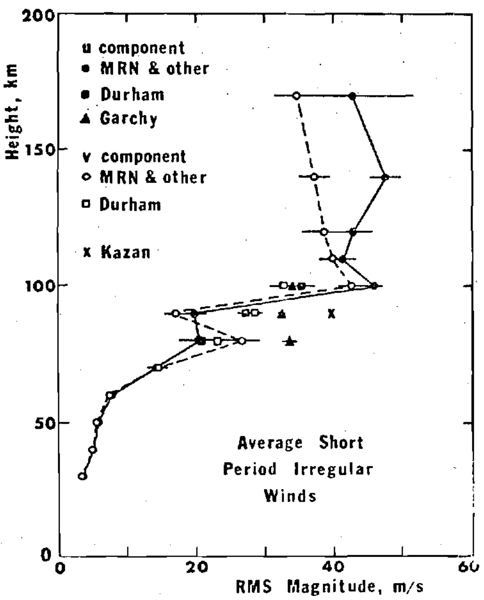

Fig. 1. Gravity wave wind magnitudes

Figure 1 shows the computed height variation of the gravity wave winds averaged over all seasons and locations from Meteorological Rocket Network, falling sphere, meteor radar, grenade release, and chemical release data. Horizontal and vertical wavelength results are shown in Figure 2. Data points

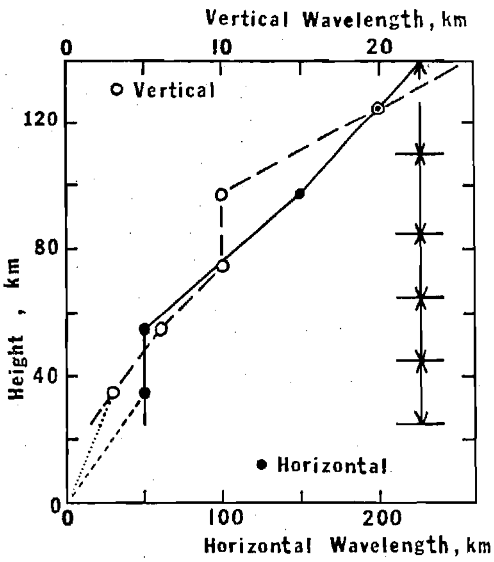

Fig. 2. Gravity wave horizontal and vertical wavelengths

in Figure 2 are plotted at the midpoints of the altitude intervals over which they were evaluated. These averaging intervals are indicated on the right hand side of the figure. Figure 2 shows that horizontal wavelengths are approximately 10 times larger than vertical wavelengths throughout the altitude interval measured.

2. Eddy Diffusion Coefficients

Hines[4,5] has developed a relation for the eddy diffusion coefficient K_D corresponding to "offset condition" dissipation of internal atmospheric gravity waves. The offset condition is the situation in which viscosity and thermal conduction just offset the normal (i.e. no-viscosity) exponential increase of gravity wave amplitudes, yielding constant amplitude. The formula, expressed in terms of horizontal and vertical wavelengths λ_x and λ_z, pressure scale height H, and Brunt-Vaisala period τ_g, is given by

$$K_D = 0.014\ \lambda_x^{\ 4}\ \lambda_z^{\ 4}\ (\lambda_x^{\ 2} + \lambda_z^{\ 2})^{-5/2}\ (\tau_g\ H)^{-1} \qquad (1)$$

The wavelength values from Figure 2, together with the 1962 U. S. Standard Atmosphere values for τ_g and H have been used to evaluate K_D from equation (1). The results are shown in Figure 3. The solid square data points correspond to the wavelength average value data points of Figure 2. The solid curve represents values computed from the interpolation curves of wavelength values in Figure 2. Some extrapolation of wavelength values was employed, but not beyond the altitude limits of the original data from which the wavelengths were determined. The portion of the solid curve between 25 and 35 km in Figure 3 is from the dashed extrapolation curve in Figure 2, and the dashed portion of the present results curve in Figure 3 is from the dotted extrapolation curve (which goes to zero at the Earth's surface) in Figure 2. In the case of $\lambda_z << \lambda_x$, equation (1) reduces to

$$K_D \approx 0.014\ \lambda_z^{\ 4}\ (\tau_g\ H\ \lambda_x)^{-1} \qquad (2)$$

Thus, since $\lambda_g \approx \lambda_x/10$ the K_D values vary approximately as the fourth power of λ_z, but inversely with λ_x. It is for this reason that small changes in λ_z in the 25 to 35 km altitude interval (where λ_z is small and small changes represent large percentage changes) make considerable difference in the values of K_D computed from equation (1).

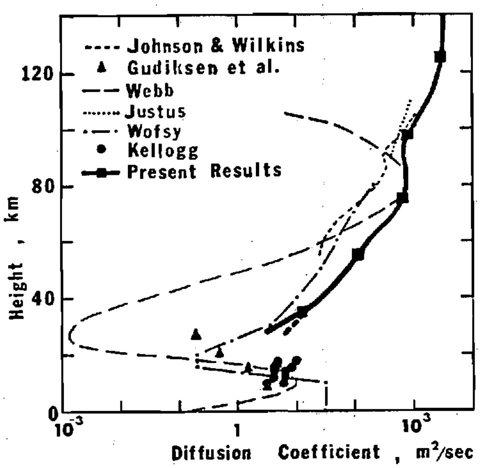

Fig. 3. Eddy diffusion coefficients

The gravity wave magnitude results of Figure 1 show that the gravity wave amplitudes are, on the average, increasing with height, contrary to the assumed offset conditions necessary for application of equation (1). However, as shown in Figure 4, there is considerable loss of gravity wave kinetic energy density with height. Between about 25 and 60 km the kinetic energy density $E = \rho V^2/2$ varies as $E = E_o \exp(-z/9.4)$ with E_o a constant and z in km. Between about 70 and 130 km E varies as

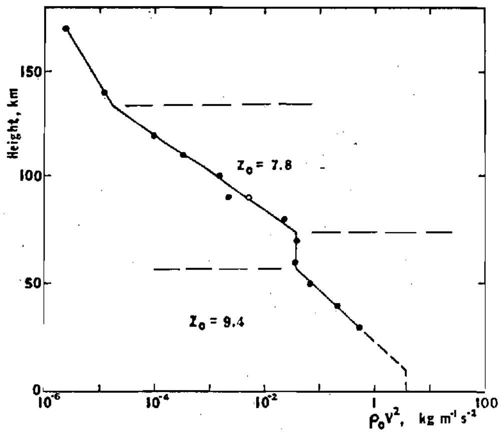

Fig. 4. Gravity wave kinetic energy density

$E_o \exp(-z/7.8)$. These values of "kinetic energy density scale height" which are not significantly larger than atmospheric density scale height, indicate that the actual rate of gravity wave dissipation more nearly corresponds to the required offset condition than the ρV^2 = constant situation without viscous effects.

3. Comparison With Other Results

Figure 3 also shows values and curves of eddy diffusion coefficients which have been measured or used below 110 km. The dashed curve denoted Webb[6] was also based on earlier work of Lettau[7] in the neutral atmosphere altitudes. The curve labeled Justus[8] was determined from measured rates of turbulent kinetic energy production ε_s by wind shear, and mean wind shear S, with ε_s assumed to be related to K_D and S by $\varepsilon_s = K_D S^2$. The curve labeled Johnson and Wilkins[9] results from their computation of the maximum eddy diffusion consistent with the amount of heat available for downward eddy transport. The data points labeled Gudiksen et al.[10] were determined by these authors from their observations of nuclear explosion released tungsten 185. The data points labeled Kellogg[11] were computed by the present author from Kellogg's reported values of turbulent velocities (v_t) and maximum observed radius (r_{max}) of smoke puffs in the 10 to 18 km height range. The data points are values of the product $v_t\ r_{max}$, which should represent a lower limit to the eddy diffusion coefficient if, as suggested by Kellogg, the linear growth velocity is due to increasing eddy diffusion coefficients with increasing scale. However, as discussed below, this conclusion is quite likely incorrect.

4. Interpretation of Results

Because vertical spread within the atmosphere is ultimately restricted compared to the unlimited horizontal spread which is possible, the large scale eddy diffusion process is not isotropic, and vertical eddy diffusion coefficients are generally smaller than horizontal eddy diffusion coefficients.

The first point to be noted with regard to the eddy diffusion coefficient data in Figure 3 is that (with the probably exception of the values computed from Kellogg's smoke puff data) these results are for vertical eddy diffusion coefficients. The Johnson and Wilkins, Wofsy, and Gudiksen et al. data all relate theoretically and/or observationally to vertical fluxes of diffusing constituents or heat. As a measure of the degree of anisotropy possible, it should be noted that the purely horizontal (meridional) eddy diffusion coefficients determined by Gudiksen et al. are roughly a factor of 10^6 larger than their purely vertical eddy diffusion coefficients plotted in Figure 3.

It is likely that the $v_t\ r_{max}$ values from Kellogg's smoke puff data overestimate the vertical eddy diffusion coefficients, because apparently the actual diffusive spread of cloud releases on small scales proceeds faster than would be accounted for by the large scale vertical eddy diffusion coefficient. This can be seen from comparison of diffusion velocity-scale products (i.e. $v_t\ r_{max}$) from chemical release clouds[12] with the eddy diffusion coefficient values (labeled Justus in Figure 3) determined from vertical energy flux measurements using chemical release data. The velocity-scale product values are consistently at least a factor of ten larger. This result is undoubtedly caused by the small scale diffusive spread being controlled by wind shear processes.

The present results curve in Figure 3 should most precisely be interpreted as a maximum value which could be produced by internal atmospheric gravity waves without an upward decrease in the amplitude of the waves. Therefore, this curve might require a slight shift to lower values before it would represent the profile of actual mean vertical eddy diffusion coefficients. Nevertheless, it is clear from Figure 3 that the present results agree with high stratospheric eddy diffusion values such as those of Wofsy and cannot agree with the Webb-Lettau curve which has stratospheric eddy diffusion coefficients which are lower by several factors of ten.

The important conclusion for these results, with regard to the assessment of environmental impact of aerospace operations, is that the vertical eddy diffusion coefficient in the stratosphere, and especially in the upper stratosphere, is higher than originally believed. From the results presented here it is reasonable to attribute these high eddy diffusion coefficients to the mixing processes generated by atmospheric gravity waves as they propagate through the stratospheric region.

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CHAPTER VII

Stratospheric Warming Effects

on Mesopause Dynamics

a) The Effects of Polar Stratwarms on the Winds at the Mesopause Level in Mid-Laditudes

(Presented at the 18th Radar Meteorology Conference, Atlanta, Ga., March, 1978, and updated at the AMS Conference on the Upper Atmosphere, Boston, Mass., October, 1978)

b) Prevailing Wind in the Meteor Zone (80-100 KM) Over Atlanta, and its Association with Mid-winter Stratospheric Warming

(Submitted to the Journal of Atmospheric Science)

Reprinted from Preprint Volume: 18th Conference on Radar Meteorology, Mar. 28-31, 1978, Atlanta, Ga. Published by the American Meteorological Society, Boston, Mass.

THE EFFECTS OF POLAR STRATWARMS ON THE WINDS AT THE MESOPAUSE LEVEL IN MID LATITUDES

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1. INTRODUCTION

For over a quarter of a century, winds have been measured over the height range 80 to 100 km by means of radio reflections from the ionized trails of meteors. Although sophisticated high frequency radars (usually operating in the 20 to 40 MHz range) have been employed, not all of the advantages of the technique have been realized. In particular, very few continuous, long term observations have been made. In the past, the major problem has been an inability to handle the large amount of data generated by continuous operation. However, the revolution in data acquisition in recent years had made such operation economically feasible.

2. THE METHOD

The radio meteor method of wind determination has been described at a previous AMS Radar Meteorology Conference (Barnes, 1972), and the Georgia Tech Radio Meteor Wind Facility detailed by Roper (1975). The Tech system is a continuous wave radar (using two continuously transmitted carriers 720 Hz apart to determine range) designed to measure some 1000 line of sight dopplers and echo positions per day. However, since the CW technique accepts echoes at all ranges, aircraft reflections are a problem, at times reducing the rate below 300 echoes per day. This, together with time lost through maintenance of the transmitter and receivers, occasional power outages at the receiving site, and the need to average several days data to produce meaningful measurements of the diurnal and semidiurnal tides, results in an apparent smoothing of the data of from a few days to two weeks. In order to meet the publication deadline, consideration is given here only to the zonal mean wind; analysis of the significance of the meridional and vertical mean winds, and the tidal winds, is proceeding.

3. RESULTS

Continuous radio meteor wind observations commenced over Atlanta in August, 1974. Figure 1a shows the variation with height and time of the zonal wind 80 to 100 km over Atlanta for the period November, 1974 through February, 1975. In interpreting the structure present, the averaging interval details in Figure 1b should be taken into consideration.

An attempt was made to correlate the observed winds with the zonal mean satellite rad-

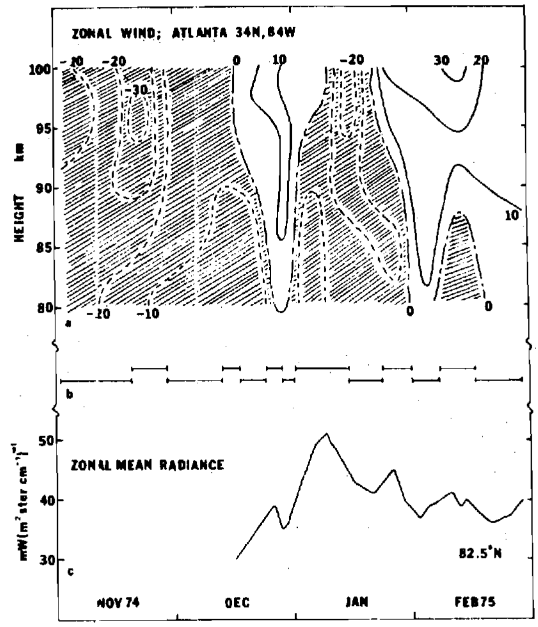

Figure 1a) The mean zonal wind over the height range 80 to 100 km above Atlanta for the period November, 1974 through February, 1975. The zonal means to which the contours have been fitted are averages for the measurement intervals depicted in b). c) is the zonal mean radiance deduced from satellite radiometer measurements, and is proportional to the stratospheric temperature averaged over the 100 to 5 mb height range at 82.5°N.

iance data at 82.5°N published by Quiroz et al (1975). The late December/early January warming did not produce a wind reversal at the 10 mb level, therefore is not characterized as a major warming.

The circulation at meteor heights during the fall and early winter of 1974, being predominantly easterly (as plotted, a negative wind is a wind vector directed toward the west, i.e. an easterly), is unusual, at least when compared to subsequent years (see Figures 2 and 3). About the best one can say from this preliminary comparison of the meteor winds with the single zonal mean radiance data curve is that there are wind reversals which

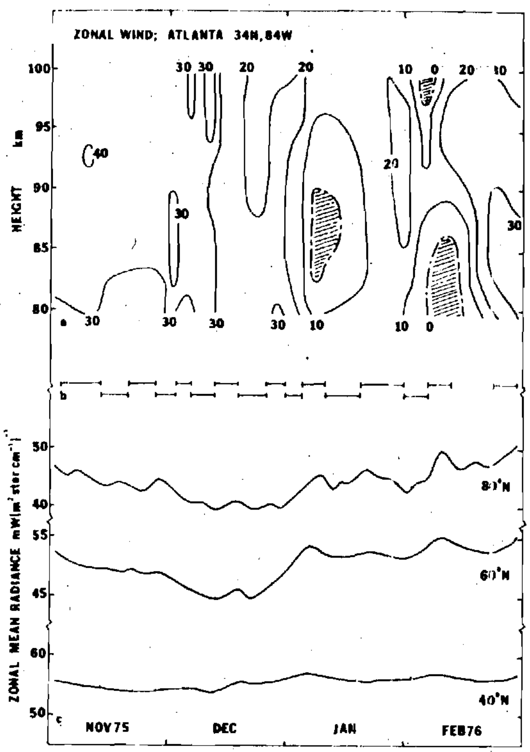

Figure 2. As for Figure 1 for November, 1975 through February, 1976, except that the zonal mean radiance is presented for three latitudes, and is proportional to the stratospheric temperature averaged over the 100 to 2 mb height range.

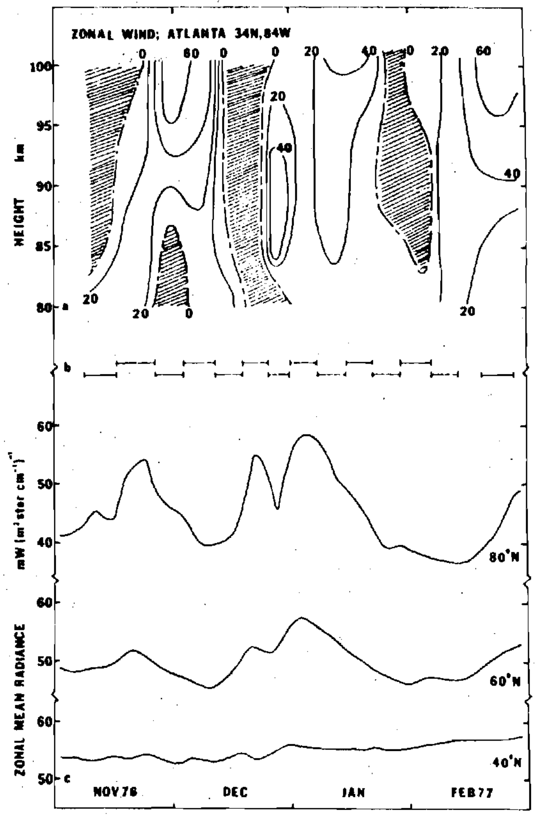

Figure 3. As for Figure 2, for November, 1976 through February, 1977.

correlate with the changes in radiance.

Proceeding to the winter of 1975-76 (Figure 2), the zonal flow at meteor altitudes is predominately westerly, with two minor reversals - in early January, and early February. Comparison with the zonal mean radiance satellite data, this time available for three latitudes (Quiroz, private communication, 1977) shows that there are local maxima in the radiance curves at all three latitudes during these two periods. However, there are not any outstanding features in the radiance data - no warmings occurred during the winter of 1975-76.

The data for the winter of 1976-77 represents the most interesting set to date. Unfortunately, no meteor wind data is available for the first week in November (and the last week in October). However, the winds through August, September and mid October were consistently westerly - thus the change to easterly in mid November could well be associated with the warming of late November. With the reestablishment of westerlies over the whole height range by the end of November, the onset of the late December warming is preceded by a rapid waning of the westerly at the upper levels, with the onset of easterly flow above 90 km occurring a week before the winds at 80 km become easterly. During the last week in December, the satellite radiance data show a reversal of the latitudinal temperature gradient from that established by the warming. This is accomplished by what one would almost call a westerly "jet" in the mid height range of the meteor winds. The subsequent warming trend in early January is not accompanied by a meteor wind reversal, and just so that we will not be tempted to think that we have all the answers, there is a tongue of weak easterlies which descends from 100 km to 83 km in late January/early February, which seems to have no correlation with the stratospheric temperature curves.

4. CONCLUSIONS

The continuous measurement of the wind profile over the 80 to 100 km region by means of radio reflections from meteor trails provides data which can be correlated with stratospheric temperature changes inferred from satellite radiance data. The very preliminary results presented here (later publications will include the consideration of the meridional wind, as well as the diurnal and semi-diurnal tidal winds) show a tendency for stratospheric warming events (which produce zonal mean temperature changes in the stratosphere at 40°N which are only just measurable) to produce dramatic changes in the circulation in the neighborhood of the mesopause at 34°N. While such changes have previously been expected and reported at higher latitudes (see, for example, Hook, 1970), to my knowledge this is the first report of similar behavior as far south as 34°N.

5. ACKNOWLEDGMENTS

Special thanks are due to Dr. R. S. Quiroz, of the National Meteorological Center, Washington, D.C., for providing the satellite radiance data used in Figures 2 and 3, prior to their publication.

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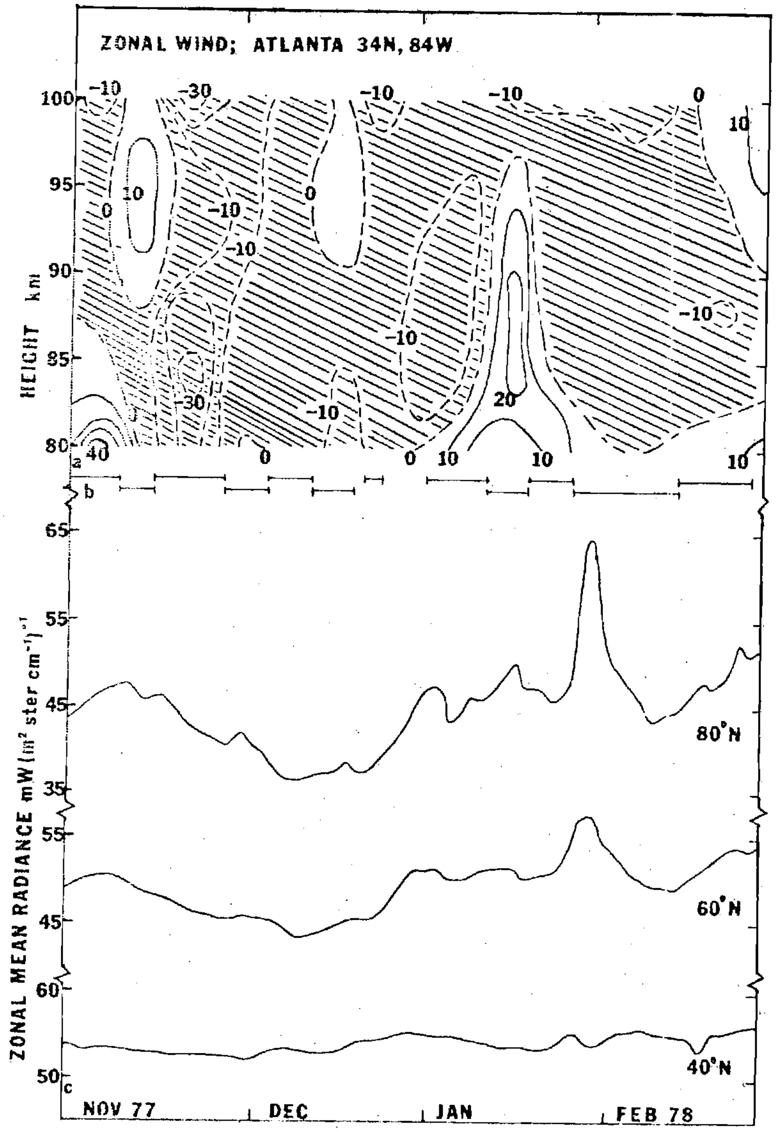

PREVAILING WIND IN THE METEOR ZONE (80-100 KM) OVER ATLANTA AND ITS ASSOCIATION WITH MID-WINTER STRATOSPHERIC WARMING

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ABSTRACT

The wind data generated by an all sky, continuous wave radio meteor wind facility at Atlanta (34°N, 84°W) is analyzed over the period of August 1974 through July 1975. Zonal and meridional components of the prevailing wind over the height range of 80-100 km, at 2 km interval represent 5 to 10 day averages where the tidal components have been removed.

Large southerly wind during winter and weak northerly wind during summer at 80-100 km altitude is consistent with other observations and mesospheric circulation models. Evidence of wave energy input from the lower atmosphere has been found in the prevailing wind.

The "minor" mid-winter stratwarm of 1974-75 is studied. Various phases of this warming, including a pre-warming pulse during the second half of November 1974, are shown to have affected the prevailing wind regime in the meteor zone, in a manner consistent with the latitudinal and vertical temperature compensation criteria in the stratosphere and above.

The implied meriodional mean temperature gradients from the one year of data point to a year round cold mesopause at high latitudes--the only exceptions during winter appear related to the perturbed stratospheric cyclonic vortex due to warming episodes.

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1. Introduction

The general circulation in the stratosphere and lower mesosphere consists primarily of ultra-long wave motions, superimposed on a zonally symmetric flow that is easterly (from the east) in the summer hemisphere and westerly in the winter hemisphere, in approximate thermal wind equilibrium with the zonal mean temperature field. The planetary scale waves of zonal wave numbers 1 and 2 appear to be vertically propagating modes that are generated in the troposphere by a variety of dynamic mechanisms, and then transport energy and momentum into the stratosphere and beyond. In addition, there is a semi-annual oscillation which has its maximum amplitude near the equatorial stratopause, and there is also a strong oscillation in the lower tropical stratosphere of somewhat irregular period, averaging about 26 months, called the "quasi-biennial oscillation". Besides these cyclic variations, there are a number of irregular short-term and year-to-year variations which are apparently due to the tropospheric forcing of vertically propagating planetary waves. The most prominent are the changes in the wind and temperature regimes associated with sudden warming of the high latitude winter stratosphere (the so-called "polar stratwarms").

On the other hand, based upon very sparsely distributed data, in time and location, a general profile of the seasonal variations of the prevailing wind in the lower thermosphere (up to 120 km) is becoming established. *Groves* (1971) finds that there are equatorial easterlies present throughout the year at an altitude of 95 $\pm$ 15 km. The seasonal variation observed at low latitudes is the result of the slight shifting of the wind belt north and south of the equator. Recently *Minina et al.* (1977) have analyzed the wind data from many years at heights of 80-100 km, primarily from meteor radar observations made at various Northern Hemisphere stations. These data indicate large scale pressure formations of the cyclone-anticyclone type, with similar synoptic structures as observed in the atmosphere below. During winter the cyclonic vortex centered near the pole dominates the circulation in the meteor zone at high and middle latitudes (up to at least 120 km; *Groves*, 1971); in summer, the anticyclonic circulation is characteristic of high latitudes (north of 65°N). The cyclonic circulation in the middle latitudes and the ring of sub-tropical highs exist year round. Thus, the winds in the meteor zone are shaped not only by the seasonal variations of the thermal balance at these heights, but also by the lower-lying layer of the mesosphere.

The present work investigates the imprint on the general circulation at lower thermospheric heights (80-110 km) of the very large perturbation of the high latitude winter stratosphere which occurs during a polar stratwarm event. *Hook* (1972) using the radio meteor technique at College, Alaska (65°N, 140°W), and *Gregory and Manson* (1975) using the partial reflection technique over Saskatoon, Canada (52°N, 107°W), have studied the effects of various major stratwarms on the wind patterns in the 80-110 km altitude range. They observe that in association with the warming phase of the stratwarm, there was found a reversal of zonal and meridional wind from westerly and southerly prevailing directions respectively. *Gregory and Manson* (1975) deduce meridional temperature gradients from their wind data and believe that the pattern of temperature changes in the vertical as suggested by *Labitzke* (1972a,b; see next section) in relation to the stratwarms is consistent with Saskatoon data. Other observational studies, for example by *Ryazanova et al.* (1976), may also be cited for higher latitudes.

In the present study, the radio meteor technique has been used to obtain wind data in the height range of 80-100 km over Atlanta (34°N, 84°W). The prevailing wind regime over this height range is analyzed for the period of August 1974-July 1975. Particular emphasis is placed on the circulation changes observed at meteor heights over this low mid-latitude station as related to the minor high latitude stratospheric warming that peaked during the first week in January 1975. This warming event is studied in the next section. A brief description of the meteor radar observation technique follows. Finally, the prevailing wind at the mesopause level is analyzed in subsequent sections.

2. Stratospheric Warmings

A. *The Minor Stratospheric Warming of 1974-75*

Since its discovery by Scherhag in 1952, the anomalous warming of the winter stratosphere has come to be recognized as the single most energetic disturbance of the entire stratosphere. Individual warming events have been termed major or minor depending on whether large scale circulation reversals were observed in the middle stratosphere (at or below 10-mb, ~30 km, level) in association with the temperature increases observed (of at least 25° in a period of a week or less), in the higher latitudes (*McInturff*, 1978).

The mid-winter warming of 1974-75 satisfied all but one criterion required to be classified as major ; that is, the mean westerly circulation poleward of 60°N did not reverse to easterly at 10-mb level. Figures 1a and 1b show 2-mb (~42 km) charts for December 11, 1974 and January 1, 1975 respectively. On December 11, 1974 (Figure 1a), representing the pre-warming upper stratosphere, a cyclonic vortex is situated close to the pole and the cold center (dashed lines represent constant temperature contours) lies about 90° westward. The chart for January 1, 1975 (Figure 1b) shows the state of the circulation close to the peaking of the warming event. Here the low pressure center is seen displaced to a position east of Greenland at about 70° latitude and the Aleutian high centered over the Bearing Straits governs the polar circulation. As shown in Figure 2, a plot of the mean zonal winds, easterlies are to be found north of 55°N at the 2-mb level on January 2, 1975. By January 15, weak westerlies appear again and within a week the stratospheric circulation returns to its near normal state.

Figure 3, which is a plot of zonally averaged temperatures, shows that the warming reached its peak during the first week of January 1975 in the upper stratosphere. On January 2, 1975, an increase in temperature of 40°C is observed at the 2-mb level, north of 70°N. Temperatures return to normal in the upper stratosphere only after the second week of January 1975. Figure 3 also indicates a low latitude warming in association with the cooling at high latitude (on January 8 and 15 contours). *Fritz and Soules* (1972) have shown that for the major warming of December 1969-January 1970 in high latitudes based upon satellite radiance data (weighting function in the middle stratosphere), a simultaneous decrease is observed in the low latitudes of both the hemispheres and more importantly, a phase lag is observed in the radiance maxima from low to high altitude.

Warming pulses of about 15-day periods have been found in the winter stratosphere of both hemispheres, and are readily detected from the satellite borne radiance measurements. In Figure 4 the mean zonal radiance (solid line) at 82.5°N is plotted for the winter of 1974-75. The peak marked II, during the first week of January coincides with the "minor" warming described above. In all, six peaks can be discerned over the period of three months.

B. *Temperature Changes up to 80 km*

During a stratwarm the height of the stratopause, defined as the temperature maximum, can vary from 20 to 60 km altitude in high latitudes. The mesopause height however, seems to vary over a much narrower range, possibly from 70 to 90 km altitude.

Labitzke (1977a) has combined several soundings from different winters to give profiles for high and low latitudes for a "composite warming" (for early winter (I), climax (II) and late stage (III)) as shown in Figure 5. Here, opposite changes in temperature between stratosphere and mesosphere (right part, Figure 5), and between high and low latitudes are apparent. The region of sign change lies between

50-55 km and Labitzke call it "the pivotal level for the interaction between stratosphere and mesosphere in winter".

3. Wind Observations Using Meteor Radar

A. *The Georgia Tech Radio Meteor Wind Facility*

The Georgia Tech system (at Atlanta; 34°N, 84°W) has been in routine operation since July 1974. It has been designed as a continuous wave, all sky system with a capability of continuous operation 24 hours a day, seven days a week with an adequate usable echo rate (*Roper*, 1975a, b).

A double sideband suppressed carrier continuous wave transmitter operating on 32.5MHz$\pm$360Hz with an RMS output of 2 Kw has been installed on the Georgia Tech campus, and a receiving site established at Technology Park/Atlanta, 27 Km northeast of the campus. The Georgia Tech ranging system uses phase comparison of the sidebands of a double sideband suppressed carrier signal to determine the echo range. For direction finding, relative phases between receiver outputs are determined by integration over 3/2 cycle of Doppler waveform, ensuring that each echo accepted meets the requirements placed on echo arrival angle. Data tapes at the receiving site contain information only on the Doppler frequency and relative phases, along with a continuously updated record of year, day of year, hour, minute and second. A series of FORTRAN computer programs handles the data reduction from determination of echo arrival angle, range, trail drift velocity and time of occurrence of echo to the plotting of height/time wind profiles and altitude dependent wind spectra deduced from the echo data. These programs have been documented by *Roper* (1975a).

B. *Decomposition of the Meteor Wind*

The Georgia Tech facility uses a technique developed by *Groves* (1959), that finds a "best fit" model to represent data by generalizing the least square solution. In this method, the zonal, meridional and vertical components of the model wind at the reflection point on the trail are assumed to be certain specified functions of height and time containing arbitrary parameters. These parameters are then chosen in such a way that the error between the model wind and the observed trail drift velocity is minimized.

Typically, a polynomial variation in height, with a periodic variation in time, is allowed. Third degree polynomials in height for zonal and meridional components, and a first (or zero) degree polynomial for the vertical component are found satisfactory for data that has at least 120 echoes per grouping. The fundamental period is taken as 24 hours with 12 and 8 hour components also computed. This choice is justified by the ample data on the presence of large tidal components (24- and 12-hour) at meteor heights.

The prevailing wind then, is the meteor wind averaged over a period of 5 to 10 days from which tidal components (24, 12 and 8 hour periods) have been removed. The echo rate dictates the period of averaging beyond five days. Figure 6 is the height-time plot of the zonal and meridional components of the prevailing wind for the period of August 1974 through August, 1975. There are no data available over the period of March 28-May 14, 1975. In this figure, isotachs are plotted at 10 m/sec intervals. The continuous lines represent westerly and southerly winds (taken positive) on the zonal and meridional plots respectively. The computed error (one standard deviation; not shown) over the entire period under consideration is season dependent. It is large during winter due to large variability of the wind at meteor heights during winter. It averages to about 20 per cent for a reasonable echo rate (i.e. about 100 usable echoes per day). The echoes are sampled over the height range of 76-106 km, though the winds below 80 km and above 100 km are not considered due to large errors involved (usable echoes outside 80-100 km height range are very small in number). The height resolution is 2 km in these experiments.

4. Results and Discussion

A. *Prevailing Wind, August 1974 through July 1975*

Zonal component. The zonal circulation is characterized by mostly easterly wind during fall through mid-winter. Westerlies are present during late winter and throughout the summer months (Figure 6).

The slope of the zero isotach during August 1974 indicates a downward progression of the westerly regime. The westerlies reach the 80 km level during the first week of September and appear to weaken while descending. A maximum westerly wind of more than 20 m/sec is reached above 95 km around the third week in August. From mid-September, easterlies re-establish beginning from lower levels, so that by the last week of October, the easterly wind takes over the entire 80-100 km height range. The easterlies gradually intensify until a maximum is reached by the fourth week of November 1974 near 95 km; a peak value of greater than - 30 m/sec is recorded.

Toward the end of November, easterlies start to weaken until a reversal begins in the third week of December 1974. The slope of the zero isotach, again, suggest a downward progression from 100 to 80 km.

The easterly wind re-establishes, starting at lower heights by the beginning of January 1975, and intensifies to a peak value of more than -20 m/sec above 90 km during mid-January.

Subsequent reversal to a westerly regime during the last week in January 1975 appears to have started from higher levels. Throughout February and March 1975, the zonal circulation is such that the westerly wind strengthens with increasing height, while below 85 km, a weak and variable wind regime persists.

There is no data for the period of April through mid-May 1975. From mid-May, where the 10 m/sec isotach is seen almost vertical, the wind is westerly through the summer and it strengthens with increasing height.

The Meridional Component. Compared to the E-W component, the north-south wind is weak and variable in strength.

The zero isotach located near 85 km divides the northerly wind, above, from the southerly wind, below, during August through mid-October 1974. During November, the zero isotach shows a progressive establishment of southerly wind over the entire 80-100 km height range, starting at 80-km. The southerly wind persists through the last week in February 1975, with a few reversals to northerly wind.

During March 1975, the wind is northerly and around the last week of March, there is a strong reversal ($u > 10$ m/sec) to southerly wind above 90 km.

During the middle of May 1975, a strong southerly current is observed below 90 km ($u > 20$ m/sec) with northerly wind above. During

June, the meridional wind is northerly, followed by southerly wind in July. From the third week of July, a reversal to wind from the north begins at higher altitudes.

The Mean Meridional Temperature Gradients. Vertical shear in the geostrophic wind implies the presence of a horizontal temperature gradient. The thermal wind, which is the vector difference between the geostrophic wind at two pressure levels, cannot strictly be applied in relation to the meteor wind between the 80-100 km altitude range. However, qualitative information about the mean meridional temperature gradients can be inferred from the height-time plot in Figure 6.

The thermal wind blows parallel to the isotherms, with warm air to the right facing downstream in the Northern Hemisphere. Then, for example, when westerly wind gains strength or easterly wind weakens with increasing altitude, positive thermal wind (zonal component) is indicated; this implies that the mean meridional temperature gradient is negative and thus, colder latitudes are indicated northward. If at the same time, the mean meridional wind component of the prevailing wind is southerly, we can infer warm advection.

Studying Figure 6 in this context, it can be seen that from mid-August 1974 through July 1975, the meridional temperature gradient is negative, implying cooler latitudes north of Atlanta, with two significant reversals. During the period of the second half of November 1974 and later on around mid-January 1975, warmer northern latitudes are indicated.

From the last week in November 1974 through the last week in February 1975, the meridional wind is southerly at 80-100 km. Thus, during this period warm advection is indicated except around mid-January where there is cold advection at meteor heights.

B. The Circulation Changes in the Meteor Zone Related to the Stratospheric Circulation System

In this section the meridional and zonal components of the prevailing wind at meteor heights are analyzed in relation to the pressure and temperature regimes in the upper stratosphere. The period under consideration is from August 1974 through July 1975 (Figure 6). Major emphasis is on the minor stratwarm in the mid-winter of 1974-75. The effect of quasi-periodic warming pulses felt in the winter stratosphere is also studied.

In Figure 7 are plotted temperature traces on Atlanta longitude (84°W) at 35°N and 50°N, at 0.4- and 5-mb levels, the temperature values being read from the synoptic charts produced by the Upper Air Branch of the National Meteorological Center (*Staff*, 1978). The values are approximate and limited significance can be attached to the 35°N to 50°N temperature gradients read from this figure. It is also necessary to keep in mind that the thermal systems in the atmosphere generally change in intensity and position with changes in altitude.

The Effect of a Strong Warming Pulse Felt During November of 1974.

The upper air synoptic charts (*Staff*, 1978) indicate that during the second half of November 1974, a warming pulse was strongly felt in the upper stratosphere. Polar latitudes at 0.4-mb (~55 km) are warmer by about 35°C on November 20, 1974 than on November 6, 1974. This can be seen from Figure 8, where temperature traces at 70°N latitude are plotted for November 6 and 20, and December 11, 1974 from 0.4-mb maps.

A warm cell has moved over the Asian land mass to a position 90°E within the Arctic Circle during November 1974. The 0.4-mb chart for November 20, 1974 (Figure 9) also shows two cold cells, one at 40°N off the China coast and another over Central Europe. There is a low pressure center over Scandinavia and a weak low is centered over the Northern North Pacific. At 5- and 2-mb levels (not shown), a weak high is found centered over North Alaska and a weak warm cell can be detected westward at the 2-mb level (*Staff*, 1978).

Figure 7 shows a warming of about 10°C during the third week of November 1974 at the 0.4-mb level for Atlanta's longitude. Also note the strong negative meridional temperature gradient (35°N to 50°N) at the 5-mb level (~36 km).

At meteor heights during the second half of November 1974 (Figure 6), the easterlies strengthen above 90 km. The meridional temperature gradient is positive, implying warmer northern latitudes. The meridional wind is from the North above 92 km. It then appears that we have a thermal structure indicated that will severely weaken the circulation associated with the stratospheric-lower mesospheric polar cyclone (i.e., westerly wind) that extends at least up to 100 km in stable winter conditions.

It is then probable that at middle and high latitudes, the warming pulse that was felt strongly near stratopause heights extended upward, possibly beyond the mesopause (80-85 km). This conclusion fits well with *Labitzke*'s (1972a,b) principle of compensating changes of temperature in the vertical, when we note that below 40 km (Figure 7) there is a tendency towards cooling in higher latitudes during the second half of November 1974.

The Effect of the Mid-Winter Warming of 1974-75. The reversal to westerly wind toward the end of December 1974 and return of easterlies in January 1975 leading to an intensification during the third week, are the principal zonal circulation features that appear to be directly linked to the various phases of the mid-winter warming (1974-75) of the stratosphere.

We have reviewed this warming in Section 2. As mentioned before, this was a major warming event in the upper stratosphere and as Figure 10 indicates, on January 1, 1975, at 5- and 2-mb levels (36 to 42 km) the Siberian Arctic has warmed by about 40°C. Its maximum descent was probably around January 5, 1975 as the radiance peak in Figure 4 would indicate.

The temperature traces for Atlanta longitude (Figure 7) toward the end of December 1974 show cooling at 35°N in the upper stratosphere. This compensating cooling in association with high latitude warming is in line with the latitudinal compensation reported by *Fritz and Soules* (1972).

What are the implications of this amplification stage of the stratwarm in relation to the circulation at meteor heights over Atlanta? From Figure 10 we notice that the warm stratopause has descended to about 35 km height in polar latitudes by the first week of January 1975. This situation, in principle, corresponds to Stage II of *Labitzke*'s (1972b) profile for "composite warming" (Figure 5). This profile leads to a very cold mesopause near 80 km height for higher latitudes; and to a warmer mesopause in tropical latitudes (Figure 5; profiles on the right).

At meteor heights we find westerlies, strengthening with height, during this period (a reversal from easterly wind). The meridional wind is southerly and more intense during the second half of December 1974 (Figure 6). The thermal wind is positive during December 1974, implying negative meridional temperature gradient over the entire height range, i.e., northern latitudes are cooler.

Thus, it can be safely concluded that the circulation changes at meteor heights over Atlanta during the period of late December-early January of 1974-75 correspond to the changes implied for these heights by the warming State II described before. The meridional temperature gradient at meteor heights is conducive to westerly circulation.

It is interesting to observe that the descent of westerlies from above 100 km during late December 1974 (Figure 6) appears comparable to the descent of easterlies (or warming; Figures 2 and 3) in the upper stratosphere, with the westerlies descending (implying gradual cooling) about a week (or more) earlier.

Now taking up the late phase of the stratwarm, the zonally averaged temperature values (Figure 3) indicate that the warming of higher latitudes is considerably reduced by January 15, 1975 and the low latitude warming is evident in the upper stratosphere. On Atlanta longitude, there is warming at 35°N and 50°N all through the 5- to 0.4-mb (36 to 55 km) height range. Also, by January 15, 1974, westerlies appear at all latitudes with a strong westerly jet centered near 40-45°N at stratopause heights (Figure 2). Figure 10 indicates warming with increasing height at 70°N, with a warm cell at 90°E.

This situation, in principle, corresponds to Stage III (Figure 5) of a major warming as projected by *Labitzke* (1972a,b). As remarked earlier, the 1974-75 mid-winter warming was a major stratwarm in all but one regard--that the circulation reversal to easterly wind did not take place in higher latitudes at 10-mb or below. We then expect that at mesopause altitude, there will be a positive meridional temperature gradient, with a warm mesopause at higher latitudes--favorable to anticyclonic circulation at those heights.

The observed changes in the zonal wind at meteor heights over Atlanta around mid-January 1975 (Figure 6) appear to support the preceding remarks. The easterlies reappear in Janaury 1975, ascending from below (i.e., 80 km), and intensify during the third week above 90 km (>20 m/sec). The thermal wind is negative, implying warmer northern latitudes. The meridional wind component has weakened in strength (southerly) through January 1975.

The establishment of westerly wind towards the end of January 1975 (Figure 6) from higher heights (i.e., from above 100 km) is the return of the normal circulation at meteor heights after the dissipation of the stratwarm. The cyclonic circulation in the upper stratosphere is well established (notice the strong negative temperature gradient on Atlanta longitude, Figure 7). The cold mesopause has made a return in higher latitudes as implied by the strong positive zonal component of the thermal wind at 80-100 km.

<u>The Summer of 1975</u>. The stratospheric circulation during summer is governed by the relatively stable circumpolar anticyclone with east-

erly wind over much of the Northern Hemisphere. At meteor heights over Atlanta however, the zonal winds follow the thermal balance. Westerlies are present at 80-100 km (June through early August) with weaker winds at lower heights. Positive thermal wind implies a colder higher latitude mesopause. The meridional wind is weak and averages over the summer as weakly northerly as suggested by various models as well as other observations (for example, *Leovy*, 1964).

C. *General Comments on the Prevailing Wind Structure*

It was observed earlier (section 4.A) that the prevailing wind (particularly the zonal component) generally exhibits a phase lag with increasing altitude. It is also observed (Figure 6) that the wind components are generally stronger at higher altitudes in the meteor zone. These observations suggest an upward propagation of wave energy from the atmosphere below into the region of decreasing density, at least up to 100 km.

Next, taking a summary view of the implied meridional gradients (section 4.A), it appears that a cold mesopause is indicated throughout the year at northern latitudes. The only indications to the contrary, during winter, have been related to the thermal perturbation of the high latitude stratosphere. In recent years *Quiroz* (1969), *Labitzke* (1972,b) and others have questioned the need of a heating requirement for the winter mesopause at high latitudes. Their speculation is that the winter mesopause may in fact be "cold" in early winter and again in the late winter, and the warm mesopause observed by rocket data (*Theon et al.*, 1972) is probably associated with the sudden warmings of the high latitude winter stratosphere. Present work is in line with this interpretation.

Finally, in this analysis the meteor wind has been interpreted in terms of the large scale circulation features of the stratospheric circulation system. In spite of the very local accumulation of the data, it is interesting to see how well the variation can be interpreted as part of the large scale circulation, and at such a relatively low latitude station at that.

5. Conclusions

In the present work, synoptic scale variations in the prevailing wind over Atlanta (34°N, 84°W) at 80-100 km are interpreted in the context of the stratospheric circulation system. The period covered is from August 1974 through July 1975. It has been shown that quasi-periodic warming pulses felt in the winter stratosphere can influence the wind regime in the meteor zone. The minor mid-winter stratwarm felt in the high latitude upper stratosphere during 1974-75 is studied and various phases of this warming have been shown to have affected the prevailing wind in the meteor zone over Atlanta in a consistent manner, by primarily making use of the latitudinal and vertical temperature compensation criteria in the stratosphere and above. What makes this analysis particularly interesting is the fact that the perturbed high latitude stratosphere interacts strongly with the mesopause level winds over a relatively low latitude station such as that at Atlanta (34°N).

The large southerly wind during winter and weak northerly wind during summer at 80-100 km altitude is consistent with other observations and mesospheric models (for example, by *Leovy*, 1964). It also appears that the implied meridional temperature gradients in the one year data point to a year round cold mesopause at high latitudes. The only exceptions during winter are shown to be related to the perturbed state of the stratospheric cyclone due to the warming episodes. Needless to add, data from more northward stations is required to confirm this interpretation.

Some wave energy input from below is indicated by the preliminary examination of the height-time variation of the prevailing wind structure. This inference along with a possible explanation of the presence of easterlies during the fall of 1974 (where westerlies are expected to be present) are the topics of a future publication. Finally, it is obvious from the present work that the radio meteor technique offers an invaluable source of continuous data at lower themospheric heights for synoptic and long term studies of the general circulation system of the stratosphere-mesosphere.

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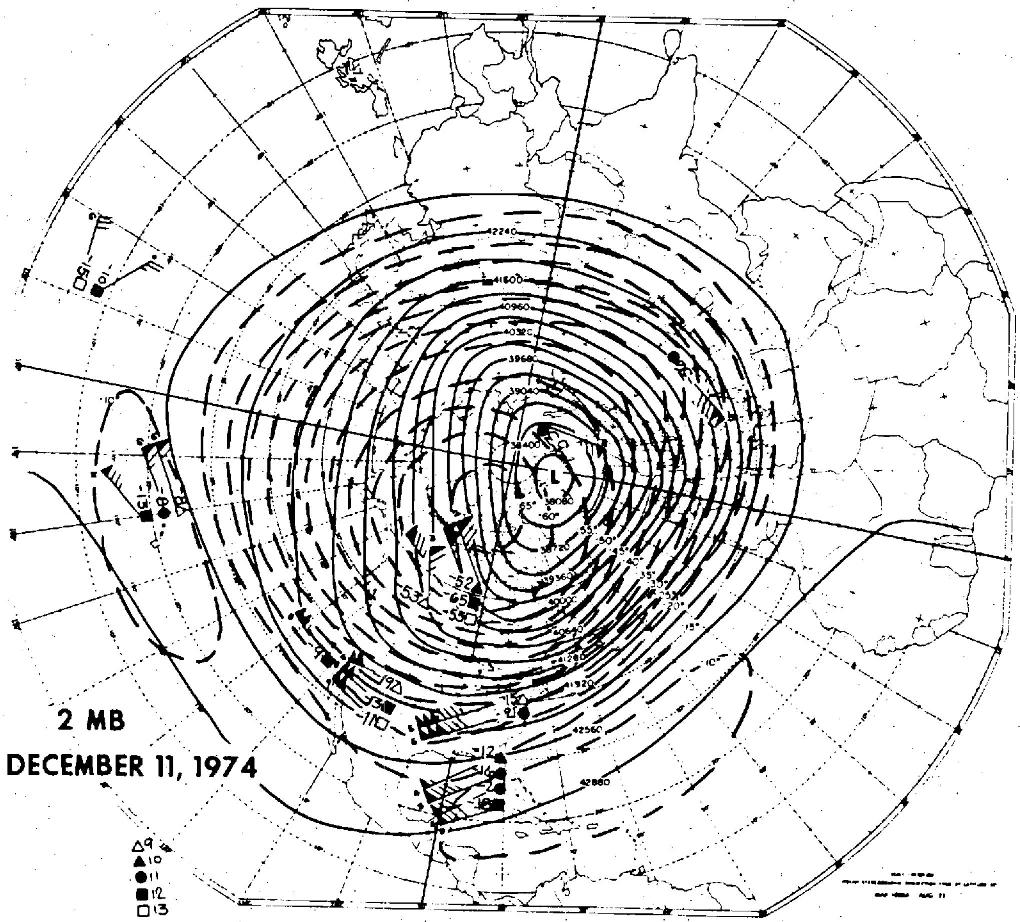

Figure 1a: 2-mb chart for December 11, 1974. Isobars are solid lines at intervals of 320 meters, and isotherms are dashed lines at intervals of 5°C (after *Staff*, 1977).

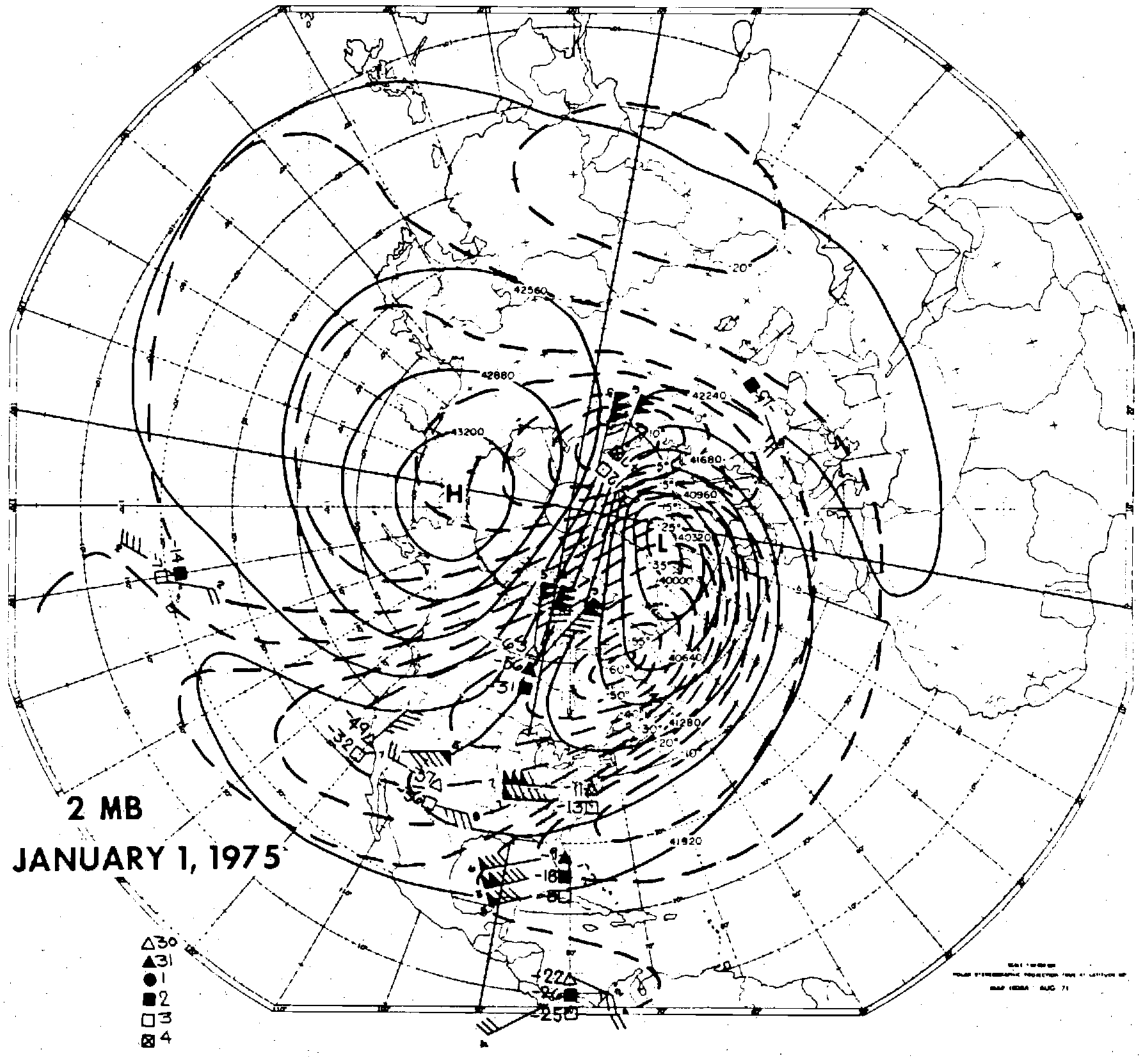

Figure 1b: 2-mb chart for January 1, 1975. Explanation as in Figure 1a.

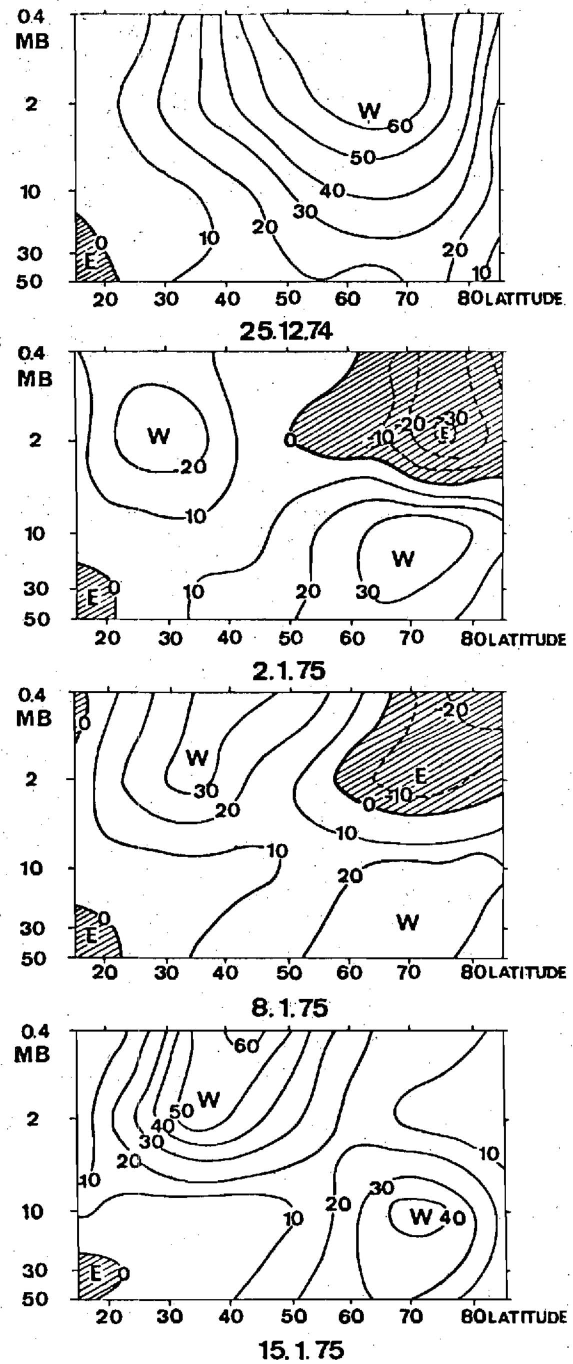

Figure 2: Mean zonal wind associated with the mid-winter warming of 1974-75 (after *Klinker*, 1976).

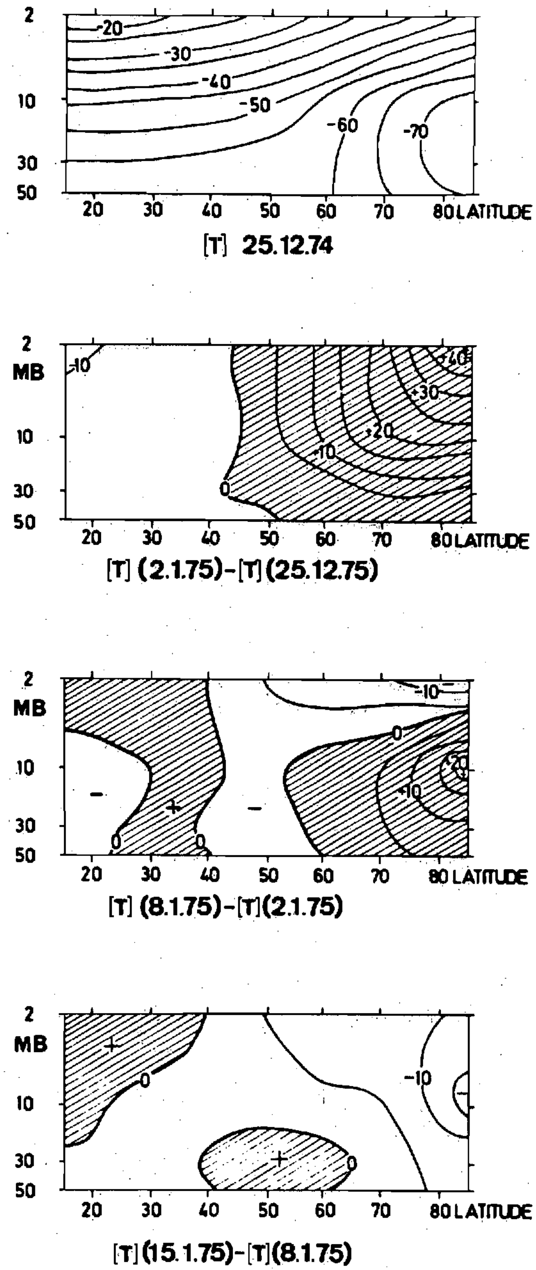

Figure 3: Mean zonal temperature associated with the mid-winter warming of 1974-75 (after *Klinker*, 1976).

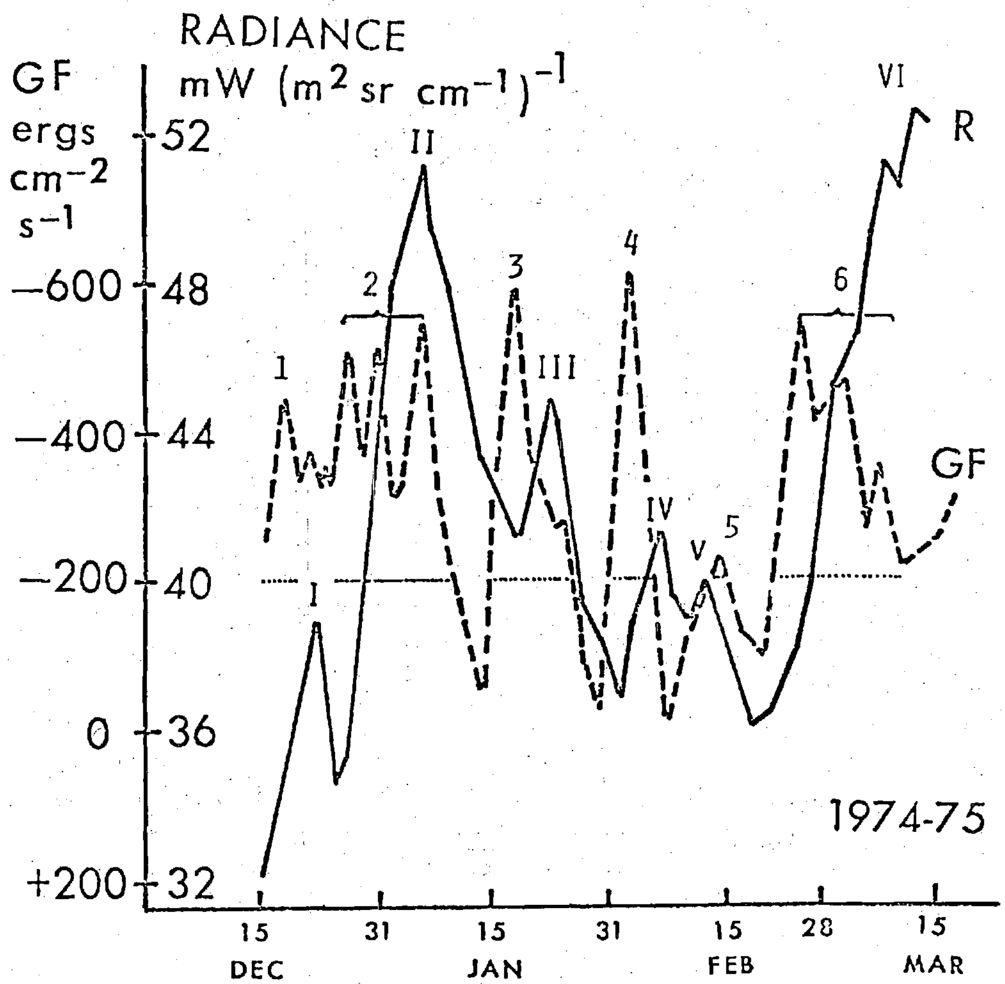

Figure 4: Eddy geopotential flux (GF) through the 100-mb surface integrated from 20°N to the North Pole (dashed line) and zonal mean radiance at 82.5°N based on VTPR (Vertical Temperature Profile Radiometer) Channel 2 measurements (solid line); 1974-75 (after *Quiroz et al.*, 1975).

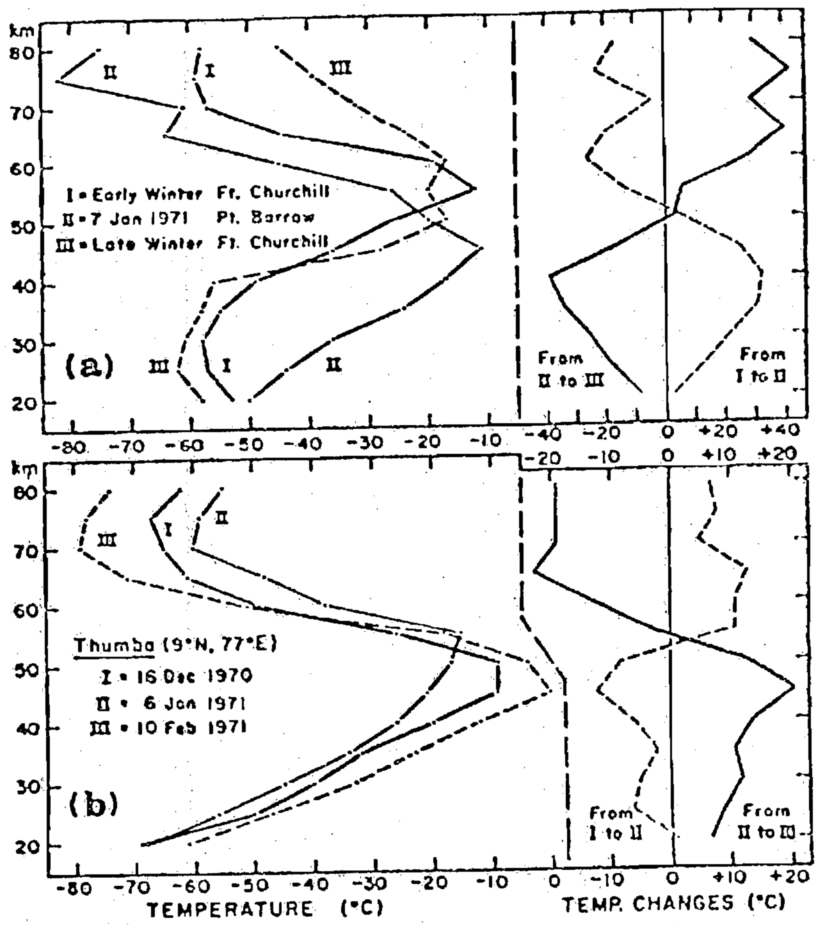

Figure 5: (a) Vertical temperature profiles at high latitudes for a composite mid-winter warming and also the changes between I and II and between II and III. (b) Vertical temperature profiles for Thumba, India. Dates as shown (after *Labitzke*, (1972b).

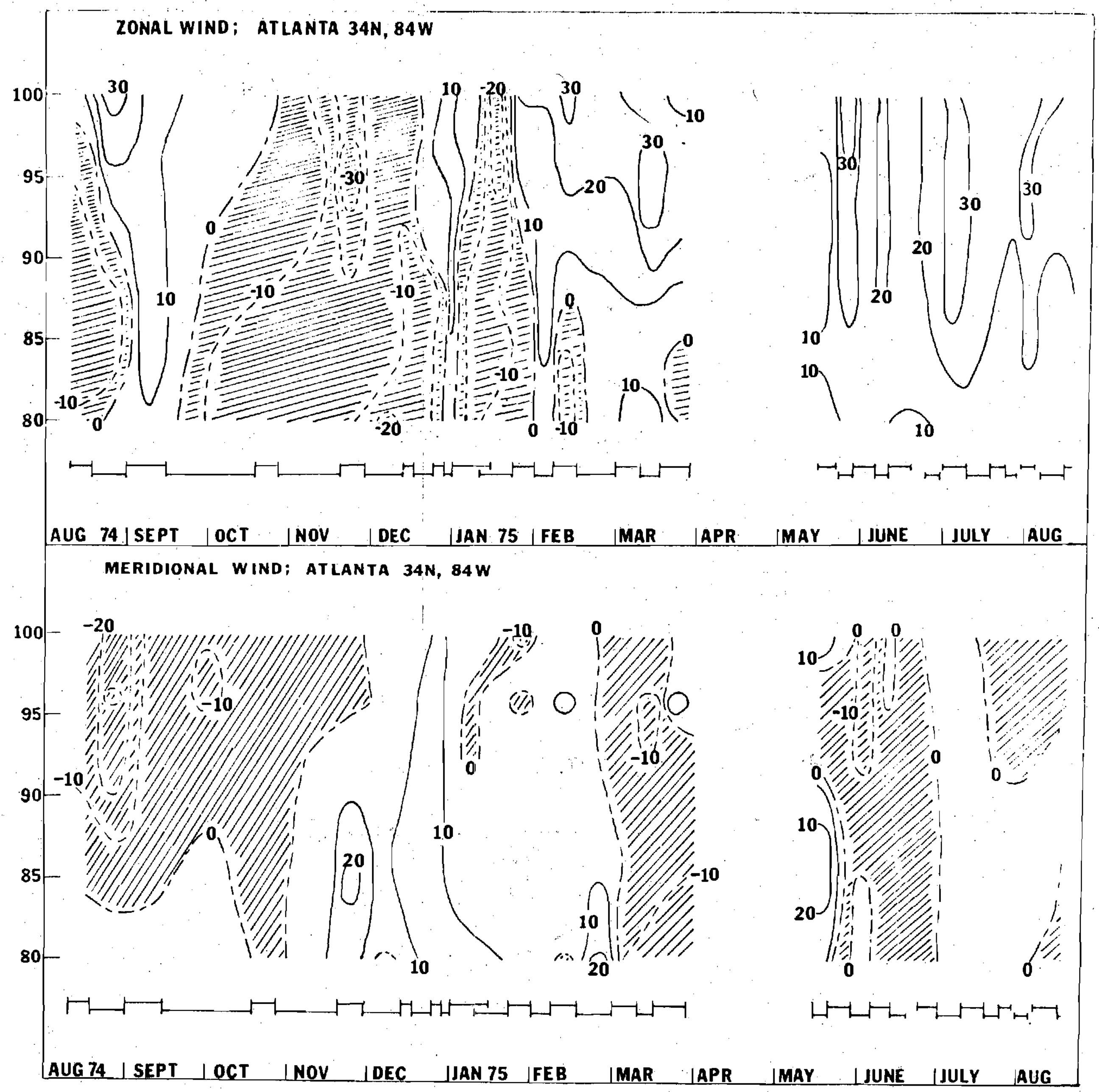

Figure 6: Height-time variation of the prevailing wind components: August 1974-August 1975. Wind speeds in m/sec.

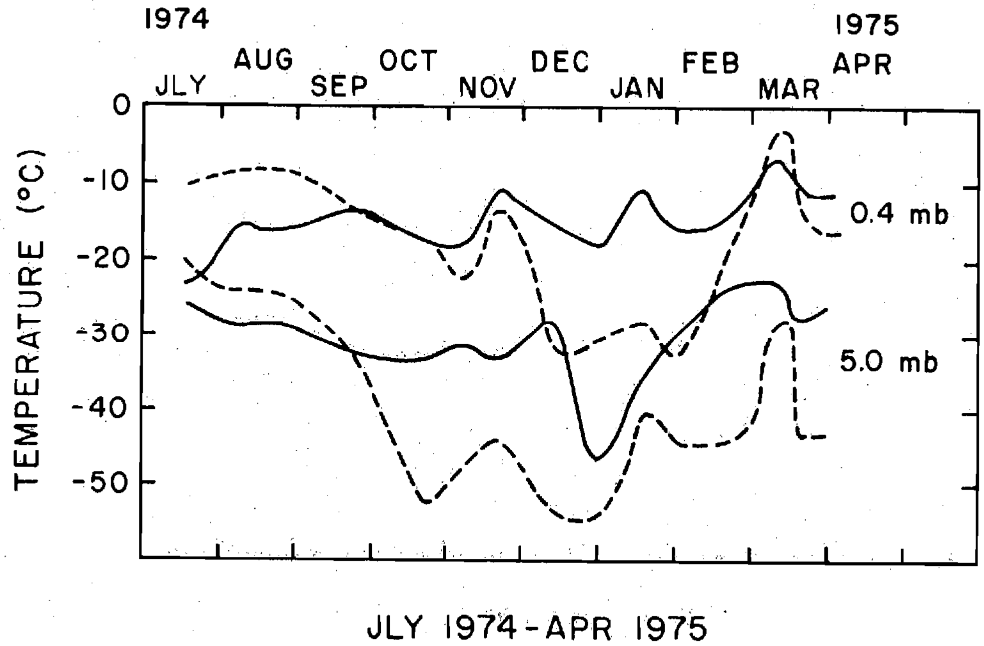

Figure 7: Temperature traces on Atlanta longitude (84°W), at 35°N (solid line) and 60°N (dashed line), at 5- and 0.4-mb levels: July '74-April '75.

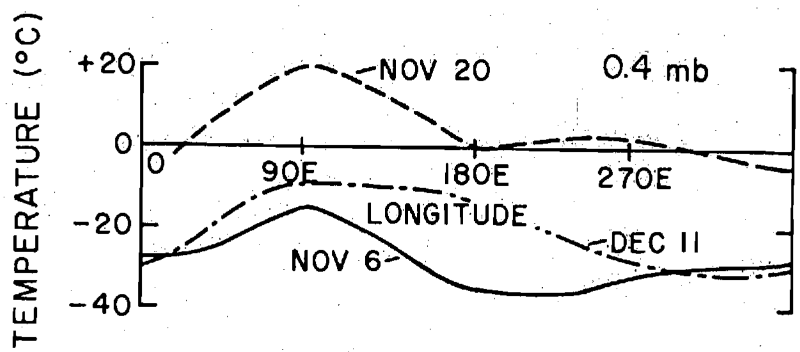

Figure 8: Temperature traces at 70°N latitude on 0.4-mb (~55 km) pressure surface during November-December, 1974.

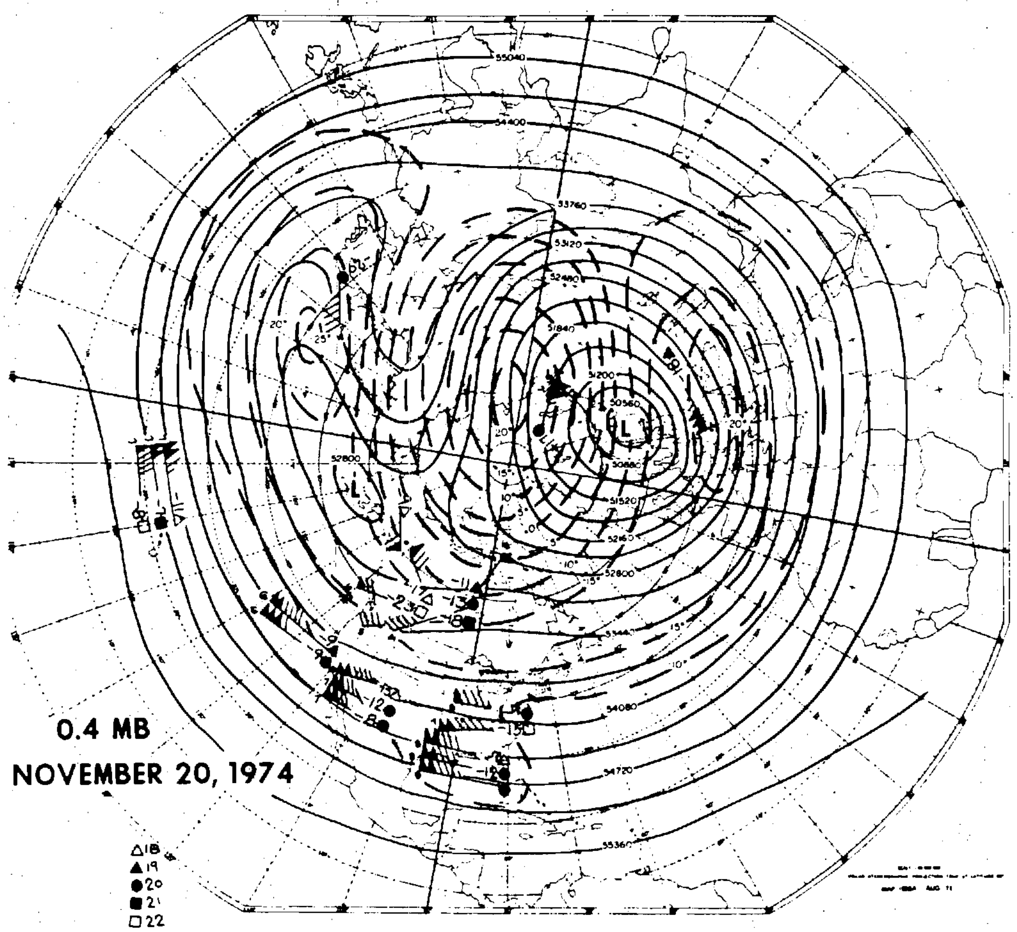

Figure 9: 0.4-mb map for November 20, 1974. Explanation is in Figure 1a.

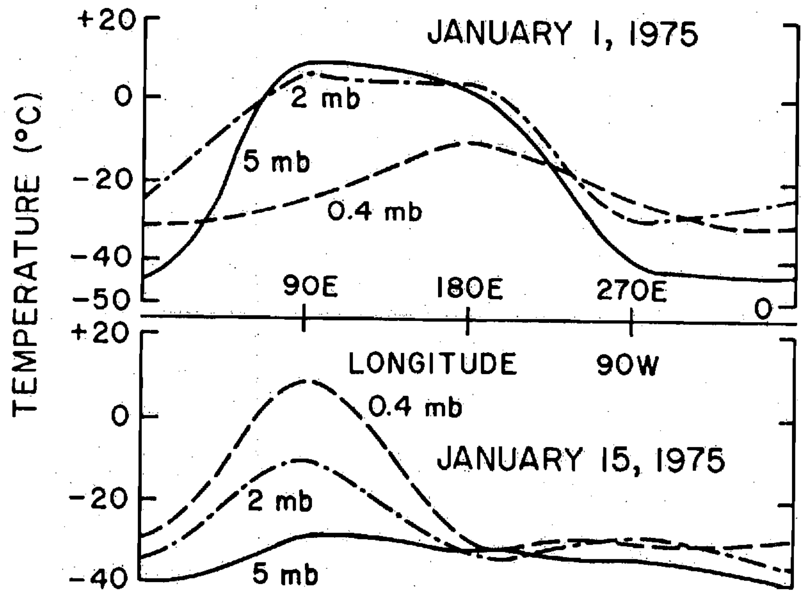

Figure 10: Temperature traces at 70°N latitude in the upper stratosphere on January 1 and 15, 1975.

CHAPTER VIII

MAP - The Middle Atmosphere Program

This report would not be complete without reference to the Middle Atmosphere Program of the Special Committee on Solar Terrestrial Physics (SCOSTEP) of the International Council of Scientific Unions. Details of this program, which is designed to be to the middle atmosphere what GARP is to the troposphere and the IMS is to the magnetosphere, may be found in the MAP Planning Document (available from Dr. S. A. Bowhill, Department of Electrical Engineering, University of Illinois 61801). A recent (1980) update on the status of MAP in the U.S. is available as "The Middle Atmosphere Program - Prospects for U.S. Participation" prepared by the Committee on Solar Terrestrial Research of the Geophysics Research Board, and published by the National Research Council, Washington, D.C. 20418.

While the major thrust of MAP has been delayed until 1982-85, pre-MAP programs are underway. Pre-MAP 1, which has as its focus a better understanding of the effect on the upper atmosphere of winter polar stratospheric warmings, was held during the northern hemisphere winter of 1978-79. A concerted effort to better understand the phenomenon of anomalous radio wave absorption in the D region is planned for 1980-81, with the West Germans spearheading an international rocket program to make in situ measurements in the mesosphere during the winter of 1980-81. Concurrent measurements of upper atmosphere dynamics using rocket released vapor trails and meteor wind radars have and will continue to provide vital knowledge of the dynamics of the upper mesosphere and lower thermosphere, a knowledge crucial to the understanding of these "anomalous" events.

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Publications produced under Grants NsG-304-63 and NGL-11-002-004

Evening Twilight Winds from 68 to 140 Kilometers for May 21, 1963, J. Geophys. Res., 68, 6062-6063 (C.G. Justus, H.E. Edwards and D.C. Kurts).

(Letter)

Photographic Instrumentation for Triangulation Studies of Luminous Clouds in the Upper Atmosphere, Appl. Optics, 3, 399-403, 1964 (M.M. Cooksey, Z. Frentress, H.D. Edwards).

ABSTRACT: *The position and velocity of clouds or other objects moving through the upper atmosphere are often determined from a study of simultaneous photographs taken from two or more observing stations against a darkened sky with a background of stars. With correction for atmospheric refraction supplied by tables, the star background is utilized to increase the accuracy of camera orientation by correcting errors inherent in the camera system produced by: (1) film shrinkage, (2) light refraction in the glass fiducial grid plate, and (3) tilt of the camera with respect to the local horizontal. Techniques and empirical formulae are developed for use in analytic data processing with a digital computer to the order of a thousandth of a centimeter. The final procedure can produce angular position determinations of 0.3 milliradian. A method for accurately determining camera focal lengths is also presented.*

A Method Employing Star Backgrounds for Improving the Accuracy of the Locations of Clouds or Objects in Space, Photogrammetric Engr., 3, 594-607, 1964 (C.G. Justus, H.D. Edwards and R.N. Fuller).

Instrumentation has been developed for obtaining triangulation photographs of luminous clouds in the upper atmosphere. These clouds are created by the release of chemicals such as sodium and cesium from rockets. The instrumentation has been used on approximately eighty rocket firings. Descriptions and operating characteristics are given of the equipment. Observations have been made from the Air Force ranges at Eglin Air Force Base, Florida, and Holloman Air Force Base, New Mexico, and from the NASA test facility at Wallops Island, Virginia. A modified version of the K-24 aerial camera with 7 in. (17.8 cm) *f*/2.5 optics was used and mounted on 60-in. (152-cm) searchlight carriages and Mk 51 gun directors. Camera control system, data chamber, fiducial light system, shutter and filter system, field installation, and alignment procedures are described.

Ionospheric Winds: Motion into Night and Sporadic E Correlations, Space Res. IV, 171-181, 1964 (N.W. Rosenberg, H.D. Edwards and J.W. Wright).

Abstract: The persistence of night winds and shears for more than 5 hours after sunset at altitudes between 100 and 150 km was evaluated during the COSPAR synoptic ionospheric winds program in December 1962. The method used was to release chemiluminous trails from rockets between 100 and 150 km with the times of release being 1720, 1801, 2145 and 2245 CST on 3 December 1962. The release times were respectively 13 minutes before sunset at the 100 km level and 28 minutes, 4 hours 12 minutes and 5 hours 12 minutes post sunset. All rockets were launched from the Air Force Gulf Test Range, Eglin Air Force Base, Florida at a longitude of 86.6° W and a latitude of 29.6° N.

It is believed that this is the first time such a morphology of winds and shears through the *E*-region has been made available. Comparisons are made between the winds and shears observed under night conditions with those for the preceeding twilight period.

Theoretical discussions by Whitehead *et al.* have suggested a wind shear origin for sporadic *E*. High east-west horizontal wind shears in the ionosphere and the natural magnetic field will act on ionized species to stratify these into relatively high vertical gradients.

The present paper reports on simultaneous measurements of ionospheric wind shears (measured by chemiluminescent trails) and of sporadic *E* (by sweep-frequency ionosondes). Occurrence at the same altitudes of maximum wind shears and strong sporadic *E* levels was found. In one case, for example (86.6° W, 29.6° N, 1720 CST, December 3, 1962) sporadic *E* was found at 96 ± 1 km to 1.0 Mc and at 109 ± 1 km at 6.2 Mc. At this time, wind shear maxima of 45 m/s km bearing 85° E at 97 ± 1 km and of 60 m/s km bearing 275° E at 109 ± 1 km were found.

These data tend to verify the correlation between wind shear and sporadic *E*.

A Triangulation Technique for Linear Objects in Space, Photogrammetric Engr., 31, 1020-1029, 1965 (H.P. Haney, W.M. Schofield and W.H. Wooten.

ABSTRACT: *Corresponding points on films, taken from two different sites, of a linear object in space cannot be paired due to unknown foreshortening. The usual triangulation techniques must be replaced by an iterative scheme to solve the problem. The particular methods developed in this paper are most useful when graphical checks of analysis are desired. Furthermore, the methods are tailored for easy computer programing and the closure technique of the iteration allows easy and quick convergence in pairing points. These methods have been used on chemical cloud releases in space to determine winds and wind shears.*

Space and Time Correlations of Ionospheric Winds, Radio Sci., 1, 149-155, 1966 (C.G. Justus and N.W. Rosenberg).

Ionospheric wind patterns between 90 and 150 km were studied by distortion of chemiluminous vapor trails released from rockets. Fifteen trails, closely paired in space or time, are reported. Profiles show well-organized sinusoidal patterns of near-constant wavelength, if the vertical coordinate is measured in integrated scale heights rather than kilometers. Dominant wavelength is 2-4 scale heights, similar in NS and EW components, and phase progresses downward about 1 wavelength per night. Horizontally spaced trails show vertical profile displacement corresponding to horizontal scales of 1000-4000 km in NS wind, longer in EW profiles. Viscous energy dissipation from observed waves is a significant heat source at 150 km altitude.

Energy Balance of Turbulence in the Upper Atmosphere, J. Geophys. Res., 71, 3767-3773, 1966 (C.G. Justus).

Turbulent winds obtained from chemical release studies and turbulent diffusion observations of globule expansion can be used to obtain estimates of the terms in the energy balance equation $\epsilon_s = \epsilon_g + \epsilon_d$, where ϵ_s is the rate per unit mass at which energy is supplied by wind shears, and ϵ_g and ϵ_d are the rates per unit mass at which energy is dissipated by buoyancy and viscous effects. Both ϵ_s and ϵ_g are found to vary slowly with altitude, having values of about 0.3 and 0.1 watt kg^{-1}, respectively, in the altitude range 90 to 110 km. However, ϵ_d increases rapidly with altitude. This rapid increase in ϵ_d is responsible for the globule cutoff observed at an average altitude of 106 km. Globule growth analysis indicates that extremely rapid anomalous diffusion takes place during the period approximately 100 to 150 seconds after release. As globule diameters increase beyond the largest scale of the anomalous growth region they expand at a rate comparable to that of molecular diffusion for a period of approximately 50 seconds, after which a transition to $d^2 \sim \epsilon_d t^3$ turbulent diffusion occurs.

Dissipation of Wind Energy in the Height Range 80 to 140 Km, J. Geophys. Res., 71, 4427-4428, 1966 (R.G. Roper).

(Letter)

Atmospheric Turbulence in the Meteor Region, J. Geophys. Res., 71, 5758-5792, 1966 (R.G. Roper).

In 1961 at Adelaide the atmospheric turbulence near 90 km was measured month by month using a spaced-station radio meteor technique. The characteristics of the large-scale turbulent motions are found to be similar to those observed in the northern hemisphere. Characteristic velocities range from 23 to 40 m/sec and the large eddies are distinctly anisotropic; they have a horizontal scale of from 50 to 250 km and a vertical scale of 7 km, in keeping with a gravity wave source of turbulent energy. Shears measured over separations of 0.5 to 3.5 km are found to be characteristic of an apparently inertial spectrum, and the turbulent dissipation rate calculated from these shears shows a marked seasonal variation, with a maximum of 400 ergs/g sec in autumn and spring and a minimum of 160 ergs/g sec in summer and winter. There is a strong correlation between the seasonal variation of the turbulent dissipation rate and the energy of the 24-hour component of the mean wind. The height shear is of particular interest, as, in keeping with rocket vapor trail wind analysis, it follows an approximate 1.4 power law.

The Eddy Diffusivities, Energy Balance Parameters, and Heating Rate of Upper Atmospheric Turbulence, J. Geophys. Res., 72, 1035-1039, 1967 (C.G. Justus).

Wind profile and turbulent wind data obtained from photographic tracking of rocket released chemical clouds have been used to compute the thermal and momentum eddy diffusivities K_t and K_m in the 90–110-km region. Exponential functions having a small increase with altitude provide a reasonable representation of the mean K_t and K_m values. The ratio $K_m:K_t$, the turbulent Prandtl number, is found to be approximately constant with a value of about 3. The mean K_t at 100 km is about 180 m^2/sec, in reasonable agreement with previous estimates. The exponential functions for K_t and K_m are also used to obtain revised estimates of the energy balance parameters and the turbulence heating rate in this altitude range.

The Spectrum and Scale of Upper Atmospheric Turbulence, J. Geophys. Res., 72, 1933-1940, 1967 (C.G. Justus).

Turbulent winds determined by photographic tracking of chemical release clouds are used to determine the turbulence structure function in the 90–110-km region. The observations indicate that the turbulence structure function is approximately isotropic with a large scale of about 5 km. The turbulence structure function is found to vary, as $r^{2/3}$ in the 1- to 5-km scale range. However, there is insufficient observational data and theoretical background to determine if this is a true inertial scale range or is a 'pseudo-inertial' region in which wind shear production and buoyancy loss are approximately balanced, leading to no net loss from the spectrum. Structure functions of the total wind profile are also discussed and related to a possible form for the gravity wave spectrum.

The Velocity Probability Density of Upper Atmospheric Turbulence, J. Geophys. Res., 72, 2460-2462, 1967 (C.G. Justus and W.B. Moseley).

(Letter)

Turbulence in the Lower Thermosphere, with W. G. Elford, Space Res. VII, 42-54, 1967 (R.G. Roper).

Abstract. Results of studies of wind shears in the lower thermosphere (80-120 km) by means of observations of chemiluminescent trails and meteor trails are summarized. The motion is highly anisotropic - in the height range 80-100 km typical scales are 200 km horizontal and 5-10 km vertical. The horizontal motions exhibit the characteristics of an inertial subrange of eddies.

The rate of dissipation of turbulent energy increases exponentially over the height range 80-105 km, with an average value of 3×10^{-2} W/kg at 93 km. The equivalent heating rates of the atmosphere over this same height range are 1-5°K per day. Turbulence decreases rapidly above 105 km.

Below 100 km the meteor observations indicate that there is a strong seasonal variation in the rate of dissipation of turbulent energy; at 93 km a minimum value of 1.5×10^{-2} W/kg occurs in summer and winter, and a maximum value of 3.5×10^{-2} W/kg during spring and autumn.

Reply (to Discussion of a Paper by C.G. Justus "Energy Balance of Turbulence in the Upper Atmosphere"), J. Geophys. Res., 73, 455-458, 1968 (C.G. Justus).

(Letter)

Atmospheric Tides in the Height Region 90 to 120 Kilometers, J. Geophys. Res., 73, 467-478, 1968 (C.G. Justus and A. Woodrum).

Winds determined from chemical release trails in the 90- to 120-km region (primarily from the early and middle fall season) are analyzed by a method that is an extension of the method used by Hines for the determination of the diurnal tide. This method is designed to reveal the prevailing, diurnal, and semidiurnal components. A diurnal tide is computed that has a vertical wavelength (19 ± 1 km) in agreement with the measurements of Hines and with tidal theory predictions for the (1, 3) mode. The observed height variation of the prevailing wind can most easily be explained in terms of a wave whose vertical wavelength is 27 ± 5 km. The small observed magnitude of the semidiurnal tide indicates that there is considerable seasonal variation of this component, since a large semidiurnal tide has previously been observed during periods near the summer and winter solstices. Observed phase differences between the northward and eastward components of the prevailing and tidal winds are discussed. Some suggestions for improvements in the tidal extraction analysis method are also given.

Some Observations of Turbulence in the 90 to 110 Km Region of the Upper Atmosphere, AMS Meteoro. Mono., 31, 122-128, 1968 (C.G. Justus and R.G. Roper).

ABSTRACT

Some recent measurements of turbulence in the 80-110 km region obtained by both the radio-meteor and chemical release methods are summarized. The velocity probability density, the structure function, eddy diffusivities, heating rate, and the altitude and seasonal variation of the energy balance parameters of the turbulence are discussed. Some aspects of the apparent disagreement between meteor trail and chemical release measurements are presented, and a proposed qualitative model is offered for the behavior of turbulence in this region of the atmosphere and for the interpretation of the experimental results. This model ascribes the chemical release wind fluctuations and the small-scale meteor wind differences to turbulence, but attributes the large-scale meteor wind fluctuations to gravity wave action. Dissipation of tidal and gravity wave energy is proposed as a source of the turbulence energy, and some aspects of the wave-to-turbulence transfer are discussed.

Measurement of the Magnitude of the Irregular Winds in the Altitude Region Near 100 Kilometers, J. Geophys. Res., 73, 7535-7537, 1968 (C.G. Justus and A. Woodrum).

(Letter)

Measurements of Tidal Period Winds in the 95 to 135 Km Region, J. Geophys. Res., 74(16), 4099-4104, 1969 (C.G. Justus, A. Woodrum and R.G. Roper).

A method developed by G. V. Groves for the analysis of meteor winds has been used on chemical release wind data from rockets. Velocity amplitudes, phases, and amplitude errors for winds with periods of 8, 12, and 24 hours were computed in the 95- to 135-km region. An average wind, assumed to represent a constant prevailing wind was also computed in this altitude range. Vertical wavelengths and phase differences of the northward and eastward components of the prevailing and tidal winds are calculated and compared with recent results. Dissipative forces are also shown to be acting on the propagating diurnal and semidiurnal tides in the height range from 95 to 135 km.

Dissipation and Diffusion by Turbulence and Irregular Winds Near 100 Km, J. Atmos. Sci., 26(5), 1137-1141, 1969 (C.G. Justus).

ABSTRACT

A review is given of some recent measurements of the following energy balance parameters in the 90–110 km height region: ϵ_0, the viscous dissipation of kinetic energy by shears in the mean winds; ϵ_d, the viscous dissipation of kinetic energy by shears in the turbulent winds; ϵ_s, the transfer of kinetic energy from the mean motion to the turbulent motion; ϵ_p, the transfer of turbulent kinetic energy to potential energy by buoyancy action; and ϵ_w, the rate of dissipation of wave energy of tides and irregular winds interpreted as gravity waves. Some measurements of the growth rates of globular structure on chemical release clouds and the growth rate of interglobular distances are presented. These data indicate that the diffusion mechanism is a mixture of the $\langle v^2\rangle t^2$ variation expected for the diffusion of a point from its initial position with the $\epsilon_d t^3$ variation expected for the variation of the separation between pairs of points.

A Theory for the Energy Spectrum of Shear-Dependent Turbulence, J. Atmos. Sci., 26(6), 1238-1244, 1969 (C.G. Justus).

ABSTRACT

A theory for the three-dimensional energy spectrum of nearly isotropic shear-dependent turbulence is presented. This theory is based on the author's general model for shear-dependent turbulence and a modified Pao spectral transfer theory which accounts for the effects of viscous loss, shear production, and inertial and velocity gradient transfer in nonstationary turbulence. The resultant spectral transfer equation is solved in closed form for the case of complete similarity. The small wavenumber stationary energy spectrum is shown to vary as k^4. The small wavenumber nonstationary solution varies as k^θ, where θ can have values from 1–4. One-dimensional spectra are computed numerically and compared with the turbulence spectra observed for wind tunnel turbulence with a wide range of Reynolds numbers. The theory successfully predicts several features of the observed spectra at both high and low wavenumbers as well as in the inertial region.

Distribution and Structure of Irregular Winds Near 100 Km, J. Geophys. Res., 75, 2171-2178, 1970 (C.G. Justus).

The velocity distribution and structure function of irregular winds near the 100-km level are presented. The irregular winds were measured by differencing chemical release wind profiles determined at times different by an integral multiple of 24 hours and by computing the deviations from the 3-day mean of winds at the same time of day. There is a significant variation of the rms irregular winds between the 90- and 115-km limits over which they were measured. The magnitudes and variations with altitude are found to be reasonably similar for data obtained from Eglin AFB, Florida; Yuma, Arizona; and Barbados, West Indies. The distributions of irregular winds seem to depart slightly from a Gaussian distribution. In particular, a low value of the flatness factor has been measured consistently. The altitude and time structure functions of the irregular winds indicate a characteristic vertical wavelength of about 16 km and time scales of up to 20 hours. The observed structure functions over small scales (2-3 km for the vertical structure function, about 2 hours for the time structure function) follow a power law variation with an exponent value near 3/2. This may result from an energy spectrum of irregular winds that varies as the −5/2 power of wave number and frequency.

Response of Winds in the 90 to 140 Km Altitude Region to Variations in Solar Activity, J. Geophys. Res., 75, 5565-5570, 1970 (C.G. Justus and J.E. Hicks).

Winds determined from 52 photographically tracked trails of rocket-borne chemical releases in the 90- to 140-km altitude range, launched between October of 1962 and June of 1966, are used in both a correlation analysis and a linear regression analysis with several solar and geomagnetic disturbance indices. Results indicate that the altitude interval considered may be broken up into two distinctly different regions of solar influence: the regions above and below 110 km. Significant increases in the geomagnetic index Kp over a 15-hour interval centered on a time lag of 21 hours after increases in atmospheric circulation below 110 km suggest either a dynamo type of interaction or an increase in the geomagnetic field strength, produced by corpuscular radiation which lags behind the electromagnetic radiation and accompanying atmospheric circulation variations. Apparently continuous associations, with indicated periods of about one solar rotation, between both the solar radio noise index $F_{10.7}$ and the Zurich sunspot number R_z and the northward wind component magnitude above 110 km indicate that long-term variations in the winds in this region are related to similar variations in solar EUV radiation.

Thermospheric Observations Combining Chemical Seeding and Ground Based Techniquest, Planet. Space Sci., 20, 761-789, 1972 (D. Rees, K.H. Lloyd, C.H. Low, B.J. McAvaney, and R.G. Roper).

Abstract—Two Skylark sounding rockets carrying chemical seeding payloads were launched from Woomera, South Australia (31°S, 137°E) in October 1969. In conjunction with these firings, the University of Adelaide conducted ground-based experiments on the upper atmosphere using the radio meteor and spaced receiver drift methods. This paper presents the measurements of properties of the neutral atmosphere above 90 km which were obtained from these experiments.

Upper Atmospheric Planetary Wave and Gravity Wave Observations, J. Atmos. Sci., 30, 1267-1275, 1973 (C.G. Justus and A. Woodrum).

ABSTRACT

Previously collected data on atmospheric pressure, density, temperature and winds between 25 and 200 km from sources including Meteorological Rocket Network data, ROBIN falling sphere data, grenade release and pitot tube data, meteor winds, chemical release winds, satellite data, and others were analyzed by a daily-difference method, and results on the magnitude of atmospheric perturbations interpreted as gravity waves and planetary waves are presented. Traveling planetary-wave contributions in the 25-85 km range were found to have a significant height and latitudinal variation. It was found that observed gravity-wave density perturbations and wind are related to one another in the manner predicted by gravity-wave theory. It was determined that, on the average, gravity-wave energy deposition or reflection occurs at all altitudes except the 55-75 km region of the mesosphere.

The Measurement of Meteor Winds Over Atlanta, Radio Sci., 10, 363-369, 1975 (R.G. Roper).

An all-sky, continuous wave radio meteor wind facility has been installed in Atlanta by the Georgia Institute of Technology under National Science Foundation sponsorship. A double sideband suppressed carrier CW transmitter, operating on 32.5 MHz $\pm$ 360 Hz, with an rms output of 2 kw, has been installed on the Georgia Tech campus, and a receiving site established at Technology Park/Atlanta, 27 km northeast of the campus. Details of the equipment, together with height/time profiles of mean wind circulation and tides between 80 and 100 km, measured from August 9 to September 4, 1974, are presented.

A Global Reference Atmospheric Model for Surface to Orbital Altitudes, J. Appl. Meteorol., 15, 3-9, 1976 (C.G. Justus, A. Woodrum, R.G. Roper, and O.E. Smith).

ABSTRACT

An empirical atmospheric model has been developed which generates values for pressure, density, temperature and winds from surface levels to orbital altitudes. The output parameters consist of components for: 1) latitude, longitude, and altitude dependent monthly means; 2) quasibiennial oscillations; and 3) random perturbations to partially simulate the variability due to synoptic, diurnal, planetary wave and gravity wave variations. The monthly mean models consist of: (i) NASA's four dimensional worldwide model, developed by Environmental Research and Technology, for height, latitude, and longitude dependent monthly means from the surface to 25 km; and (ii) a newly developed latitude-longitude dependent model which is an extension of the Groves latitude dependent model for the region between 25 and 90 km. The Jacchia 1970 model is used above 90 km and is faired with the modified Groves values between 90 and 115 km. Quasibiennial and random variation perturbations are computed from parameters determined from various empirical studies, and are added to the monthly mean values. This model has been developed as a computer program which can be used to generate altitude profiles of atmospheric variables for any month at any desired location, or to evaluate atmospheric parameters along any simulated trajectory through the atmosphere. Various applications of the model are discussed, and results are presented which show that good simulation of the thermodynamic and circulation characteristics of the atmosphere can be achieved with the model.

Turbulence in the Lower Thermosphere, Chapter 7 of "The Upper Atmosphere and Magnetosphere", F.S. Johnson, ed., one of the Studies in Geophysics pub. by the National Academy of Sciences, Washington, D.C., 1977 (R.G. Roper).

(See Chapter 2)

Winds from the Atlanta (34°N, 84°W) Radio Meteor Facility, J. Atmos. Terrest. Phys., 40, 891-894, 1978 (R.G. Roper).

Abstract—A brief description of the Georgia Tech radio meteor wind facility is followed by a tabular presentation and discussion of winds measured over Atlanta (34°N, 84°W) for the first three intervals of the URSI/IAGA Cooperative Tidal Observations Program (CTOP). The pervailing zonal wind measured during August 1974, being easterly, is significantly different from that measured during October 1975 and January 1976, and is not typical of winds measured in August 1975 and August 1976, when westerlies predominated. The complicated tidal picture is detailed, but is not easily summarized.

A Comparison Between Radio Meteor and Airglow Winds, J. Geomag. Geoelectr., 31, 419-426, 1979 (G. Hernandez and R.G. Roper).

A comparison between the winds near 97 kilometers altitude has been made from observations of the 17924K (5577Å) OI line emission at Fritz Peak Observatory (39.8N, 195.5W) and with a meteor radar facility at Atlanta (34N, 84W), from August 1974 to November 1975. Since the optical emission measurements are made only at night, the nighttime meteor radar measurements have been used, weighted by an airglow emission rate profile. The results show general agreement in both the zonal and meridional wind vectors, but with the variations in the amplitude of the meridional winds at the northernmost station (Fritz Peak Observatory) larger than those at Atlanta, a result of the smoothing inherent in producing the meteor winds.