Unclassified
Security Classification

AD-748 540

DOCUMENT CONTROL DATA - R&D

(Security classification of title, body of abstract and indexing annotation must be entered when the overall report is classified)

1. ORIGINATING ACTIVITY (Corporate author) Air Force Cambridge Research Laboratories (LKB) L.G. Hanscom Field Bedford, Massachusetts 01730	2a. REPORT SECURITY CLASSIFICATION Unclassified 2b. GROUP

3. REPORT TITLE

THE MEASUREMENT OF ATMOSPHERIC STABILITY FROM 30 TO ~90 KM

4. DESCRIPTIVE NOTES (Type of report and inclusive dates)

Scientific. Interim.

5. AUTHOR(S) (First name, middle initial, last name)

S.P. Zimmerman
A.C. Faire
E.A. Murphy

6. REPORT DATE 28 August 1972	7a. TOTAL NO. OF PAGES 8	7b. NO. OF REFS 8

8a. CONTRACT OR GRANT NO.	9a. ORIGINATOR'S REPORT NUMBER(S) AFCRL-72-0489
b. PROJECT, TASK, WORK UNIT NOS. 8605-10-01	
c. DOD ELEMENT 61102F	9b. OTHER REPORT NO(S) (Any other numbers that may be assigned this report)
d. DOD SUBELEMENT 681300	

10. DISTRIBUTION STATEMENT

Approved for public release; distribution unlimited.

11. SUPPLEMENTARY NOTES Reprinted from Space Research XII, Akademie-Verlag, Berlin 1972.	12. SPONSORING MILITARY ACTIVITY Air Force Cambridge Research Laboratories (LKB) L.G. Hanscom Field Bedford, Massachusetts 01730

13. ABSTRACT

Three years of wind and temperature data determined from grenade experiments at six sites are utilized in the calculation of the Richardson number Ri. The results show that, for a critical Richardson number of 1, there is quite often a turbulent region around 85 km at all latitudes. Furthermore, the winter polar mesospheric data indicate a more intense turbulence than do the summer data. There also appears a region of marginal stability (Ri~1) at or slightly above the stratopause which is more often present during the winter season than in the summer. Latitudinal and seasonal distributions of Richardson number, wind shears and estimates of heating rates due to viscous dissipation of turbulent kinetic energy are given.

KEYWORDS: Richardson number, Atmospheric stability, Turbulence, Internal gravity waves, Viscous deposition

9

DD FORM 1 NOV 65 1473

Unclassified
Security Classification

Space Research XII Akademie-Verlag, Berlin 1972

THE MEASUREMENT OF ATMOSPHERIC STABILITY FROM 30 TO ~90 KM

S. P. Zimmerman, A. C. Faire and E. A. Murphy

Air Force Cambridge Research Laboratories, Bedford, Mass., USA

Three years of wind and temperature data determined from grenade experiments at six sites are utilized in the calculation of the Richardson number Ri. The results show that, for a critical Richardson number of 1, there is quite often a turbulent region around 85 km at all latitudes. Furthermore, the winter polar mesospheric data indicate a more intense turbulence than do the summer data. There also appears a region of marginal stability ($Ri \sim 1$) at or slightly above the stratopause which is more often present during the winter season than in the summer. Latitudinal and seasonal distributions of Richardson number, wind shears and estimates of heating rates due to viscous dissipation of turbulent kinetic energy are given.

1. Theory and Analysis

In a stably stratified atmosphere, the necessary condition for the maintenance of turbulence against buoyancy forces is that the flux Richardson number R_f be less than some critical number $R_f < R_{f_c}$, where R_{f_c} is the critical flux Richardson number. If this flux Richardson number cannot be measured since it requires the *in situ* simultaneous measurement of the turbulent transport coefficients of heat and momentum as well as the local atmospheric temperature and wind shear, then a measure of stability or instability may be provided by examination of the ordinary Richardson number Ri defined as

$$Ri = g[\partial T/\partial Z + \Gamma]/T[\partial U/\partial Z]^2 \tag{1}$$

where g is the acceleration of gravity, T is the atmospheric temperature (°K), Γ is the adiabatic lapse rate ($\simeq 10$ km^{-1}), Z is the vertical coordinate, and U is the amplitude of the mean horizontal wind velocity, such that $\partial U/\partial Z$ is the vertical wind shear, defined by

$$[\partial U/\partial Z]^2 = [\partial U_x/\partial Z]^2 + [\partial U_y/\partial Z]^2 ,$$

and where x and y are the horizontal directions.

The amplitude of the critical Richardson number Ri_c necessary for the onset of turbulence in a stably stratified atmosphere has been theoretically shown to be $Ri_c \leq 1/4$, and has been experimentally verified. Woods [1] has shown that when a region has been placed into a condition of fully developed turbulent flow, it may be maintained in this turbulent flow under the condition $Ri \leq 1$, and since for this study we do not have a time and space history of the temperature and wind fields, we shall assume that the observations where $Ri \leqq 1$ are indicative of local turbulence.

Thus given this condition for turbulence ($Ri \leqq 1$), let us now consider the turbulent energy balance equation, which may be expressed as

$$\varepsilon / K_{\mathrm{m}} = (1 - \alpha\, Ri)\,(\partial U / \partial Z)^2 \tag{2}$$

where ε is the rate of viscous dissipation of turbulent kinetic energy, K_{m} is the coefficient of turbulent momentum transfer, K_{h} is the coefficient of turbulent heat transfer, and $\alpha = K_{\mathrm{h}}/K_{\mathrm{m}}$, assumed unity here.

Utilizing the Heisenberg [2] relation for K_{m} and assuming an "inertial" sub-range over the turbulent energy spectrum, an expression for ε, valid within uncertainties of 2, may be derived using one-dimensional theory,

$$\varepsilon \approx 0.250\, l^2 \,[(1 - Ri)\,(\partial U/\partial Z)^2]^{3/2} \tag{3}$$

where l is the large scale end of the "inertial" spectrum. To limit the eddy wavelength to values smaller than the wavelengths of the wind shear, we shall use a scale size l of 500 m as the value of this large scale limit. This will give a wavelength of 3.14 km, which for the most part is small compared with the wind system vertical wavelengths of 5–20 km usually reported.

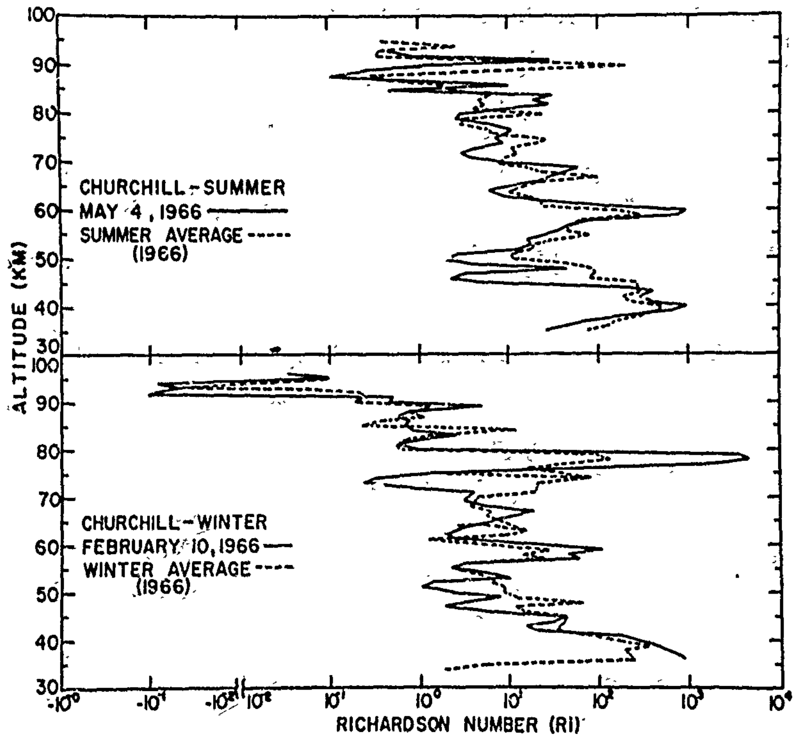

Fig. 1. The Churchill (60° N Lat.) winter and summer Richardson numbers. The average values are shown by the dashed line, and one immediate sample, as dated, is given by the solid line.

Fixing the large scale limit can introduce some error in the determination of ε, since by theory (Eq. (3)) ε and l are closely coupled. However, this method allows K_h, which is not calculated here, to vary as the rate of dissipation and thus, within the limit of this scale size, affords a more self-consistent determination of ε than would be accomplished by predetermining K_h: a rational procedure which would lead to serious inconsistencies due to the aforementioned interrelationship of ε, K_h and l. The errors generated by holding l constant would be to overestimate ε in regions of high shear, and thus lower Richardson numbers, and to underestimate ε in the region of Ri near unity, or low shear region. The occurrence average of ε and Ri, to be given later, should be considered conservatively because of the small data sample.

2. Results

Murphy et al. [3] used eight years of wind and temperature data (1959–1967), as determined from grenade data [4], in their analysis, but because of large errors, particularly at the uppermost data point in the years up to 1965, the results of the analysis are open to some question. We shall limit this analysis to the years 1965, 1966 and 1967; even so, the general points raised in the

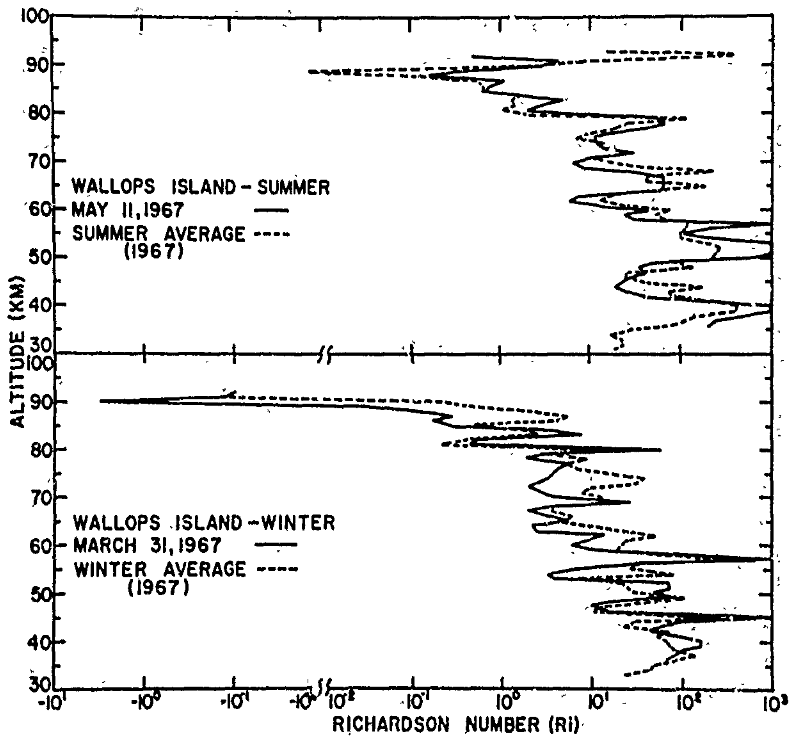

Fig. 2. The Richardson numbers for the Wallops Island (38° N) data. The identification of the curves is the same as in Fig. 1.

eight-year analysis are repeated in this three-year analysis. The winter northern (60° N) mesosphere shows a more intense turbulence than does the summer northern mesosphere (Fig. 1), and the summer data are somewhat sporadic in the occurrence of turbulence above 80 km, while the winter data show turbulence as being present 100% of the time at some altitude above 75 km. The mid-latitude data (38° N) (Fig. 2) show little turbulence for the years 1965 and 1966, but more frequent occurrences for 1967, and, as an overall average, approximately two thirds of all data have some turbulence above 80 km. The

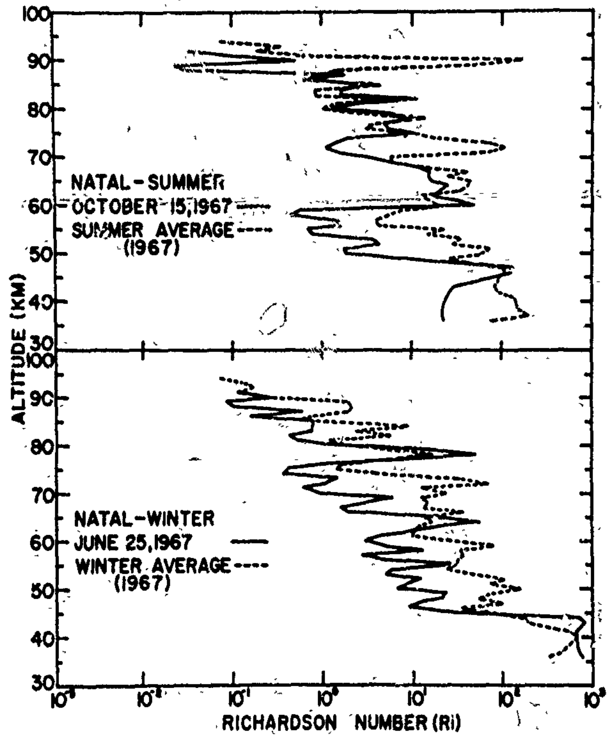

Fig. 3. The Richardson numbers for the Natal (5° S) data. The identification of the curves is the same as Fig. 1.

equatorial region (5° S) shows fairly intense turbulence, present most of the time. The region at or slightly above the stratopause (Figs. 1, 2 and 3) shows sections of marginal stability or even turbulence ($Ri \leqq 1$, and occasionally $\leqq 1/4$), and as theoretically demonstrated by Hines [5], these regions may correspond to a critical layer, signified by $Ri \leqq 1/4$, which results in amplification of the background wind system. Alternatively if the condition $Ri \leqq 1$ signifies a turbulent region, as hypothesized here, the radiation of short internal waves from this region of turbulence arises from the growth of wave modes in the interactions among the turbulent components as noted by Phillips [6]. In either event, it is possible that these stratospheric regions ($Ri \leqq 1$) are a source of wave generation or amplification which propagate through the mesosphere to ionospheric heights and provide the stratosphere-ionospheric coupling as noted by ionospheric physicists.

The winter–summer averages of the rate of viscous dissipation for the three latitudes are given in Figs. 4, 5 and 6 as a function of altitude. The values of ε upon occasion go as high as $\sim 10^6$ ergs g^{-1} s^{-1} and fairly often up to and greater than 10^5 ergs g^{-1} s^{-1}, in regions of unusually high shears ($\sim 10^{-1}$ s^{-1}), particularly at the northern latitudes. Because of the probable error generation of ε discussed earlier, values of $\varepsilon > 1 \times 10^5$ are removed from the calculation

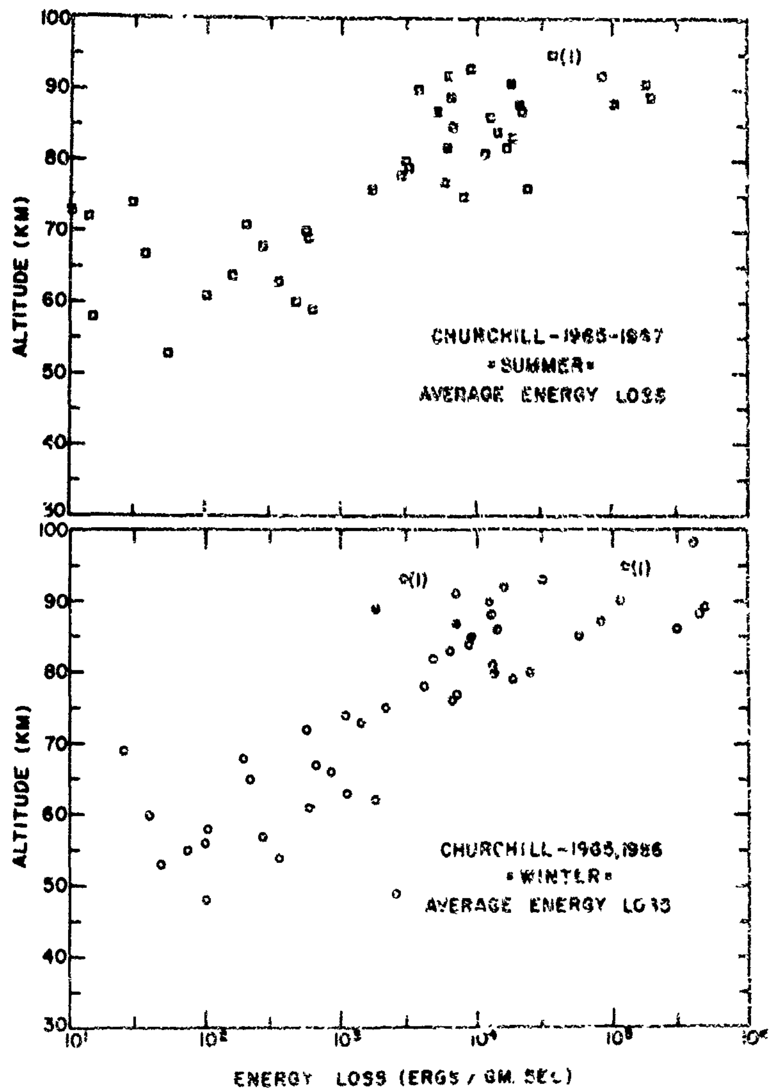

Fig. 4. Average winter [illegible] heating rates at [illegible] Churchill. [illegible] the averages when values [illegible] omitted from the data.

of the mean heating rate, also shown in Fig. 4, 5 and 6. Another factor to consider in removing these values is the low size of the data sample, where a single value of $\varepsilon \gtrsim 10^6$ in a sample of 11 would overwhelm the more statistically significant sample for $\varepsilon \lesssim 10^5$ ergs g^{-1} s^{-1}. It is quite apparent that the average winter northern mesosphere has a rate of heating which is equivalent to or higher than that calculated from the subsidence and recombination of atomic oxygen, and that of Johnson and Gottlieb [7] who calculated a heat source due to the meridional wind distribution which was required to balance the lack of symmetry in the solar heat input for solstice

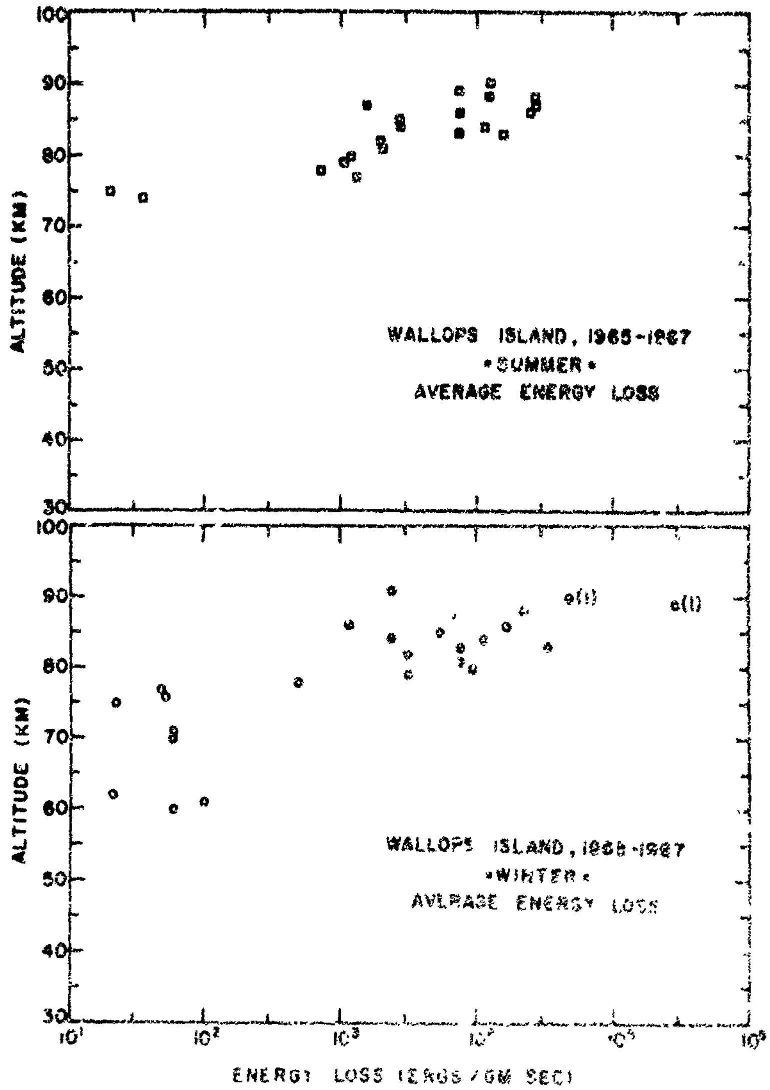

Fig. 5. [illegible]

conditions, and which would then provide the energy necessary to maintain the warm polar mesosphere. The mid-latitude heating rate is obviously much lower than that of the northern latitudes or equatorial region, and is approximately of the same amplitude for the yearly average as that determined by Zimmerman and Rosenberg [8] from wind trail analysis at approximately 95 km. The equatorial region also indicates large heating rates, similar to those of the northern latitude.

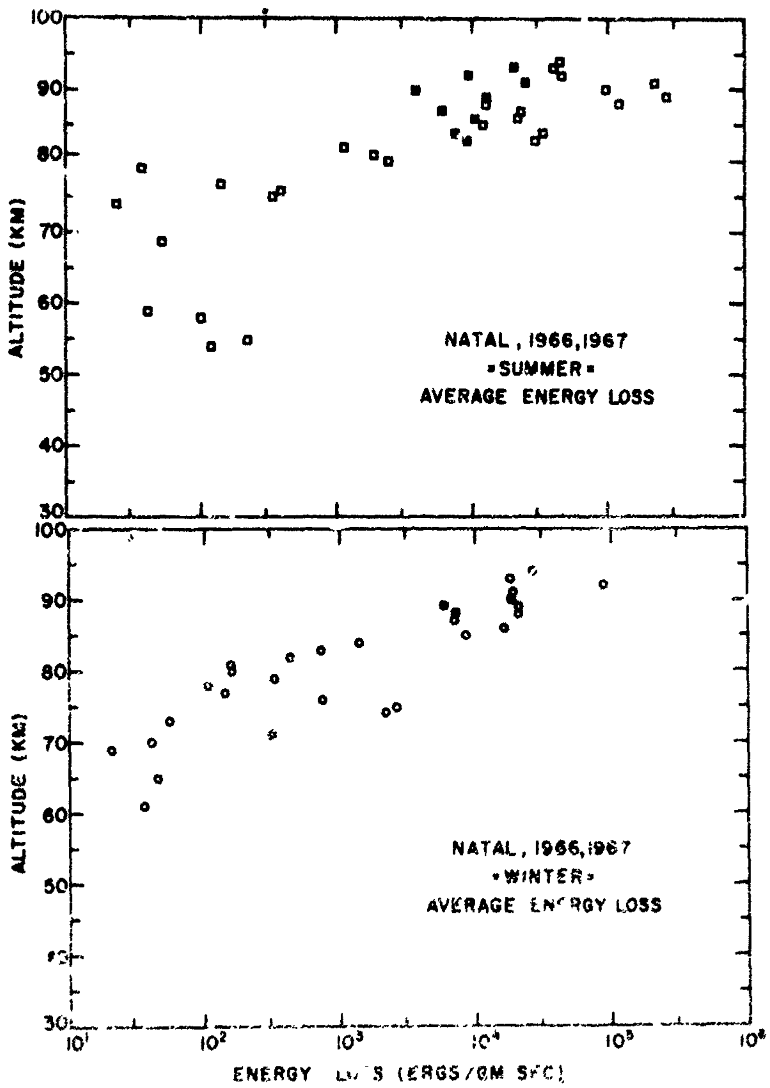

Fig. 6. Same as Fig. 4 for Natal, Brazil

3. Summary

This study shows the atmosphere above 80 km is usually turbulent, with the occurrence of turbulence, at the moment, as some non-predictive function of latitude and season. The average heating of the atmosphere by the turbulent flow system is of the order of or greater than the heating due to euv and Schumann–Runge energy absorption, and apparently compensates for the lack of symmetry in the solar heating input at solstice conditions.

The region around the stratopause shows conditions which are sporadically favorable for wind system amplification; however, many more comparisons of these data with ionospheric mesospheric measurements and analysis must be performed to ascertain this possible influence.

4. Acknowledgments

The authors take this opportunity to thank G. Pereira, of this laboratory, for his aid in the handling and analysis of the data and Drs. N. W. Rosenberg and A. Quesada for their aid and advice with the computations

References

[1] J. D. WOODS, Radio Sci. 4, 1289 (1969).
[2] W. HEISENBERG, Z. Phys. 124, 628 (1948).
[3] E. A. MURPHY et al., Trans. Am. Geophys. Un. 52, 293 (1971).
[4] W. S. SMITH et al., NASA Tech. Rep. Nos. TR R-263, TR R-288 and G-938 (1965, 1966, 1967).
[5] C. O. HINES, J. Atmos. Terr. Phys. 30, 837 (1968).
[6] O. M. PHILLIPS, The Dynamics of the Upper Ocean, Cambridge Univ. Press, Cambridge, England 1966
[7] F. S. JOHNSON and B. GOTTLIEB, Space Research IX, 442, 1969
[8] S. P. ZIMMERMAN and N. W. ROSENBERG, Space Research XII, 623 (1972)