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...and Vertical Variability of the
...ntainty Diffusion Coefficients in

PROJECT 6670
...PHYSICS LABORATORY

...SYSTEMS COMMAND, USAF

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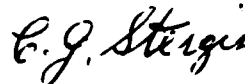
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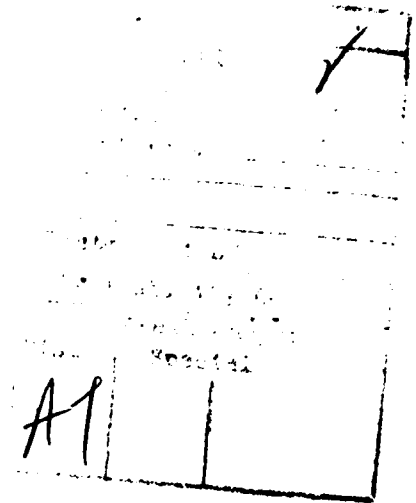
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Utilizing the premise of turbulent energy balance and an altitude-invariant non-dimensional relation of turbulent heat flux, vertical turbulent diffusivities are obtained from bodies of data consisting of winds and temperature in the upper atmosphere. These are specific to the small scale turbulent motions of the atmosphere and should be differentiated from those determined from the study of wave interactive damping mechanisms. The average values of the turbulent diffusivities derived in this work are presented as a function of latitude, and are compared to those derived by an alternative method.

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On the Horizontal and Vertical Variability of the Vertical Turbulent Eddy Diffusion Coefficients in the Mesosphere

1. INTRODUCTION

The parameterization of turbulent diffusivity coefficients in the mesosphere and lower thermosphere is a difficult[1] task that has been approached by many researchers, and rather than enumerate them here, we shall reference some as they occur in this report. Because of the formidable distribution of scale sizes and associated energies that vertically and horizontally transfer material, momentum, and heat in this altitude region, the turbulent diffusivities and dynamics still remain largely undetermined; particularly so in the altitude region 90 km to the turbopause, say ~120 km.

The measurement of turbulence, per se, in the upper atmosphere is extremely difficult. There have been many schemes that, covering the same period and region of the upper air, have produced a number of these diffusivities that differ in orders of magnitude. An example of these large differences between the derived vertical diffusivities are the tropospheric and stratospheric diffusivities reported

1. Ebel, A. (1980) Eddy diffusion models in the mesosphere and lower thermosphere, J. Atmos. Terr. Phys. 42:617-628.

by Danielsen and Louis[2] and those of Zimmerman and Murphy.[3] The former are based upon the distribution of radioactive tracers in the stratosphere that are transported via vertical diffusivity to the troposphere. The latter analysis is based upon the determination of the local Richardson number using Rawinsonde data, and only calculate a diffusivity when the Richardson number (Ri) is less than 0.25.

$$Ri = \frac{\omega_B^2}{(\partial V / \partial Z)^2}$$

where

ω_B is the Brunt Vaissala frequency and $= \left[\frac{g}{T} \left(\frac{\partial T}{\partial Z} + \Gamma \right) \right]$,

$\frac{\partial V}{\partial Z}$ is the magnitude of the vertical shear of the horizontal wind, and Γ is the adiabatic lapse rate. The Danielsen and Louis analysis (Figure 1) shows large penetration of high amplitude vertical diffusivities into the stratosphere, and very little parallelism between the iso-diffusivities and isodensity contours of the radioactive tracers. The diffusivities based upon the average local dynamic situation (Figure 2) strongly parallel the iso-density contours. In the middle atmosphere, the stratosphere into and through the mesosphere, there have been many contributions of the diffusivities. Those covering large latitudinal distances and derived by dynamic analyses are by Ebel,[1] Nastrom and Brown,[4] Blum and Schuchardt,[5] while those following species distributions are exemplified by Brasseur.[6]

2. Danielsen, E. F., and Louis, J. F. (1977) Transport in the Stratosphere, The Upper Atmosphere and Magnetosphere, Studies in Geophysics, National Academy of Sciences, pp. 141-168.

3. Zimmerman, S. P., and Murphy, E. A. (1977) Dynamical and Chemical Coupling Between the Neutral and Ionized Atmosphere, D. Reidel Dodtrecht-Holland pp. 35-47.

4. Nastrom, G. D., and Brown, D. E. (1978) Eddy Diffusion Coefficients and Wind Statistics, Final Report NAS 2-5978, Control Data Corp., Minnesota.

5. Blum, P. W., and Schuchardt, K. G. H. (1978) Semi theoretical models of the Eddy diffusion coefficients based on satellite data, J. Atmos. Terr. Phys. 40:1137-1142.

6. Brasseur, G. (1978) On models bidimensional du comporment de l'ozone dans la 'stratosphere, Planet Space Sci. 26:139-159.

Figure 1. Spring Values of the Isocontours of the Vertical
Turbulent Diffusivities (light solid and dashed lines) as
Contrasted to the Observed (heavy solid lines) Radioactivity
Isodensity Contours
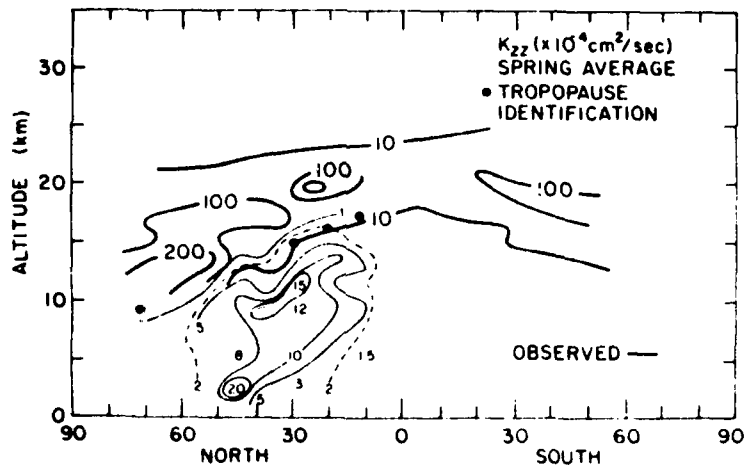
Figure 2. Comparison of the Isodensity Contours of
Radioactive Debris (heavy solid lines) With the Iso-
contours of Vertical Turbulent Diffusivities in the
Northern Hemisphere (light solid and dashed lines).
The numbers on the contours of turbulent diffusivity
are values expressed in $K_{zz} \times 10^{-4}$ cm^2/sec

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2. THEORY

The approach used in this paper follows from the work of Zimmerman and Murphy[3] and Zimmerman and Keneshea.[7] In these works the diffusivity is not derived from the effective vertical damping of gravity wave amplitude by an enhanced diffusivity that is presumed to be turbulent.[1,4,8,9,10] Instead it is based upon the logic that local regions of the atmosphere are in a turbulent state when the measured Richardson number is less than 0.25. Given this region of instability, (Ri $\leq$ 0.25) we can derive the local values of the turbulent heat flux $<w' \theta'>$

where

w' is the amplitude of the vertical component of the turbulent motions,

and

θ' is the potential temperature fluctuation.

Then, given the heat flux we may determine the vertical diffusivity (K_θ) by the relation

$$K_\theta = -\frac{<w' \theta'>}{\partial \theta / \partial Z}$$

where K_θ is the thermal diffusivity

θ is the potential temperature defined by $\theta = T\left(\dfrac{P_o}{P}\right)^{\frac{\gamma-1}{\gamma}}$,

P is the local atmospheric pressure,

P_o is the pressure at earth's surface,

γ is the ratio of specific heats (= 1.4), and

T is the kinetic temperature.

7. Zimmerman, S.P., and Keneshea, T.J. (1981) Stratosphere and mesosphere turbulence, Handbook for M.A.P., Vol. 2, Univ. of Illinois Press, pp. 311-322.

8. Hines, C.O. (1970) Eddy diffusion coefficients due to instabilities in internal gravity waves, J. Geophys. Res. 75:3937-3939.

9. Justus, C.G. (1973) Upper atmospheric mixing by gravity waves, AIAA Paper No. 73-493, 1-3.

10. Zimmerman, S.P. (1974) The effective vertical turbulent viscosity as measured from radio meteor trails, J. Geophys. Res. 79:1095-1100.

Thus, to calculate the vertical component of turbulent diffusivity (K_θ), the spatial distribution of the turbulent vertical heat flux must be determined. Zimmerman and Keneshea[11] have demonstrated a non-dimensional heat flux that is altitude invariant from 5 m to ~ 20 km. Their relation for this non-dimensional spatially invariant flux is the ratio of the outer scale of the spectrum of $<w' \theta'>$ to the vertical scale of the horizontal wind, and is a function of gradient and flux Richardson numbers, the local shear, temperature gradient, and horizontal wind amplitudes.

Using atmospheric measurements of turbulence from Heck and Panofsky,[12] Delay and Dutton,[13] Kennedy and Shapiro,[14] Kaimal et al,[15] Wyngaard and Cote,[16] and Izumi,[17] we calculate the non-dimensional heat flux as a function of Richardson number. The comparison of these data with the length scales ratio are shown in Figure 3 for positive flux Richardson numbers. As observed there is good correspondence between the higher altitude data (~ 20 km) and the boundary layer measurements.
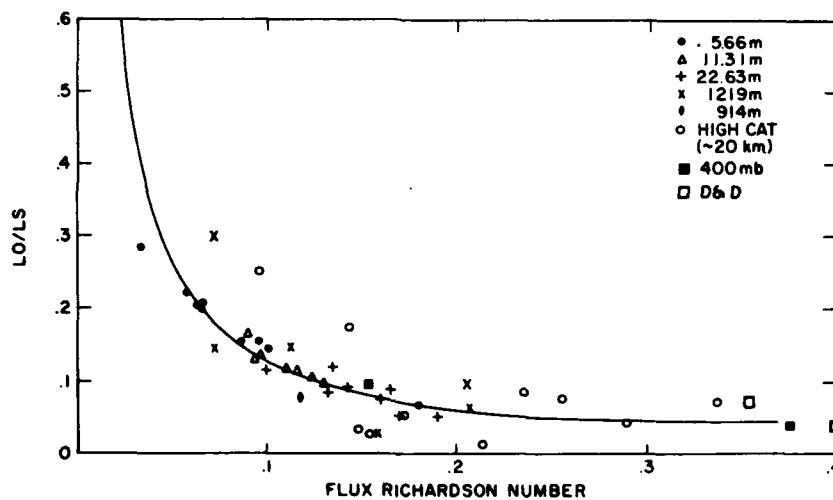

Figure 3. The Non-Dimensional Turbulent Heat Flux as a Function of R_f for Positive R_f. The solid line is a least squares fit to the data

(Due to the large number of References cited above, they will not be listed here. See References, page 15.)

3. DATA

The data base used to determine these turbulence coefficients is the rocket grenade data as used by Zimmerman and Murphy.[3] In this report we discuss the results from the six-month period around the summer solstice. These data provide[18] information on local temperature and winds, with the separation of data points being 2 to 7 km. They were joined with a cubic spline and interpolated to height steps of 1 km (see Figure 4). Then the vertical spatial derivatives were determined and used to calculate the gradient Richardson number (R_i). Inter .ting to 1 km vertical separation can introduce a high frequency amplitude that m   ot be in the real data. This in essence means that larger wind shears and tem_   _ ture gradients are generated than may be in the real data sample. Given th   )dy of data, with a random sampling of data separation, rigorous analysis woulc   ' use data lengths of the largest separation, in this case approximately 7 km.   _w- ever as observed in Figure 5, the use of this large data separation would eliminate completely the observed large positive and negative gradients generated where the experimental data separation are significantly smaller than 7 km. This would be tantamount to saying that the atmosphere is less structured than observed, and would lead to an incorrect assessment of turbulence occurrence. Further, in Figure 5 we observe that the 1-km interpolated points between the measured data points follow the straight line connections between data points quite nicely, except near rapid changes in the sign of the gradient. In these regions, the linear interpolation significantly underestimates the gradients that must be there, while generating infinitely small cusps points. On the other hand, the splined interpolation, as it must, generates a smoothly continuous curve between data points. But again we ask, "Does the splining process generate significant overshoots, and thus larger shears than are present in the original data?" We cannot answer this question directly without having measurements of small scale, equidistant data points. However, we can reduce any erroneous and also some real gradients, by filtering the 1-km splined points. Towards this direction, we used a three-point and five-point running mean. The results of the three-point mean, that is equal to a two-km spacing, are also displayed in Figure 5. As observed, this filtering does not modify the temperature structure to any significant degree, but does somewhat smooth the profile in regions of very large gradients, particularly in regions of gradient sign reversals. This filtering reduces the total number of samples for Richardson number less than or equal to 0.25 by approximately five percent and slightly reduces the total number of large negative Richardson numbers. However, it also somewhat

18. Smith, W.S., Theon, J.S., Swartz, D.C., Katchen, L.B., and Howath, T.T. (1961-1967) NASA Tech Report, TR R211, 245, 263 and 288.

increases the number of occurrences in the region $-0.0625 < R_i < 0$. In calculating the annual mean turbulence parameters, the filtering produces results that differ insignificantly from those obtained using the unfiltered ensemble. The five-point, or 4-km, running mean filter, however, severely reshapes the original profile, so that in many areas the splined and filtered curve no longer passes through many of the observed data points, and thus is excluded from our analyses.
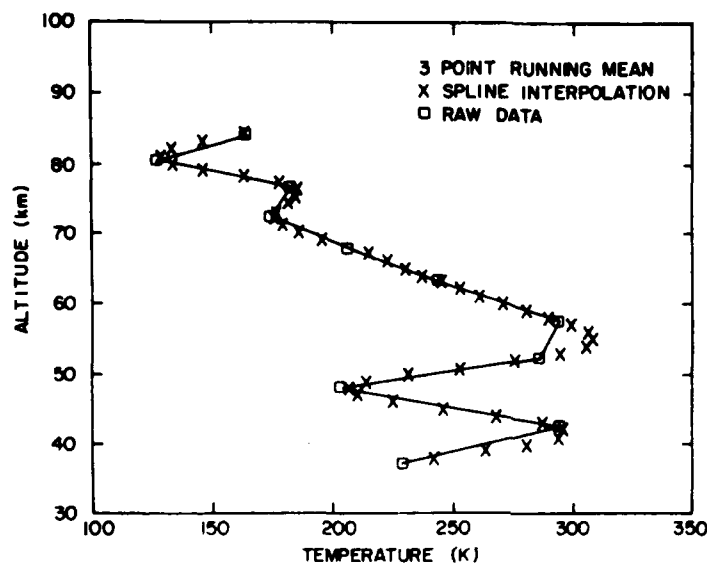

Figure 4. Temperature Measurements From the Grenade Data Analysis Showing the Measured Parts, a Straight Line Interpolation, and the Splined Analysis
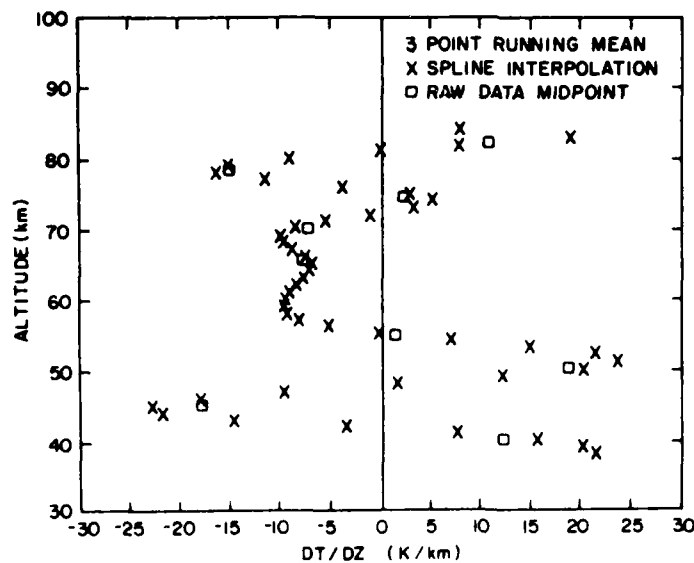

Figure 5. The Vertical Gradients of the Temperature Measurements Shown in Figure 4. The comparisons of the different analyses are discussed in the text

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The specific results using these data show too large a variability in the diffu-
sivities for them to be incorporated into a meridional model, thus the analyzed
data was further filtered with an eleven-point running mean (= 5.5 km). The re-
sults deduced here are given in Figure 6, and for comparison purposes we use
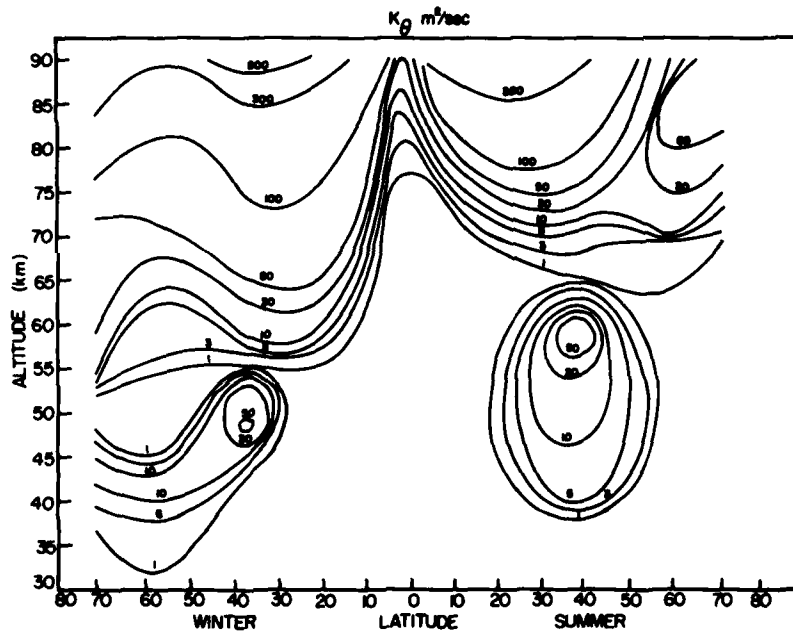Ebel's[1] results deduced from energy degradation (see Figure 7).

Figure 6. The Altitude-latitudinal Variation of Smoothed Vertical Turbulent
Diffusivities (in MKS units) as Deduced From the Turbulent Heat Flux
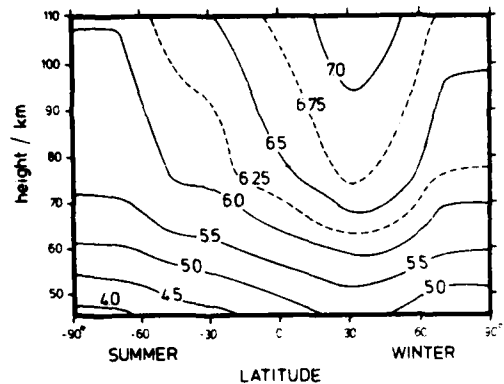$< \omega' \theta' >$ Term

Figure 7. The Altitude-latitude Variation of K_{zz} (in CGS units), Derived by Ebel[1] From the Data on Internal Gravity Wave Degradation

4. RESULTS

Our results, from the northern latitude winter-summer, shown an extreme paucity of diffusion coefficients in the summer northern latitudes and for all seasons in the equatorial latitudes. We feel this is partly due to the lack of resolution inherent in the grenade measurement technique. In the future, we shall supple-ment these analyses with the meteorological Rocket Network data. These new data will enable us to more accurately define the diffusivities where they are now defi-cient due to the lack of high resolution data, and to supplement and correct or add to the values derived from the grenade measurements.

In those regions where there are sufficient occurrences of $R_i < 0.25$, we can more accurately quantify the diffusivities. This is apparent in the mid-latitudes, for both seasons, and the northern winter latitudes. For those regions, we can also make good comparisons with Ebel's work. But wherever we have only sketchy results, we cannot do more than discuss generalities.

Basically, the altitude/latitude variabilities of the two models are somewhat similar, with large diffusivities in the northern latitude winter, decreasing strongly towards the equator and increasing again towards the summer pole. However, it is in the summer polar regions, and to some extent in the northern polar regions, where we observe large differences in the diffusivities at the same altitude. Ebel's results generally show larger diffusivities than we do, indicating that the wave energy is undergoing significant viscous degradation, more so than our results suggest as directly occurring due to a turbulent viscosity. However, the method used for analysis of wave degradation by Ebel and others has no formulation for the inclusion of wave energy transfer to and from the mean motion by wave or turbulent[7] momentum transfer. This would be a possible explanation of the above results, since any mechanism that causes a loss to a propagating wave would, in the context used by Ebel and others, be constituted as only a damping mechanism. Further,

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as shown by Zimmerman and Keneshea[7,11] in their midlatitude summer analysis (Figures 5 and 6) there is large momentum transfer to the mean wind at 80 to 90 km level. Coincident with this, and as expected, the diffusivities at these altitudes are markedly reduced (Figure 7). Thus the latitudinal reduction of Reynolds stress is accompanied by transfer of energy from wave motions into mean mass motion with the accompanying reduction of wave energy and reduced dissipation.

5. CONCLUSION

In conclusion, we demonstrate, using incomplete and coarse data (the only data available to 90 km), the latitudinal variability of stratospheric and mesospheric turbulent diffusivity in the northern hemisphere over portions of the Americas. It is the only analysis, to date, that incorporates the effects of momentum, and thus energy transfer to and from the mean motions. These results thus show some agreement as well as some disagreement with those analyses that do not incorporate these other sources and sinks.

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