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4.4B ESTIMATION OF VERTICAL DIFFUSION FROM OBSERVATIONS OF ATMOSPHERIC TURBULENCE LAYERS

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There have been numerous studies addressing the turbulent diffusion in the stratosphere and mesosphere during the last two decades. The motivation for such studies was the need for an understanding of the thermal and constituent structure of the middle atmosphere. Observational estimates of the horizontal and/or vertical diffusion were obtained using chemical release, rocket vapor trail, aircraft, balloon, and radar techniques (ZIMMERMAN and CHAMPION, 1963; KOCHANSKI, 1964; REES et al., 1972; LILLY et al., 1974; ROSENBERG and DEWAN, 1975; CADET, 1977; VINCENT and STUBBS, 1977; MANSON et al., 1979; SATO and WOODMAN, 1982; FRITTS, 1984). Typical estimates of vertical diffusion near the mesopause ranged from ∿200 to 600 $m^2 s^{-1}$. During the same period, a number of theoretical studies were performed to infer the level of vertical diffusion needed to account for observed constituent profiles (JOHNSON and WILKINS, 1965; COLEGROVE et al., 1966; CUNNOLD et al., 1980; ALLEN et al., 1981; APRUZESE et al., 1984). These estimates were for the most part near 100 $m^2 s^{-1}$. Thus, there appears to be a discrepancy between the level of vertical diffusion required for the dissipation of gravity wave and tidal motions on the one hand and for the maintenance of observed temperature and constituent profiles on the other. The purposes of the present contribution are to outline a possible explanation of this discrepancy and to suggest the measurements that may help verify this explanation.

The theory relies upon laboratory and atmospheric observations that suggest that the saturation of internal gravity waves via either convective or dynamical instabilities results in the generation of thin layers of turbulence which act to mix the local environment (CADET, 1977; KOOP, 1981; SATO and WOODMAN, 1982; WAND et al., 1983; PHILBRICK, private communication, 1983). Provided that these turbulent regions remain associated with the unstable portion of the wave field (as observed), it is then appropriate to examine the consequences of a turbulent diffusion that varies throughout the wave field in a systematic manner. Such an approach was taken in the study by FRITTS and DUNKERTON (1984). The results of that study and their implications for atmospheric observations will be described here.

The evolution of the wave and mean fields in the middle atmosphere are described by the nonlinear, viscous momentum and thermodynamic energy equations and the continuity equation. Using the continuity equation, the mean and perturbation potential temperature equations may be written

$$\overline{\theta}_t + (\overline{w'\theta'})_z = (\overline{\nu\theta}_z + \overline{\nu'\theta_z'})_z \tag{1}$$

and

$$\theta_t' + \overline{u}\theta_x' + w'\overline{\theta}_z + u'\theta_x' + w'\theta_z' - (\overline{w'\theta'})_z = (\nu'\overline{\theta}_z + \overline{\nu}\theta_z' + \nu'\theta_z' - \overline{\nu'\theta_z'})_z + (\overline{\nu}_h\theta_x' + \nu_h'\theta_h')_x \quad , \tag{2}$$

where subscripts denote differential and ν_h is the horizontal component of turbulent diffusion assumed proportional to ν. Then, consistent with the laboratory observations by KOOP (1981), we assume a symmetric distribution of turbulent diffusion with a maximum at the point at which the total potential temperature is minimum (negative for convective instability). The total potential temperature may be written as

$$\theta_z = \overline{\theta}_z\,(1 + \alpha\cos\phi) \tag{3}$$

where ϕ is the phase function

$$\phi = kx + mz - kct \tag{4}$$

and α is a measure of wave amplitude ($\alpha > 1$ for convective instability).

We now multiply (2) by θ', average horizontally, and note that all triple correlations are small by virtue of the assumed negative correlation of ν' and θ_z'. Solving $\overline{w'\,\theta'}$, assuming steady-state saturation, and substituting into (1) yields, after some manipulation,

$$\overline{\theta}_t = \frac{\partial}{\partial z}\left(\overline{\nu\theta_z} + 2\,\overline{\nu'\theta_z'} + \frac{\overline{\nu}}{\overline{\theta}_z}\,\overline{\theta_z'^2} + \frac{\overline{\nu}_h}{\overline{\theta}_z}\,\overline{\theta_x'^2}\right) \quad . \tag{5}$$

Alternatively, if we write

$$\nu = \nu_o\, f(\cos\phi) \tag{6}$$

and define a positive coefficient

$$\beta \equiv -\frac{\overline{\nu\cos\phi}}{\overline{\nu}} \tag{7}$$

then (5) becomes

$$\overline{\theta}_t = \frac{\partial}{\partial z}\left\{(\overline{\nu\theta_z})\left[1 - 2\alpha\beta + \frac{\alpha^2}{2}\left(1 + \frac{k^2}{m^2}\frac{\overline{\nu}_h}{\overline{\nu}}\right)\right]\right\} \quad . \tag{8}$$

This expression relates all of the wave and turbulence contributions to the mean turbulence flux of potential temperature. In particular, we note that both wave and turbulence fluxes include a countergradient diffusion term due to the occurrence of a maximum turbulent diffusion in regions of minimum θ_z. These countergradient terms reduce the total down-gradient flux of heat (and constituents) expected on the basis of a uniform turbulent diffusion.

The magnitude of this effect can be illustrated by assuming a distribution for turbulent diffusion of the form

$$\nu = \nu_o\left(\frac{1 - \cos\phi}{2}\right)^n \tag{9}$$

for $n \geq 0$ so that

$$\beta = \frac{n}{n + 1} \tag{10}$$

The magnitudes of the turbulence and total (wave plus turbulence) fluxes are shown for various values of n (and negligible $\overline{\nu}_h$) in Figure 1 with dashed and solid lines, respectively. Note that even for relatively small n (broad distributions of turbulent diffusion), the countergradient terms results in an appreciable reduction of the total flux expected for a uniform turbulent diffusion (with $n = 0$). In the case where ν and ν_h contribute equally to the down-gradient wave flux, it is possible to show that, while the reduction of the total flux may be significant, the total flux remains down-gradient for all α and $n \geq 0$ (FRITTS and DUNKERTON, 1984). Additionally, this theory predicts

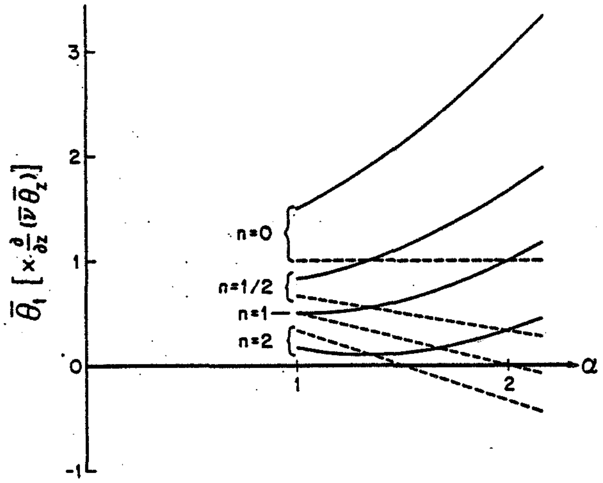

Figure 1. Variation of the turbulent (- - -) and net (——) rate of change of mean potential temperature normalized by the turbulent transport of the mean for the case of a vertical turbulent diffusivity alone. Note that even small values of n cause significant reductions in the net flux.

Prandtl numbers in the mean and perturbation equations much larger than unity for nonuniform turbulent diffusion, as anticipated by CHAO and SCHOEBERL (1984).

These results imply that, while a large turbulent diffusion may be required for the dissipation of gravity wave and tidal motions, a substantially smaller effective diffusion may act to mix mean gradients of potential temperature and constituents. The reason is that intense turbulence (and diffusion) is expected in precisely those regions of the gravity-wave field in which the total gradients of potential temperature and constituents are small or opposite to those of the mean state. Thus, it may not be sufficient to infer a mean vertical diffusion by assuming an average level of turbulence applied to the mean state. It may be necessary, instead, to infer the effective vertical diffusion from observations of the intensity and distribution of turbulent layers as well as the thermal structure. Such measurements could resolve the discrepancy between the level of turbulence required to dissipate observed wave motions and that needed to explain middle atmosphere constituent distributions.

REFERENCES

Allen, M., Y. L. Yung and J. W. Waters (1981), Vertical transport and photochemistry in the terrestrial mesosphere and lower thermosphere (50-120 km), J. Geophys. Res., 86, 3617-3627.

Apruzese, J. P., D. F. Strobel and M. R. Schoeberl (1984), Parameterization of IR cooling in a middle atmosphere dynamics model; 2. Non-LTE radiative transfer and the globally averaged temperature of the mesosphere and lower thermosphere, submitted to J. Geophys. Res.

Cadet, D. (1977), Energy dissipation within intermittent clear air turbulence patches, J. Atmos. Sci., 34, 137-142.

Chao, W. C. and M. R. Schoeberl (1984), A note on the linear application of gravity wave saturation in the mesosphere, J. Atmos. Sci. (in press).

Colegrove, F. D., F. S. Johnson and W. B. Hanson (1966), Atmospheric composition in the lower thermosphere, J. Geophys. Res., 71, 2227-2236.

Cunnold, D. M., F. N. Alyea and R. G. Prinn (1980), Preliminary calculations concerning the maintenance of the zonal mean ozone distribution in the northern hemisphere, Pure Appl. Geophys., 118, 329-354.

Fritts, D. C. (1984), Gravity wave saturation in the middle atmosphere: A review of theory and observations, Rev. Geophys. Space Phys. (in press).

Fritts, D. C. and T. J. Dunkerton (1984), Fluxes of heat and constituents due to convectively unstable gravity waves, submitted to J. Atmos. Sci.

Johnson, F. S. and E. M. Wilkins (1965), Thermal upper limit on eddy diffusion in the mesosphere and lower thermosphere, J. Geophys. Res., 70, 1281-1284.

Kochanski, A. (1964), Atmospheric motions from sodium cloud drifts, J. Geophys. Res., 69, 3651-3662.

Koop, C. G. (1981), A preliminary investigation of the interaction of internal gravity waves with a steady shearing motion, J. Fluid. Mech., 113, 347-386.

Lilly, D. K., D. E. Waco and S. I. Adelfang (1974), Stratospheric mixing estimated from high-altitude turbulence measurements, J. Appl. Meteorol., 13, 488-493.

Manson, A. H., C. E. Meek and R. J. Stening (1979), The role of atmospheric waves (1.5 hr - 10 days) in the dynamics of the mesosphere and lower thermosphere at Saskatoon (52°N, 107°W) during four seasons of 1976, J. Atmos. Terr. Phys., 41, 325-335.

Rees, D., R. G. Roper, K. H. Lloyd and C. H. Low (1972), Determination of the structure of the atmosphere between 90 and 250 km by means of contaminant releases at Woomera, May 1968, Phil. Trans. R. Soc., 271, 631-666.

Rosenberg, N. W. and E. M. Dewan (1975), Stratospheric turbulence and vertical effective diffusion coefficient, Environ. Res., 535, Air Force Cambridge Res. Labs.

Sato, T. and R. F. Woodman (1982), Fine altitude resolution observations of stratospheric turbulent layers by the Arecibo 430 MHz radar, J. Atmos. Sci., 39, 2546-2552.

Vincent, R. A. and T. J. Stubbs (1977), A study of motions in the winter mesosphere using the partial reflection drift technique, Planet. Space Sci., 25, 441-455.

Wand, R. H., P. K. Rastogi, B. J. Watkins and G. B. Loriot (1983), Fine Doppler resolution observations of thin turbulence structures in the tropo-stratosphere at Millstone Hill, J. Geophys. Res., 88, 3851-3857.

Zimmerman, S. P. and K. S. W. Champion (1963), Transport processes in the upper atmosphere, J. Geophys. Res., 68, 3049-3056.