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14. Abstract/Notes

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A NOTE ON VERTICAL EDDY DIFFUSION COEFFICIENT FROM THE LANDSAT IMAGERY

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Abstract

The vertical eddy diffusion coefficient (K_z) for five stable cases of the smoke plumes is estimated from the published LANDSAT imagery data sets of Viswanadham and Torsani (1982) and Weinstock's formulation. These results agree well with the previous experimental values obtained over water surface by various workers.

1. Introduction

There is a strong tendency in the atmosphere and oceans for the development of layers in which the local gravitational stability of the air or water is either very large or very small. With larger stable layers in the atmosphere or in the bodies of water there are often found sublayers, sometimes very numerous, in which neutrally stable layers are sandwiched between thin stable layers (Stommel and Fedorov, 1967). Some of this is understood to some extent, but many of these layering phenomena still have no acceptable explanations. The dispersion of atmospheric pollutants in such stable layers is limited by the height of the inversion lid.

In this note we will present the calculated values of the vertical eddy diffusion coefficient for five stable cases in the atmospheric boundary layer from the LANDSAT imagery of the smoke plumes. It will be shown that the diffusion coefficient values derived under some approximations, with relatively simple expressions account qualitatively in a reasonable way for the various effects.

2. Theoretical considerations

Determination of the vertical standard deviation σ_z(m) and the corresponding vertical eddy diffusion coefficient, K_z($m^2\ s^{-1}$), depend on measurements of concentration at different heights and hence are less reliable than those of the lateral standard deviation, σ_y(m) and the apparent lateral eddy diffusion coefficient, K_y($m^2\ s^{-1}$). Ozmidov (1965) and Lilly et al. (1974) have proposed the following expressions for K_z

$$K_z = 0.1 \frac{\varepsilon}{N^2} \qquad \text{(Ozmidov)} \tag{1}$$

and

$$K_z = \frac{\varepsilon}{3N^2} \qquad \text{(Lilly et al.)} \tag{2}$$

in the ocean and in the stratosphere, respectively. Here, N(s^{-1}) is the Brunt-Väisälä frequency. The vertical turbulent diffusion coefficient in a stably stratified fluid is also derived analytically by Weinstock (1978), i.e.,

$$K_z \simeq 0.81 \frac{\varepsilon}{N^2} . \tag{3}$$

Eqs. (1) and (2) are similar to the expression (3) derived by Weinstock (1978) except that Weinstock uses a coefficient of 0.81 rather than 0.1 and 1/3. Recently, this expression was tested by experiments of Caldwell et al. (1980). Their measurements and analysis suggest that Eq. (3) is valid for strong or intermediate stratification (actually, for Cox numbers less than 2500) and is not valid for weak stratification (Cox numbers exceeding 2500). The fact that Eq. (3) cannot be valid for weak stratification is evident from its form, which shows a singularity at N=0.

From the previous discussion we know that Eq. (3) is valid for strong or intermediate stratification. So, it is better to compare the critical wave number $k_o(m^{-1})$ of the energy-containing region of the spectrum with the microscale wave number $k_N(m^{-1})$ of Kolmogorov (1941). An expression for the critical wave number, k_o, was derived theoretically by Dougherty (1961) as

$$k_o = \varepsilon^{1/2} \left(\frac{g}{\theta}\frac{d\theta}{dz}\right)^{3/4} = \varepsilon^{1/2} (N^2)^{3/4} \tag{4}$$

where θ (K) is the potential temperature. Eq. (4) provides an estimate of the wave number at which buoyancy cannot be neglected (see Weinstock, 1981; Fig. 1, p. 9927). The Kolmogorov's microscale wave number, k_N, is

$$k_N = \varepsilon^{1/4} \nu^{-3/4} . \tag{5}$$

Here, $\nu(m^2\ s^{-1})$ is the kinematic viscosity of air. Thus, the isotropic inertial range exists when $k_N > k_o$ or $L_N < L_o$, where L (m) is the wavelength (= k^{-1}). If, on the other hand, $k_N < k_o$ or $L_N > L_o$, there is no inertial range in the usual sense. In other words, we have weak stratification (i.e., most turbulence scales are not influenced by buoyancy). In case, when $k_N < k_0$, we should not use Eq. (3) to calculate K_z in a stably stratified atmospheric flows. Assuming an inertial range turbulence of the atmosphere, the values of K_z are obtained from Eq. (3) in the present study.

3. Results

The details of site description and data sets are presented elsewhere (Viswanadham and Torsani, 1982). The theoretical information in section 2 and the data sets of Viswanadham and Torsani (1982) were used to make estimates of the K_z values in five stable cases. The critical wavelength L_0 and the Kolmogorov's microscale wavelength L_N were also evaluated from Eqs. (4) and (5), respectively. In Eq. (5), the value of 0.15×10^{-4} m^2 s^{-1} is used for the kinematic viscosity of air ν (Sutton, 1953, p. 38). The results are presented in Table 1 together with other parameters. It is of interest to consider the numerical values of L_0 and L_N in Table 1. In all 5 cases, L_0 is much greater than L_N. Thus, the isotropic inertial subrange spectrum exists in all cases. This argument shows that the utilization of Eq. (3) to calculate K_z in these 5 cases is well justified.

Although one might expect some values for the parameter K_z, a check with actual measurements is more important. In Table 2, we have summarized results of K_z from four available publications. Note that the reported values of K_z in Table 2 are directly comparable with our results in Table 1. Lilly et al. (1974) have showed that ε is subjected to large variations with underlying surface characteristics, height, latitude and season. Also, the measurements over water are difficult; because the spectra extend to very small scales.

TABLE 1. Estimation of the vertical eddy diffusion coefficient (K_z) from smoke plumes of the LANDSAT imagery

n	RI	$\overline{T}$ K	$\frac{\partial\theta}{\partial z}$ K/100m	N^2 s^{-2} x 10^{-4}	Mean ε $m^2\ s^{-3}$ x 10^{-5}	Mean K_z $m^2\ s^{-1}$	L_o m	L_N x 10^{-4}m
5	0.046	296.3	0.56	1.85	0.3	0.01	1.07	58.55
7	0.006	301.1	0.06	0.20	8.1	3.42	30.91	25.31
8	0.081	294.3	0.47	1.57	7.0	0.37	6.01	26.31
10	0.040	299.3	0.70	2.29	0.9	0.03	1.63	43.80
11	0.088	296.1	0.27	0.89	0.2	0.02	1.48	65.44

n is the reference number of LANDSAT image cases, RI is the Richardson number, $\overline{T}$ is the mean temperature of the screen temperature T_s and 76m stack height temperature T_h, $\partial\theta/\partial z$ is the potential temperature gradient below 80m and N is the Brunt Väisälä frequency. Values of RI, T_s, T_h, $\partial\theta/\partial z$ and ε are taken from Viswanadham and Torsani (1982, Tables 1 and 4).

TABLE 2. Values of K_z from the literature

Source	Location	K_z $m^2\ s^{-1}$
Roll (1965)	Over water	0.05 - 3.50
Bowden et al. (1974)	Sea areas	
	Irish sea	10^{-5} - 10^{-2}
Lyons (1975)	Over water	1.50 - 3.00
Pond and Bryan (1976)	Ocean area	0.01
Present results	Ocean area	0.01 - 3.42

Over water, while the surface roughness tends to be more nearly uniform than one could find at land sites, sharp variations in water surface temperatures, especially in near shore region, can make the interpretation of data exceedingly difficult. Furthermore, during periods of offshore flow, intense air mass modification occurs. Thus, as air flows from land to cold water, not only does the boundary temperature change, but the value of the eddy conductivity begins to decrease to a lower value. There is a destruction of the mechanism involved in creating high K_z values over land, since much of the vertical eddy transport probably results from penetrative convection acting over and above the effect of mechanical turbulence. It is this convective component that disappears very rapidly with fetch. The drop in K_z from around 10 to 10^{-1} m^2 s^{-1} (or less) probably occurs within the first few kilometers. The air's vertical diffusion rates should indeed be low in such an environment. There is little or no quantitative information on fetch, wind duration, sea state, etc., with which Viswanadham and Torsani (1982) have categorized σ_y in their analysis. Their effects (if any) remain obscure in the data scatter of individual data sets (Table 1). In summary, the present data sets give evidence that the average vertical diffusion in the lower troposphere is somewhat lower in the stable marine boundary layer. On the other hand, the vertical diffusion coefficients of smaller magnitude compared to ours are found appropriate when the quasi-horizontal processes are explicitly represented in the global atmosphere.

4. Concluding remarks

It is obvious that this study is an attempt to a statistical description of the turbulent microstructure in the marine boundary layer rather than a definite study. Until the nature of the microstructure is proven conclusively our understanding of the mechanisms producing the mixing events will be limited. Therefore, much more needs to be known about the physical processes of stratified turbulence before its signature can be properly identified in the marine boundary layer. In the marine atmosphere, however, ingeneous observational techniques had to be applied before any relevant results could be obtained.

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