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14. Abstract/Notes

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VERTICAL EDDY DIFFUSION COEFFICIENT FROM THE LANDSAT IMAGERY

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ABSTRACT

Analysis of five stable cases of the smoke plumes that originated in eastern Cabo Frio (22^{o} 59'S; 42^{o} 02'W), Brazil using LANDSAT imagery is presented for different months and years. From these images the lateral standard deviation (σ_y) and the lateral eddy diffusion coefficient (K_y) are obtained from the formula based on Taylor's theory of "diffusion by continuous moment". The rate of kinetic energy dissipation (ε) is evaluated from the diffusion parameters σ_y and K_y. Then, the vertical diffusion coefficient (K_z) is estimated using Weinstock's formulation. These results agree well with the previous experimental values obtained over water surfaces by various workers. Values of ε and K_z show the weaker mixing processes in the marine stable boundary layer. The data sample is apparently too small to include representative active turbulent regions because such regions are so intermittent in time and in space. These results form a data base for use in the development and validation of mesoscale atmospheric diffusion models.

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. It is evident that the dispersion of atmospheric pollutants in such stable layers is limited by the height of the inversion lid.

Considerable interest has been evidenced in the properties of the idealized planetary boundary layer contributing to the dispersion of trace materials introduced into it. While the list of technologically significant atmospheric dispersion problems is very long, those dealing with the fate of pollutants from industrial processes have received special attention in the last decade as part of an effort to evaluate and predict air quality. Efforts to specify atmospheric eddy diffusion coefficient for use in the models have centered primarily on an analysis of different studies (Tennekes, 1973; Bauer, 1974; Mellor and Yamada, 1974; Yamada and Mellor, 1975; Zeman and Tennekes, 1977; Freeman, 1977; Bauer et al., 1978; Remsberg, 1980). We mention some of the meteorological properties involved in this important and difficult problem: wind speed, thermal stratification, wind shear and turning. These properties affect the dispersion of pollutants through the advective transport of the wind field and the diffusion which the pollutant experiences due to the turbulent motion of the atmosphere.

In this paper we investigate the vertical eddy diffusion coefficient 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 above effects.

2. Theoretical considerations

The relationship between the apparent lateral eddy diffusion coefficient, K_y (m^2 s^{-1}), and the lateral standard deviation, σ_y (m), along the y- axis was derived by Batchelor (1949):

$$K_y = \frac{1}{2} \frac{d\sigma_y^2}{dt} \tag{1}$$

where t (s) is the time. This expression, with Taylor's (1921) hypothesis becomes

$$K_y = \frac{1}{2} \bar{u} \frac{d\sigma_y^2}{dx} \tag{2}$$

where $\bar{u}$ ($m\ s^{-2}$) is the mean wind speed along the x-axis. The turbulent energy dissipation rate ε (m^2 s^{-3}) is estimated from the atmospheric turbulent diffusion study of Richardson (1926):

$$K_y \propto \varepsilon^{1/3} \quad \sigma_y^{4/3} \tag{3}$$

where here σ_y is the scale of the diffusion process. The above treatment has already been presented in a detailed form elsewhere (Viswanadham and Torsani, 1982).

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 σ_y^2 and K_y. Ozmidov (1965) and Lilly et al. (1974) have proposed the following expressions for K_z as

$$K_z = 0.1 \frac{\varepsilon}{N^2} \qquad \text{(Ozmidov)} \tag{4}$$

and

$$K_z = \frac{\varepsilon}{3N^2} \qquad \text{(Lilly et al.)} \tag{5}$$

in the ocean and in the stratosphere, respectively. Here, N (s^{-1}) is the Brunt-Väisälä frequency. Eq. (5) for K_z was based on the assumption that the flux Richardson's number (R_f) is 1/4 and that the energy production is locally dissipated. The vertical turbulent diffusion coefficient in a stably stratified fluid is also derived analytically by Weinstock (1978), i.e.,

$$K_z = 0.81 \frac{\varepsilon}{N^2} . \tag{6}$$

Eqs. (4) and (5) are similar to the expression (6) derived by Weinstock (1978) except that Weinstock uses a coefficient of 0.81 rather than 0.1 and 1/3. Since, theoretically most occurrences of turbulence should have small enough scales in the inertial subrange, it follows that Eq. (6) might be fairly generally applied to stable atmospheric conditions (Weinstock, 1978). Weinstock's derivation does not require that the flux Richardson number be known or specified. Recently, this expression was tested by experiments of Caldwell et al. (1980). Their measurements and analysis suggest that Eq. (6) 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. (6) cannot be valid for weak stratification is evident from its form, which shows a singularity at $N = 0$.

Incidentally, Weinstock (1981) has also derived another expression for K_z that applies to all stable stratifications (all positive N^2). This derivation is based on one improvement of the derivation of Eq. (6). In the derivation of Eq. (6), it was assumed that the vertical velocity spectrum is "cut off" at the buoyancy wave number regardless of how weak the stratification is. However, a cutoff is not valid when the stratification is very weak ($N \sim o$), as the stratification is then simply too weak to influence the spectrum. Hence the improvement Weinstock (1981) made is to account for general stable stratification without using a cutoff at all. Eq. (6) gives a larger diffusion coefficient, by a factor of 2.4, than Eq. (5). Weinstock (1978) presented a detailed explanation to this fact. He suggested that the smaller value of K_z obtained by Lilly et al. might be attributed to the fact that turbulent fluxes were neglected (i.e, production was assumed to be local). This led to a smaller value of R_f ($= 1/4$) and a smaller value of K_z.

From the previous discussion we know that Eq. (6) 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

$$k_o = \varepsilon^{1/2} \left(\frac{g}{\bar{\theta}} \frac{d\theta}{dz}\right)^{3/4} = \varepsilon^{1/2} \left(N^2\right)^{3/4} \tag{7}$$

where θ (K) is the potential temperature. Eq. (7) 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

$$k_N = \varepsilon^{1/4} \quad \nu^{-3/4}. \tag{8}$$

Here, ν ($m^2\ s^{-1}$) 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 $k_N < k_o$, we should not use Eq. (6) 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. (6) in the present study.

3. Site and data sets

Fábrica Nacional de Álcalis (National Chemicals Industry) is located on the east shore of Cabo Frio 120 km east of Rio de Janeiro city in the state of Rio de Janeiro, Brazil. The Cabo Frio area (22^{o} 59'S; 42^{o} 02'W) of eastern Brazil is a flat coastal region bounded north-west by land and the rest by Atlantic Ocean. The coast is straight uncomplicated and typical of much of the northern and southern seaboard from Cabo Frio town (see Viswanadham and Torsani, 1982).

The LANDSAT data were obtained during the period April 1975 to July 1978 at the above site. In all, 5 LANDSAT images have been obtained during this period showing stable conditions. Figs. 1 and 2 are typical examples of the LANDSAT images. The processing of the LANDSAT images, and the determination of σ_y and K_y from the LANDSAT images are presented elsewhere (Viswanadham and Torsani, 1982). The mesoscale meteorological conditions existing during the period (1975 - 1978) of interest could be approximated from available data of two agencies. Due to the lack of adequate meteorological measurements for inland and oceanic areas of the Cabo Frio region at the time of LANDSAT imagery, the upper air observations data were taken from the international airport of Galeão, Rio de Janeiro, Brazil (Força Aérea Brasileira, 1974 - 1978). Surface meteorological parameters are available at a weather station very close to the industry (Instituto Nacional de Meteorologia, 1968 - 1978). The radiosonde soundings for each LANDSAT imagery were plotted for the three-day period (one day before and one day after the passage of satellite) along with the surface data. Then, the interpolated mean wind speed was obtained at 76-m stack height level. In Table 1, the

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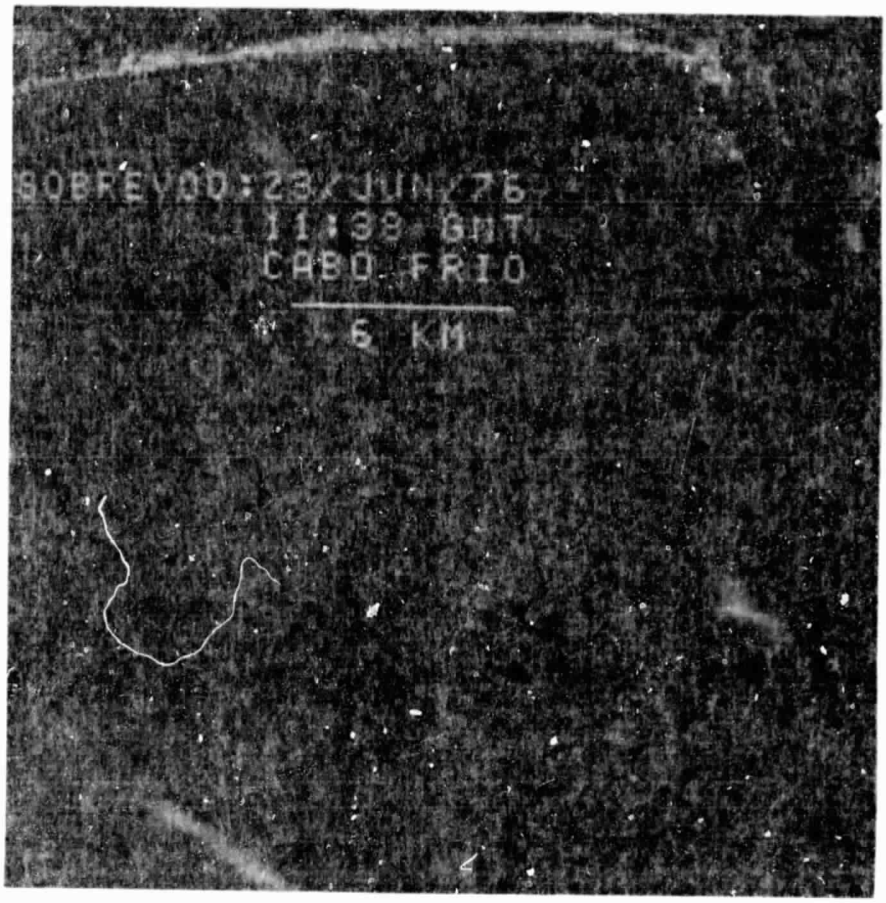

Fig. 2

TABLE 1. Reference number of LANDSAT image cases (n) and various meteorological parameters.

n	Date	Time in hours		C	Wind Speed $m\ s^{-1}$		DDD	Temperature ^{o}C		$\frac{\partial\theta}{\partial z}$	Inversion height m		RI
		G.M.T.	Local		10 m	76 m		T_s	T_h	K/100m	Base	Top	
5	23-06-76	1130	0830	0	8.3	4.1	NNE	23.0	23.5	0.56	1300	1500	0.046
7	11-03-77	1153	0853	10	4.1	8.0	NE	28.0	28.2	0.06	Surface	400	0.006
8	22-05-77	1145	0845	40	1.1	4.0	W	21.0	21.5	0.47	Surface 1800	900 2500	0.081
10	31-10-77	1136	0836	20	10.0	5.0	NNE	26.0	26.6	0.70	Surface	400	0.040
11	01-07-78	1205	0905	0	2.5	4.6	N	23.0	23.2	0.27	Surface	500	0.088

C is the cloud cover in percent, CDD is the prominent wind direction, T_s is the screen temperature, T_h = 76-m stack height temperature, $\partial\theta/\partial z$ is the potential temperature gradient below 80 m, and RI is the Richardson gradient number. The wind speed at 76-m stack height level, T_h and $\partial\theta/\partial z$, are obtained from radiosonde ascents at the international airport of Galeão. The stability parameter RI is calculated by using mean value of T_s and T_h and the wind shear between 10 and 76 m. In the case of missing wind data at 10-m level, the wind shear is obtained after dividing the 76-m level wind speed by 76 m.

base and top of the inversion heights are presented along with other details. The LANDSAT images have been characterized by a gradient Richardson number (RI) computed from radiosonde and surface observations. The RI values given in Table 1 are approximate due to the interpolation involved in the data sets. However, on the basis of RI values, we have five stable cases (Table 1).

The lapse rates were estimated over land due to lack of marine meteorological observations. Unfortunately, we don't have evidence that the determined lapse rate over land is a reasonable estimate for the lapse rate in the boundary layer over water. There are probably more important differences in mesoscale climate in the 200 km east of the Cabo Frio region. The data are summarized in Table 1 and do in fact suggest some meaningful interrelationships (see Viswanadham and Torsani, 1982). The LANDSAT image cases show that the smoke plumes are very long and well defined except the case n=8 (Table 2). From these long and well defined visible plume characteristics of the LANDSAT images, we may have some evidence that the smoke plumes are associated with stable lapse rates in the marine boundary layer. In the normal lower atmosphere N^2 is of the order of 10^{-4} s^{-2}. Generally, N^2 does not show abrupt variations in its order of magnitude from near inland through marine atmosphere. Therefore, in the present study the estimated lapse rate over land does not create abnormal differences in the calculations of vertical diffusion coefficient in the marine atmosphere.

TABLE 2. Estimation of the rate of kinetic energy dissipation (ε) and vertical eddy diffusion coefficient (K_z) from smoke plumes of the LANDSAT imagery.

n	X_t km	RI	$\overline{T}$ K	$\frac{\partial\theta}{\partial z}$ K/100m	N^2 s^{-2} x 10^{-4}	Mean σ_y m	Mean K_y $m^2\ s^{-1}$	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	130	0.046	296.3	0.56	1.85	238	21	0.29	0.01	1.07	58.55
7	100	0.006	301.1	0.06	0.20	724	282	8.23	3.42	30.91	25.31
8	20	0.081	294.3	0.47	1.57	241	62	7.05	0.37	6.01	26.31
10	80	0.040	299.3	0.70	2.29	424	67	0.92	0.03	1.63	43.80
11	100	0.088	296.1	0.27	0.89	250	19	0.18	0.02	1.48	65.44

N is the Brunt Väisälä frequency. X_t is the total plume length. $\overline{T}$ is equal to $(T_s + T_h)/2$ from Table 1. Mean values of σ_y and K_y are taken from Viswanadham and Torsani (1982, Table 2).

4. Results

As suggested in the introduction, the phenomenon of plume spread can be quite complex unless some simplifications can be made by careful scrutiny of the physics of the problem. Therefore, in this section important deductions are made from the LANDSAT imagery of smoke plumes and established physical laws of fluid mechanics. On the basis of RI values, we classified the LANDSAT images into 5 stable cases (i.e, n = 5, 7, 8, 10 and 11, in Table 1). Then the measurements of plume geometry provided the estimate of lateral standard deviation σ_y and the lateral eddy diffusion coefficient K_y for 5 cases (Viswanadham and Torsani, 1982).

Monin (1959) suggested that the diffusion along the vertical in the lower layer of the atmosphere is mostly due to small-scale turbulence. And, the Richardson law $K(\sigma_y) \propto \sigma_y^{4/3}$ was explained by Obukhov (1941) as a consequence of the similarity hypotheses of Kolmogorov (1941) for turbulence with a large Reynolds number. According to Kolmogorov's hypothesis, the rate of kinetic energy dissipation ε is the only parameter determining the turbulent regime within the inertial range. If the diffusion occurs as a result of turbulent motions of scales within the inertial range, then Eq. (3) is obtained. In other words we obtain Richardson's law. Therefore, the parameter ε by viscous process can be regarded as a good indicator of the strength of turbulence. With the aid of theoretical considerations in the section 2, the LANDSAT data sets were used to make estimates of

ε and the vertical eddy diffusion coefficient K_z in 5 stable cases. The critical wavelength L_0 and the Kolmogorov's microscale wavelength L_N were also evaluated from Eqs. (7) and (8), respectively. In Eq. (8), 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 2 together with other parameters. In Table 2, values of ε and K_z show the weaker mixing processes in the marine boundary layer. The mean ε values show how sensitive the thermal inertial boundary layer can be to changes in wind, insolation, etc. They must be regarded as tentative, due to the subjective nature of the identification of turbulence categories, and the diverse data sources used in computations. In addition, it is of great interest to consider the numerical values of L_0 and L_N in Table 2. 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. (6) 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 3, we have summarized results of K_z from four available publications. Note that the reported values of K_z in Table 3 are directly comparable with our results in Table 2. 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 3

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

There are several possible sources of error in the above computations. Acceptable errors would have been encountered in all instances in any case because of satellite position in relation to the study area, and the size of an individual grid cell compared to the magnitude of possible error (Ernest and Lyons, 1974; Ernest, 1975). Some error may arise from assuming an inertial range spectrum to compute dissipation rate. Also, the vertical, lateral and longitudinal turbulence differed substantially from the assumption of isotropic turbulence. There are no direct atmospheric data against which to test these preditions.

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 we can categorize σ_y in the present analysis. Their effects (if any) remain obscure in the data scatter of individual data sets (Tables 1, 2). 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.

5 Concluding remarks

Measurements in the atmosphere even under the most favourable conditions could hardly be interpreted directly in any particular case, even if ε and the distribution of concentration could be measured accurately; because the turbulence is certainly not sufficiently uniform over its depth to permit uncritical application of a theory which depends on uniformity as well as local isotropy. The reason why Kolmogorov's analysis seems to apply so well to Richardson's synthesis of all observation of diffusion has been touched on by Batchelor and others. In spite of the importance of those diffusion coefficients for the quantitative treatment of atmospheric exchange processes, it must be stated that no theoretical approach attempting to interpret these highly variable factors by characteristic quantities of the turbulent and convective motions have become known up to now. Hence, we have to be satisfied with empirical data derived from these suitable measurements.

Finally, it is obvious that this study is a first attempt at 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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List of figures

Fig. 1 - The June 23, 1976 LANDSAT image at 1130 GMT of smoke plume of the Cabo Frio region drifting NE ward across the Atlantic Ocean. The air was so stable that the plume could be seen extending for 130 km with $u = 4.1\ m\ s^{-1}$ and $\frac{\partial \theta}{\partial z} = 5.6\ K\ km^{-1}$. But, over water fetch the air temperature was slightly colder than the ocean surface temperature. The formation of cumulus cloud streets are visible over the ocean in right side of photo. Image photographically enhanced from original 35 m m slide.

Fig. 2 - As in Fig. 1 except for 1205 GMT July 1, 1978. The air was so stable that the plume could be seen extending for 130 km with $u = 4.6\ m\ s^{-1}$ and $\frac{\partial \theta}{\partial z} = 2.7\ K\ km^{-1}$. A strong clockwise surface water current is visible on right side of the island.