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石田, 広史

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On the Stability Effect of Local Isotropy in the Atmospheric Boundary Layer over the Sea

Hiroshi ISHIDA

Abstract

This paper reports the estimation and analysis on the stability effect of the first Kolmogorov's hypothesis, local isotropy, in the atmospheric boundary layer from the data of the observation of AMTEX '74 which was done over the East China Sea. Local isotropy is estimated to depend on the atmospheric stability and the characteristic of the stability effect of local isotropy obtained from the data corresponded to the estimation.

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1. Introduction

In the practically atmospheric shear flow, the condition in which the turbulence becomes isotropic may be exerted influences by the atmospheric stability, roughness, and so on. The problem concerned with the first Kolmogorov's hypothesis, local isotropy, was partly investigated on the land by Kaimal et al. (1972)[1], while there is rarely the investigation concerned with this problem over the sea except the result of Pond et al. (1963)[2], which was neglected the effect of the atmospheric stability. It is necessary to know the characteristic of local isotropy in the atmospheric boundary layer over the sea. This reports the estimation and analysis from the data obtained in the observation of AMTEX '74 (Air Mass Transformation Experiment) with a relatively good accuracy.

2. Observation, Data Acquisition and Processing

The observation was done in the area, N28° E125°, by R. V. Keifu in the spring of 1974. During the observation she was drifting. All the sensors of meteorological entities were installed on the top of the foremast which was the position where the effects of the ship body on winds were small, this is certified by the results of the wind tunnel experiment by H. Ishida and Y. Mitsuta.[3] They were a sonic anemometer (Kaijo Electric Co.) and a fine thermocouple psychrometer. Each component of wind relative to the ship measured with them are corrected for ship motion by the method of Y. Mitsuta and T. Fujitani(1973)[4]. All the signals of fluctuating component were multiplexed into an FM signal with a multiplexer (TEAC, AU-151) and recorded on a magnetic tape recorder (TEAC, R-200). Further details of data aquisition and processing can be found in the reference(4).

3. Basic Analysis

The turbulent kinetic energy equation per unit density assuming horizontal uniformity, incompressibility, and neglecting molecular transports may be written as the following,

$$\frac{\partial \bar{e}_3}{\partial t} = -\overline{u'w'}\frac{\partial \bar{u}}{\partial z} + \frac{g}{\theta}\overline{\theta' w'} + 0.61g\overline{w'q'} - \frac{\partial \overline{e_3 w'}}{\partial z} - \frac{1}{\rho}\frac{\partial \overline{p'w'}}{\partial z} - \varepsilon \quad \cdots\cdots(1)$$

where, $e_3 = \frac{1}{2}(u'^2 + v'^2 + w'^2)$

u ; longitudinal wind velocity

v ; lateral wind velocity

w ; vertical wind velocity

g ; earth's gravitation

θ ; temperature

q ; specific humidity

ρ ; air density

p ; pressure

ε ; energy dissipation rate

prim and bar mean fluctuations from the mean value and the mean or covariance respectively.

In the surface layer, equation (1) may conveniently expressed in nondimensional forms,

$$\frac{k \cdot z}{u^3_*}\frac{\partial \bar{e}_3}{\partial t} = \phi_m - \frac{z}{L} - \frac{z}{L_q} - \frac{z}{L}\frac{\partial \phi_D}{\partial (z/L)} - \phi_\varepsilon \quad \cdots\cdots(2)$$

where, u_*; friction velocity

k ; Von Karman constant

L ; Monin-Obukhov length

L_q; Monin-Obukhov length for latent heat

ϕ_m and ϕ_ε are nondimensional functions and ϕ_D is the corresponding flux term. Equation (2) directly suggests the hypothesis that all the ϕ functions may be unique one of z/L, therefore ε is estimated to be a function of z/L too.

Extension of Monin-Obukhov scaling to the spectra of velocity leads to the assumption that the logarithmic spectra, $\kappa F(\kappa)$, should be the function of the reduced wave number, κz, and z/L only. Recently experimental analyses based on Kansas data show that Monin-Obukhov hypotheses are applicable to spectra.

In the local isotropy and the inertial subrange neglected the molecular viscosity, the relation for one-dimensional velocity spectra can be written,

$$F_V(\kappa) = F_W(\kappa) = \frac{4}{3}F_U(\kappa) = \frac{4}{3}\alpha \cdot \varepsilon^{\frac{2}{3}} \cdot \kappa^{-\frac{5}{3}} \quad \cdots\cdots(3)$$

where α is Kolmogorov's constant.

$F(\kappa)$ and ε are the functions of z/L in the natural atmosphere, so isotropic wave number, κ_i, in which equation(3) comes to balance may depend on z/L.

In the discussion to fllow, symbols and plots most commonly used in atmospheric work will be used.

$$\int_0^\infty F(\kappa)\,d\kappa = \int_0^\infty S(f)\,df = \overline{u'^2} \quad \cdots\cdots(4)$$

provided that we take Taylor's hypothesis, frozen turbulence, $\kappa = 2\pi f / \bar{U}$, so,

$$\frac{2\pi}{U} F_U\left(\frac{2\pi f}{\bar{U}}\right) = S_U(f)$$

$$x \cdot F_U(x) = f \cdot S_U(f)$$

4. Result and Discussion

The logarithmic spectra of longitudinal and lateral wind velocity in Fig. 1, $S_U(f)$ and $S_V(f)$, follow the $-\frac{5}{3}$ law. Even in the range of relatively low frequency, logarithmic spectra fit the $-\frac{5}{3}$ law, but the range does not always agree with the inertial subrange and equation (3) does not balance in such a frequency Spectra of wind velocity depended on the atmospheric stability just as the results of Kaimal et al. (1972)[1] and Busch and Larsen(1972)[5] in Kansas. Fig. 2 shows the nondimensional spectra of lateral wind velocity in the range of z/L between 0.05 and 0.6, −0.07 and −0.4.

$$\frac{f \cdot S_V(f)}{\sigma_V^2} = \frac{A \cdot f^*}{1 + \alpha \cdot f^{*\frac{5}{3}}}$$

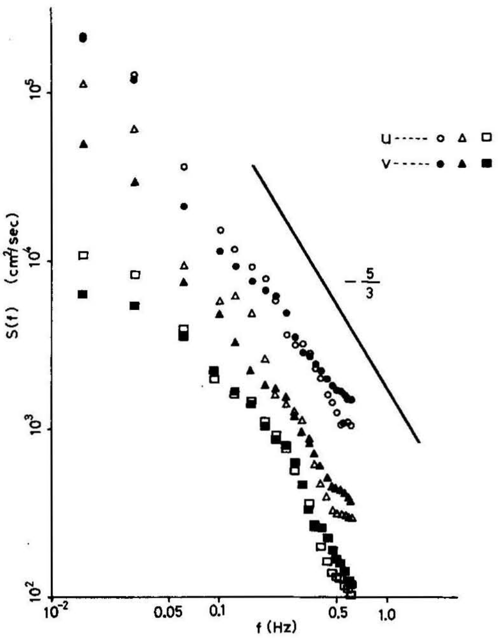

Fig. 1 The logarithmic spectra of longitudinal and lateral wind velocity

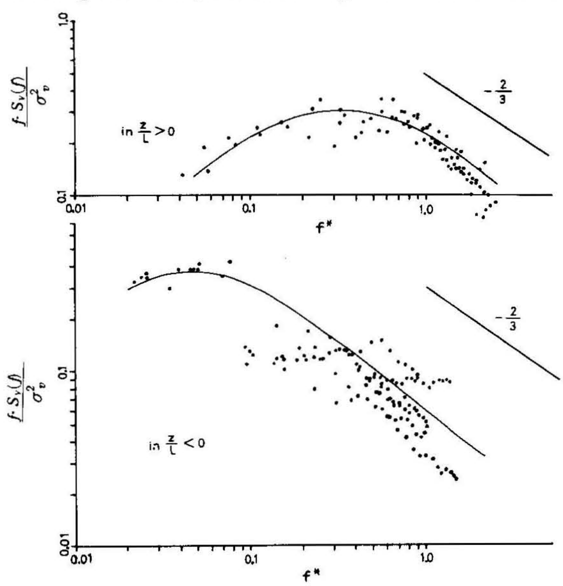

Fig. 2 The nondimensional logarithmic spectra of lateral wind velocity versus reduced frequency.

where, A and α ; constant

$f^* = \frac{fz}{U}$; reduced frequency

σ_v^2 ; variance of lateral wind velocity

Nondimensional logarithmic spectra have relatively well fitted with a universal function in the stable atmosphere, but they have been scattered in the unstable one. f_m^* is 0.3 in the stable atmosphere and 0.05 in the unstable one, which are in good agreement with the earlier results, where f_m^* is the reduced frequency for which $f \cdot S(f)$ has its maximum.

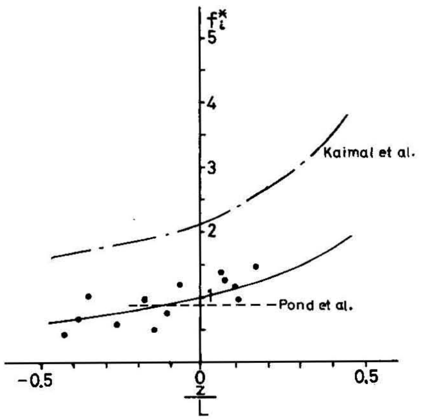

Fig. 3 The reduced frequency of local isotropy versus stability.

Fig. 3 shows f_i^* in which equation (3) comes to balance, where f_i^* is a reduced isotropic frequency, $f_i^* = f_i z / U$, and f_i is the lowest isotropic frequency. Pond et al. (1963)[2] obtained $2\pi f_i z / U = 4.5$, which is shown in Fig. 3 and has not the effect of the atmospheric stability. As shown in Fig. 3, it is noticed that f_i^* shows the same dependence on the atmospheric stability as the results of Kaimal et al. (1972)[1]. Although f_i^* is estimated from only the spectra of longitudinal and lateral wind velocity, in the range of istropic frequency, $\overline{u'w'} = -\overline{u'w'}$, so $\overline{u'w'} = 0$, that is, fluxes are 0, cospectra which showed momentum, heat, and water vapor fluxes showed nearly 0 in higher frequency than f_i^*. Therefore, it suggests that f_i^* generally depends on the atmospheric stability from the experiment too. But it is necessary to do more accurate measurement.

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