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RADIOLOGICAL AND ENVIRONMENTAL RESEARCH DIVISION ANNUAL REPORT

Atmospheric Physics

January—December 1978

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ARGONNE NATIONAL LABORATORY, ARGONNE, ILLINOIS

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Part IV

ARGONNE NATIONAL LABORATORY
9700 South Cass Avenue
Argonne, Illinois 60439

RADIOLOGICAL AND ENVIRONMENTAL
RESEARCH DIVISION
ANNUAL REPORT

Atmospheric Physics

January through December 1978

R. E. Rowland, Division Director
B. B. Hicks, Section Head

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Preceding Report: ANL-77-65 Part IV, January-December 1977

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FOREWORD

The programs addressed by this Section's research activities continue to evolve. In some cases changes occur slowly, but in others a casual reader might have difficulty in relating the reports presented here to specific projects as addressed in previous Annual Reports. For example, our work on dispersion over water near shorelines has grown to include a large component investigating the flux of pollutants to the surface. Likewise, our early studies of thermal exchange from heated water surfaces led to a current emphasis on steam fog and on the exchange of gases across an air-water interface. This latter program, focusing on environmental problems associated with the operation of industrial cooling ponds, is now in its final year.

Work on the optical and radiative effects of air pollutants commenced under the Multistate Atmospheric Power Production Pollution Study (MAP3S) and led to the operation of a network of atmospheric turbidity stations during 1977 and 1978. This network is no longer in operation; however, the work has led to a separately funded, broader investigation of the regional-scale effects of atmospheric haze.

The transfer of the MAP3S program from the Department of Energy to the Environmental Protection Agency took effect toward the end of 1978, with little immediate impact upon the scientific programs involved. We look forward to a period of closer cooperation with the EPA. Prior to this transfer, this Section had already profited from the receipt of EPA support for our studies of the dry deposition of atmospheric pollutants to natural surfaces. With the transfer of the MAP3S program, most of our work is now for the EPA.

Collaboration with our sister Ecological Sciences Section has continued to expand, especially since the delivery of their new research vessel, the RV Ekos. This unique research facility has made possible a series of cooperative experiments to measure the flux of small atmospheric particles into Lake Michigan by application of direct eddy-correlation methods. Much of the initial success of these studies is described in Part III of this 1978 series.

A highlight of the past year has been the success attained by several of our modeling programs. For example, our second-order closure schemes have been applied to a number of practical questions and are now being used in studies of flow over complex terrain. The Section's experimental data on the dry deposition of sulfur compounds have been combined with land-use information to produce a first estimate of deposition velocities for use in evaluations of pollutant transport, dispersion, and deposition over the region of interest to MAP3S. These same data have been averaged and combined with improved parameterizations of the wet removal process and of the depth of the mixed layer (both derived from studies conducted by this Section) in order to develop an Advanced Statistical Trajectory Regional Air Pollution (ASTRAP) model.

These models form the end product of much of our research. It is through their application that answers are best obtained to the difficult questions which confront our DOE and EPA sponsors and other regulatory and planning authorities. We look forward to a continued strong involvement in this field.

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RELATIONSHIPS OF ACOUSTIC AND OPTICAL C_n^2 TO SURFACE HEAT FLUX

M. L. Wesely and R. L. Coulter

Introduction

The measurement of C_n^2 and its relationships to sensible heat flux have been considered several times in previous annual reports and associated publications. The current intent, as explained elsewhere in this report (Coulter and Wesely, 1978), is to use acoustic and optical methods to infer spatial averages of sensible heat flux, mostly in support of experimental studies of the planetary boundary layer (PBL). It is anticipated that such work will eventually take place over complex terrain, where conventional, point-sampling techniques for measuring sensible heat flux are quite difficult. In this report, we assume that the acoustic parameter C_{na}^2 and the optical parameter C_{no}^2 can be successfully measured at known altitudes in the lower PBL. The task is to relate these parameters to the surface sensible heat flux H during unstable and slightly stable conditions in the lower atmosphere.

Interpretation of C_n^2

The parameter C_n^2 , the refractive index structure function coefficient, is a measure of the magnitude of the variance of refractive index fluctuations at the spatial scales or frequencies associated with the inertial subrange of the variance spectrum. In the present method of measurement, C_n^2 is a function of temperature fluctuations and, to a lesser extent, humidity fluctuations. Thus, C_n^2 can be expressed in terms of the corresponding temperature and humidity structure parameters, C_T^2 and C_e^2 . In the present discussion, C_e^2 is not mentioned explicitly but is included through the Bowen ratio of sensible to latent heat flux in the surface boundary layer. That is, the expression $C_T \bar{p} / C_e \bar{T} = 5\beta$ suggested earlier (Wesely and Alcaraz, 1973), where $\bar{p}$ is the mean atmospheric pressure and $\bar{T}$ is the mean air temperature, is assumed valid. With this relationship the analysis can be carried further to show that acoustic C_n can be related to C_T as

$$C_{na} = C_T (2\bar{T})^{-1} [1 + (0.06/\beta)^2]^{1/2}, \quad (1)$$

and optical C_n can be related to C_T as

$$C_{no} = C_T A_1 \bar{p} \bar{T}^{-2} [1 + (0.03/\beta)^2]^{1/2} \quad (2)$$

(Wesely, 1976b), where $A_1 = 78.7 \text{ K mb}^{-1}$ is a refractive index constant for light. The factors in the brackets take into account the effect of humidity fluctuations on C_n . These "correction" factors are derived with the assumption that at the small spatial scales of refractive index inhomogeneities that are associated with the scattering of acoustic radiation and light, the temperature and humidity fluctuations are uncorrelated. Field measurements indicate that a correlation coefficient of zero is indeed appropriate at these scales (Wesely and Hicks, 1978).

Equations 1 and 2 are easily rearranged to find the needed expressions for C_T , which is now to be related to H . As reviewed by Tatarskii (1971), a large amount of work has been done describing the structure parameter for both forced and free convection in the atmospheric surface layer. Similarity theory leads to a convenient way of expressing C_T (Obukhov, 1960);

$$C_T = T_* z^{-1/3} f(z/L), \quad (3)$$

where z is the height above the surface and $T_* = H/(\rho c_p u_*)$ is a temperature scaling parameter (ρ is air density, c_p is the specific heat of air at constant pressure, and u_* is the friction velocity). The function $f(z/L)$ is an empirical relationship dependent upon stability expressed in terms of z/L , where L is the Obukhov length scale,

$$L = -\rho c_p u_*^3 \bar{T} / (kgH), \quad (4)$$

where $k = 0.4$ is the von Karman constant and g is the acceleration due to gravity. With knowledge of $f(z/L)$, $\bar{T}$, u_* , and β , the sensible heat flux can be evaluated from C_n with the use of Eqs. 1 or 2 and Eqs. 3 and 4.

Inclusion of moisture effects in Eqs. 1 and 2, however, implies that the effects should also be considered in Eqs. 3 and 4. The contribution of

water vapor flux to buoyancy effects can be included in Eqs. 3 and 4 by replacing H with $H(1 + 0.07/\beta)$ (e.g., Busch, 1973). Equation 3 may be further revised by replacing T with the virtual temperature $T_v \approx T/(1 - 0.378e/p)$, where e is the water vapor pressure, so that C_{T_v} replaces C_T . These substitutions, made with the assumption that the scaling arguments which apply to temperature fluctuations also apply to density fluctuations, result in

$$C_{T_v} = T_* (1 + 0.07/\beta) z^{-1/3} f(z/L_B) , \quad (5)$$

where the total-buoyancy Obukhov length scale is

$$L_B = -\rho C_p u_*^3 \bar{T} / kgH(1 + 0.07/\beta) . \quad (6)$$

An expression for C_{T_v} similar to those for C_n given by Eqs. 1 and 2 can be derived by procedures similar to those described earlier (Wesely, 1976b). That is, by expanding T , e , and p into mean and fluctuating components in the relation for T_v , squaring and averaging the result, and making several simplifying assumptions, it can be shown that an appropriate expression is

$$C_{T_v}^2 = C_T^2 [1 + r_{eT} (0.756 C_e \bar{T} / (C_T \bar{p})) + (0.378 C_e \bar{T} / C_T \bar{p})^2] , \quad (7a)$$

where r_{eT} is the correlation coefficient of e and T at the spatial scales associated with the scattering of acoustic and electromagnetic energy. If, as is presently assumed, the value of r_{eT} is zero, Eq. 7a reduces to

$$C_{T_v} = C_T [1 + (0.07/\beta)^2]^{1/2} \quad (7b)$$

Substitution of Eq. 7b into Eq. 5 results in

$$C_T = T_* z^{-(1/3)} f(z/L_B) (1 + 0.07/\beta) [1 + (0.07/\beta)^2]^{-1/2} . \quad (8)$$

If r_{eT} had the value of 1.0, or if β were extremely large compared to 0.07, Eq. 8 would be identical to Eq. 3. It is noteworthy that this improved formulation was not used in an earlier study of optical remote sensing using blurring, with the result that the latent heat flux above an industrial cooling pond was probably underestimated by an average of 10% (see Table 3 of Wesely, 1976a).

For the present optical data (Coulter and Wesely, 1978) collected in unstable conditions ($0 > z/L > -1.5$), H is found by an iterative technique using Eqs. 2, 6, and 8. The value of u_* is found from eddy-correlation measurements, and β is estimated as $H/(0.93 R_n - H)$, where $0.93 R_n$ represents the fraction of the net radiation that is partitioned into sensible and latent heat flux. The function $f(z/L)$ is calculated as suggested by Wesely and Alcaraz (1973), and agrees within 5% of the values given by Wyngaard et al. (1971).

For strongly unstable conditions ($z/L > -1.5$), C_T is not directly dependent on u_* (Obukhov, 1960; Wyngaard et al., 1971), so that the resulting estimate of H is not dependent on the eddy-correlation measurements of u_* . This is usually the case for sodar measurements but occurs less often for optical line-of-sight measurements. Here we employ the formula of Wyngaard et al. (1971)

$$C_T^2 = 2.68 (\bar{T}/g)^{2/3} (H/\rho c_p)^{4/3} z^{-4/3} , \quad (9)$$

where the modification, as previously discussed, of replacing H with $H(1 + 0.07/\beta)$ and C_T with the expression for C_{T_V} given by

$$C_T^2 = 2.68 (\bar{T}/g)^{2/3} (H/\rho c_p)^{4/3} z^{-4/3} (1 + 0.07/\beta)^{4/3} [1 + (0.07/\beta)^2]^{-1} . \quad (10)$$

The surface heat flux H_{so} from sodar information can be determined by combining Eqs. 1 and 10:

$$H_{so} = 0.48 \rho c_p (g/\bar{T})^{1/2} (C_{na}/2\bar{T})^{3/2} z \gamma_{so} , \quad (11a)$$

$$\gamma_{so} = \{ [1 + (0.07/\beta)^2] / [1 + (0.06/\beta)^2] \}^{3/4} / (1 + 0.07/\beta) . \quad (11b)$$

Likewise, H_{sc} from the optical scintillation measurements can be found from optical C_n as

$$H_{sc} = 0.48 \rho c_p (g/\bar{T})^{1/2} [C_{no} \bar{T}^2 / (A_1 \bar{p})]^{3/2} z \gamma_{sc} , \quad (12a)$$

where

$$\gamma_{sc} = \{ [1 + (0.07/\beta)^2] / [1 + (0.03/\beta)^2] \}^{3/4} / (1 + 0.07/\beta) . \quad (12b)$$

Obviously, the factors γ_{so} and γ_{sc} take into account the effects of humidity fluctuations on the calculation of H. Figure 1 shows these effects are quite significant, especially if β is less than about 0.2. Note that for $\beta > 0.4$, the differences of either γ_{so} or γ_{sc} from 1.0 are mostly due to the term $(1 + 0.07/\beta)$ on the right-hand sides of Eqs. 11b and 12b. In fact, a first approximation for γ_{so} for all values of β is

$$\gamma_{so} = (1 + 0.07/\beta)^{-1} . \quad (13)$$

This approximation is not dependent upon the correlation between T and e. Thus, the estimates of heat flux from the sodar measurements are almost independent of the assumption made with regard to the extent of T and e correlation. This is not the case for the optical measurements, however, except in the trivial case where β is large.

FIG. 1.--Values of factors that take into account humidity in Eqs. 11b and 12b, plotted as a function of Bowen ratio. (ANL Neg. 149-79-77)

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SODAR CALIBRATION METHOD

D. L. Sisterson and R. L. Coulter

Introduction

Acoustic remote probing systems (sodars) have been used in most of the planetary boundary layer (PBL) experiments conducted by this laboratory in recent years. In usual applications, they have yielded pictorial representations of the structure of the lower atmosphere, from which mixing layer heights can be evaluated. Since digitization of the graphical records is time-consuming and arduous, a number of exploratory attempts to quantify sodar returns automatically have been made. It is also possible to derive useful information from the intensity of returned signals, this being a measure of the acoustic refractive index structure function C_n^2 . For this purpose, it is not only required that the returned signals be digitized and processed automatically, it is also necessary that the transmitting and receiving characteristics of the apparatus be thoroughly understood. In this report, the methods employed to calibrate an ANL sodar will be described. In the following articles (Coulter, 1978), results will be discussed, and some preliminary applications of the calibrated systems will be reported (Coulter and Wesely, 1978).

Experimental Procedure

Calibration is accomplished by determining the total output power of the sodar transmitter and the sensitivity of the sodar receiving antenna in terms of sound-pressure levels. While the sensitivity of the receiver can be estimated directly, the total output power of the sodar is dependent upon the directivity or beam pattern of the antenna (Coulter, 1978). To determine the beam pattern a microphone is positioned at various locations relative to the transmitting antenna and is used to measure a continuous signal produced through the sodar transmitter at its usual operating frequency, 1600 Hz. In the calibration procedure used at Argonne, a microphone was suspended from a carriage on a cable between a 44.5 m tower and a 23 m pole (see Figure 1),

FIG. 1.--Photograph of sodar calibration site .

high enough above the sodar to be in the far field and thus, high enough to eliminate short-range perturbations of the acoustic pressure field.

It is necessary that all measurements be made in a vertical plane through the center of the sodar receiving antenna, since the beam width is narrow ($< 10^\circ$ at 1600 Hz) and the intensity of the signal falls off sharply at small angles away from the center of the beam. The beam pattern, a function of the size and shape of the receiving antenna and the noise-reducing enclosure, is assumed to be axially symmetric and can therefore be "mapped out" by traversing the microphone across the transmitting antenna along only one direction. Once the shape of the beam pattern is determined, the total output power of the sodar can be evaluated by integrating the sound-pressure level or

intensity of the signal over the surface area of the beam pattern.

The sensitivity of the sodar receiver is determined by mounting a speaker on the tram and placing the microphone near the receiver. A signal generator and amplifier are used to produce a continuous 1600 Hz signal through the suspended speaker while the sodar is operating in a continuous receiving mode. The sensitivity of the sodar receiver is derived as the ratio of sound-pressure level readings of the receiver and the microphone. If the total output power of the transmitter and the sensitivity of the receiving antenna are known, the intensity of the return signal can be compared directly to the intensity of the transmitted signal.

Determination of the Beam Pattern

The beam pattern can be characterized in terms of θ , the angle of departure from the vertical axis, and r , the distance from the sodar antenna, since the intensity of the output signal is inversely proportional to the square of this distance. In the Argonne calibration tests, a double-theodolite system was used, as illustrated schematically in Figures 2 and 3. The angle θ cannot be measured directly and must be determined within the framework of known measurements shown in Table 1.

FIG. 2.--Schematic of sodar calibration method with non-zero departure angle.

h_o	= 1.66 m	a	= 63.5 m
ω	= measured directly	δ	= 20.0°
χ	= measured directly	ψ	= 42.46°
ξ	= measured directly	η	= 3.66°
σ	= 49.84°	ϕ	= 3.58°
β	= 84.77°	n	= 1.83 m
α	= 7.15°	γ	= 44.38°

$$\theta = \cos^{-1} \left(\frac{h}{r} \right) \quad , \quad (1)$$
$$r = (f^2 + g^2 + h^2)^{1/2} \quad (2)$$
$$h = c \tan \omega + h_0, \quad (3)$$

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for e, we find

$$e = \frac{a \sin \chi}{\sin \{\pi - (\chi + \xi)\}}$$

where a (64.5 m) is the baseline distance between theodolites. Azimuth angles χ and ξ (taken at the same microphone position as ω) are measured directly. Therefore, h can be determined for each position of the microphone.

To solve Eq. 2, f and g must be determined. From Figure 2

$$g = e \sin \{\tau + (\sigma - \xi)\} - i, \quad (4)$$

and

$$f = e / \cos \{\tau + (\sigma - \xi)\} + k, \quad (5)$$

where σ (49.84°) is measured directly.

τ (Figure 3) is determined by

$$\tau = \pi/2 - \nu - \{\sigma - (\beta - \alpha)\}, \quad (6)$$

where α (71.5°) and β (84.77°) are measured directly by theodolite 1. The angle ν is determined by solving two equations with two unknowns:

$$\frac{d}{\sin \sigma} = \frac{b}{\sin \{(-\nu - \alpha) - \mu\}}, \quad (7)$$

where σ (20.0°) is the azimuth reading of theodolite 2; and

$$\frac{d}{\sin \alpha} = \frac{c}{\sin \nu}. \quad (8)$$

The angle μ is found from

$$\mu = \pi - (\psi + \beta),$$

where ψ (42.46°) is the azimuth reading by theodolite 2. Therefore, $\mu = 52.77^\circ$.

The distance between theodolite 2 and the pole (length b in Figure 4) is determined from

$$b = \frac{(n-m)}{\tan \eta}$$

and

$$m = b \tan \phi,$$

FIG. 4.--Schematic of determination of the distance between the pole and theodolite 2.

where $n = 1.83$ m, $\eta = 3.66^\circ$, and $\phi = 3.58^\circ$. Solution of these equations yields $b = 14.5$ m. Length c (Figure 3) can be found from

$$c = \frac{a \sin \psi}{\sin \mu}$$

and $c = 53.8$ m. Equations 7 and 8 thus yield $\nu = 51.36^\circ$, which in turn gives $\tau = 207^\circ$ (Eq. 6).

The lengths i and k can be obtained from the special case of $\tau = \xi$ (the microphone directly over the sodar); equations 4 and 5 then become

$$i = \ell \sin \tau, \quad (9)$$

and

$$k = \ell \cos \tau, \quad (10)$$

where ℓ is given by

$$\ell = \frac{a \sin \gamma}{\sin \{ \pi - (\gamma + \sigma) \}}.$$

The above equations can now be solved, and θ and r can be calculated (Eqs. 1 and 2, respectively) for successive positions of the microphone.

The signal received by the microphone is sent directly through an rms detector and then digitized and averaged, using a Digital Equipment 1134 mini-computer. From these data and the position data derived from the theodolite measurements, the beam pattern can be determined directly.

Determination of Receiver Sensitivity

To calculate the receiver sensitivity, the microphone on the tram is replaced by a speaker and a 1600 Hz signal is transmitted. The sound-pressure level of this signal is converted to a DC signal by the sodar receiver and is digitized, averaged by the minicomputer, and stored on magnetic tape. A calibration curve relating the sound-pressure level to the signal output of the receiver can be produced by varying the transmitted acoustic output level.

Total Output Power of the Sodar

It is necessary to know the absolute sound-pressure level along the axis of the antenna (see Coulter, 1978) in order to evaluate the total output power. With the microphone suspended directly above the sodar (using the two theodolites for accurate positioning), the sodar transmitter is used at the maximum pulse duration (200 ms). The signal level measured by the calibrated microphone is digitized at 100 Hz and averaged to determine the absolute sound-pressure level.

Therefore, the total output power of the sodar transducer and the sensitivity of the sodar receiver, both measured in terms of sound-pressure levels, are obtainable. In fact, most of the calculations discussed in this report need to be performed only once, thereby permitting precise position determination of the microphone and operation by as few as two people.

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SODAR CALIBRATION RESULTS

R. L. Coulter

The use of acoustic sounders (sodars) to obtain quantitative measurements of atmospheric turbulence parameters (e.g., temperature and velocity structure parameters) necessitates the calibration of the total transmitting and receiving apparatus; that is, one must be able to relate the voltage level of the receiver output to the actual acoustic pressure level resulting from back-scattering of the transmitted acoustic wave and be able to determine the actual output of acoustic energy. During the past year, this calibration has been carried out successfully with the ANL sodar. The theory and results of this project are discussed in this paper.

The transmitter emits a pulse of acoustic energy at a frequency of 1600 Hz for a known duration τ ($\tau = 50, 100, \text{ or } 200 \text{ ms}$, corresponding to a length of approximately 17, 34, or 68 m). At a location in space sufficiently distant to be within the far field of the transmitting antenna, the acoustic power density $W(r, \theta, \phi)$ is related to the peak power output (P_o) and to the directivity ($F(\theta, \phi)$) of the antenna through the relationship (Kristenson, 1978)

$$W(r, \theta, \phi) = P_o F(\theta, \phi) / r^2, \quad (1)$$

where r is the distance to the point in question and θ and ϕ are as in Figure 1. That is to say, both the antenna (a parabolic dish in this case) and the noise limiting enclosure modify the pressure field produced by the transducer. The function $F(\theta, \phi)$ must be less than or equal to 1; for the ANL parabolic dish antenna, $F=1$ when $\theta=0$. The total acoustic energy output of the system can then be written

$$\begin{aligned} E &= v_p^{-1} \int_{r_1}^{r_1+\ell} dr \int_0^{2\pi} r d\phi \int_0^\pi d\theta W(r, \theta, \phi) r \sin(\theta) \\ &= 2\pi(\ell/v_p) P_o \int_0^{\pi/2} d\theta F(\theta) \sin(\theta), \end{aligned} \quad (2)$$

FIG. 1.--Coordinate system and variables used relative to the antenna enclosure.
(ANL Neg. 149-79-214)

where ℓ is the pulse length and v_p is the phase velocity of sound (340 m/s). We have assumed that $W(\theta, \phi)$ is axially symmetric about $\theta=0$ (i.e., W is not a function of ϕ ; hence, the 2π term from integration), and is zero for θ greater than $\pi/2$.

At any point in space, a certain portion of the incident energy is scattered in any given direction. The system under discussion has one transducer acting as both transmitter and receiver; thus all the energy received has been scattered through an angle of 180° , if possible refraction effects are ignored. The amount of backscattered energy is the product of the output power density $\beta(\pi)$, the acoustic scattering cross section at 180° . The amount which is detected by the receiver will once again be a function of the directivity of the antenna and enclosure. A certain proportion of the energy $[1 - \exp(-2\alpha r)]$, where α =attenuation coefficient] will be absorbed by the atmosphere during propagation to and from the scatterer. We can express the power received at a given instant from a certain point in space as

$$\delta P(\theta, r, \phi) = W(r, \theta) \beta(\pi) (A_R/r^2) F(\theta) e^{-2\alpha r} \delta V, \quad (3)$$

where δV is the volume element at the position in question, and A_R/r^2 is the receiver aperture. To obtain the total power received at a given instant we must integrate over the length, ℓ_v , of the volume contributing to the signal at any instant ($\ell_v = \ell/2$ for a monostatic system). This length is defined by the locus of points swept out by the forward edge of the pulse as it returns toward the receiver until such time as it meets the outward bound trailing

edge of the pulse. Then,

$$p(r) = \exp(-2\alpha r) \int_{r-\ell_v/2}^{r+\ell_v/2} \int_0^{2\pi} r d\phi \int_0^{\pi/2} (P_o/r^2) F(\theta) (A_R/r^2) r \sin(\theta) d\theta \quad (4)$$

$$= 2\pi P_o \beta(\pi) e^{-2\alpha r} A_R \ell_v / [r^2 (1 - \ell_v^2/(4r^2))] \int_0^{\pi/2} F^2(\theta) \sin(\theta) d\theta \quad (5)$$

$$\simeq 2\pi A r^{-2} \int_0^{\pi/2} d\theta \sin(\theta) F^2(\theta) \quad (6)$$

where

$$A = \ell_v \beta(\pi) e^{-2\alpha r} A_R P_o .$$

The term A_R is difficult to evaluate separately. However, for the receiver as a whole, the output voltage is proportional to the amplitude of the received signal amplitude. Since A_R is constant,

$$P(r) r^{-2} A_R^{-1} = E^2 C_a , \quad (7)$$

where C_a is a calibration constant that can be determined by measuring the voltage E when a known acoustic signal level is present. Thus, three calibration constants need be determined: P_o , the power output by the transmitter along the antenna boresight ($\theta=0$); $F(\theta)$, the directivity or beam pattern of the antenna-enclosure combination; and C_a , the receiver calibration constant. The actual procedure for the determination of these parameters is discussed by Sisterson and Coulter (1978).

In practice, $F(\theta)$ cannot be reliably determined over the complete half space, because of equipment limitations. However, as θ increases, $F(\theta)$ decreases much more rapidly than $\sin \theta$ increases.

Hence, one need measure $F(\theta)$ through only 10 or 20 degrees to permit calculation of the last term in Eq. 6. Figure 2 illustrates the beam patterns of the two ANL antenna enclosures. While the two curves are similar in their gross features, the beam pattern for antenna 1 (Figure 2a) is seen to be narrower than that for antenna 2 (Figure 2b). The half-width (θ such that $F(\theta) = 0.5 F(0)$) for antenna 1 is 5° , while that for antenna 2 is 6° . The differences are due mainly to the differences in enclosures. The enclosure

FIG. 2.--Directivity for the two Argonne sodars. Relative signal strength in dB increases outward (signals were normalized to the signal at $\theta=0$). Positive angles (θ) are for $\phi=0$, negative angles are for $\phi=180^\circ$. (ANL Neg. 149-79-213)

sides of antenna 1 slope outward with height, while the sides of antenna 2 are vertical and about two feet shorter. The larger effective diameter of antenna 1 leads to a smaller beam width. Although $F(\theta)$ is probably not precisely symmetric (regular polygons) about $\theta=0$, the departure will have only a small effect on the final calibration. An indication of the amount of departure from axial symmetry can be seen in the differences in $F(\theta)$ for ϕ equal to 0 and 180° .

The parameter C_a can be determined from the intercept of the plot of $\log(E^2)$ versus $\log$ of the sound pressure level. The value of P_0 is determined as outlined elsewhere (Sisterson and Coulter, 1978). The values for P_0 and C_a are listed in Table 1 for the ANL sodar 1.

The directivity is a function of the overall geometry of the antenna and enclosure. It will not change with time or meteorological conditions and need be measured only once. The other two parameters are more variable (particularly with the present receiver), which necessitates re-evaluation of those constants under different temperature conditions or when the electronics are changed.

Table 1. Calibration constants and operating output power for the ANL sodar.

$\text{Log}_{10}(C_a)$	-9.73 ± 0.3
Peak power output (P_o)	0.135 ± 0.05 watts
Total output power (energy/time)	5.22 watts (acoustic)
Beamwidth	$\pm 6^\circ \pm 0.5^\circ$

Whenever the transducer is replaced, a recalibration must be done. As indicated in Table 1, C_a and P_o are the most uncertain constants. This is due in part to the natural variability of acoustic signals, but to a greater extent it is due to the limited range of the receiver. With only a linear receiver for signals that vary logarithmically for the most part, it is difficult to establish the intercept in Eq. 7. An additional source of difficulty is the relatively wide frequency bandwidth of the receiver, which limits the range of known signals for determination of C_a . Nevertheless, these obstacles to calibration have been overcome, and values of atmospheric turbulence variables are now determinable.

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A METHOD FOR THE AUTOMATIC ESTIMATION OF SURFACE HEAT FLUX

R. L. Coulter

Introduction

The flux of heat (H) into the lowest part of the atmosphere from the surface of the earth during daytime is a major determinant in the evolution of the well mixed layer, throughout which pollutants are diffused, as well as a dominant force in the development of convection. For estimates of the surface boundary conditions in nonhomogeneous terrain, spatially averaged values are desirable. Finally, parameterization of heat flux requires a statistical knowledge of its variation with season, time of day, cloudiness, turbidity, etc.

Remote sensing techniques (sodar, in particular) for the estimation of surface heat flux satisfy many of these requirements. Because the sodar can be operated in a nearly continuous fashion, and because the signal return is a function of a volume average (see Coulter, 1978; Wesely and Coulter, 1978), real time estimates of heat flux averaged over time and space can be obtained and conveniently stored for statistical analysis.

This report details the development of a real-time continuous monitoring system, using the ANL sodar and a minicomputer, which can provide estimates of the surface heat flux under certain theoretical constraints, as well as provide an indication of the reliability of those results.

Theory

The details concerning the theoretical relationship between an acoustic signal return and heat flux can be found elsewhere (Wesely and Coulter, 1978; Coulter, 1977). The highlights will be touched upon here. The acoustic scattering cross-section ($\sigma(z)$) is related primarily to the temperature structure parameter (C_T^2) for acoustic energy scattered through 180° by

$$\sigma(z) = A C_T^2, \quad (1)$$

where A is a function of the air temperature and acoustic energy wavelength. This cross-section is, in turn, proportional to the signal strength detected by the sodar. A system calibration (Coulter, 1978; Sisterson and Coulter, 1978) permits one to estimate the actual magnitude of C_T^2 .

When the atmosphere is very unstable, the surface heat flux can be related to the temperature structure parameter by

$$C_T^2 = B(H/z)^{4/3} f(\beta) , \quad (2)$$

where B is a function of the average surface temperature, $f(\beta)$ is a relatively complicated but known function of the Bowen ratio (β), and z is the height above the surface. As a first-order approximation, one can assume $f(\beta) = 1$, particularly for larger values of β . In this case, then, we see that a measure of C_T^2 via Eq. 1 allows an estimate of H through Eq. 2, so long as the free convection assumption is valid. This can be determined by investigating the dependence of C_T^2 upon height. If a $-4/3$ power dependence is observed, we assume free convection conditions, and the value of heat flux can then be found by iteration.

Method

Data from the sodar are sampled and digitized at a 68 Hz rate by a minicomputer and analog-to-digital converters. This rate is equivalent to sampling every 2.5 m as the acoustic pulse moves upward in the atmosphere. These samples are then averaged, on-line, to a user selectable height range. This time series is also averaged with a user selectable number of subsequent acoustic pulses and combined with the appropriate calibration constants to form an averaged vertical profile of C_T^2 ; i.e.,

$$\overline{C_T^2(z_k)} = n(2\Delta z/5)^{-1} \sum_{i=1}^N \sum_{j=k-\Delta z/5+1}^{k+\Delta z/5} C_T^2(z_j, t_i) , \quad (3)$$

where k is the number of the sample, Δz is the height interval over which the data are averaged, and N is the number of pulses over which the data are averaged. A 30 min average requires that $N = 100$ and the height interval is

usually 5 or 10 m.

After this series is formed, $\log(C_T^2(z_k))$ is calculated and displayed versus $\log(z_k)$ to provide a visual check of the data. A least squares fit is made to the data between the heights of 40 and 300 m. The calculated slope of this line is compared to $-4/3$; if the slope is within -1.33 ± 0.05 , the data are assumed good, and the value of H is calculated, using Eq. 2. If the slope is not within -1.33 ± 0.05 , the process is repeated with the maximum height decreased in steps of 10 m until either an acceptable fit is found or the maximum height is 60 m. If no fit has been determined acceptable, the height interval with slope closest to 1.33 is used to approximate H . The value of H , the maximum height used, the slope calculated, and C_T^2 at 40 m are then printed on the plot, as well as the beginning and end times for the data recorded (Figure 1). This whole process usually takes less than 10 sec. For permanent storage purposes, the time, C_T^2 (40 m), H , slope, maximum height, and correlation coefficient of the data are written on a disk. If the user desires, the averaged time series ($C_T^2(z_k)$) can also be stored.

FIG. 1.--Logarithmic profile of temperature structure parameter made on-line during data collection on Sept. 7, 1978. The straight line is a reference $-4/3$ slope corresponding to 100 W/m^2 . (ANL Neg. 149-79-28)

The value of the correlation coefficient is quite useful in determining the quality of the data used to calculate the heat flux. Estimates of H with small correlation coefficients can be ignored or treated in a special manner. This is of some benefit in circumstances in which an estimate of H every 30 min is useful, since often the atmosphere changes character considerably within this time interval. Storage limitations of the minicomputer used in the initial development do not permit a continuous running mean to be calculated; however, there is enough storage to permit the division of the averaging time into as many as five segments. That is, with a 30 min averaging time, we can produce a 30 min average every 6 min. This procedure has been incorporated into the data processing in such a manner that both the basic averaging time and the number of divisions within that time are selected by the user.

Figure 2 is a plot of heat fluxes calculated from data of Sept. 7, 1978; a value of 0.2 has been assumed for β in these estimates. In reality, β changes with time and should be calculated separately for each estimate. A measure of net radiation is needed for this determination; this value can be easily included in the data sampling routine to enhance the computations and provide better on-line estimates.

These data can be gathered and heat flux estimates obtained automatically. With such an operation, a large amount of data can be gathered and reduced quite easily. The advent of microprocessors makes the use of such a system in remote locations quite possible. There are, however, some improvements which offer promise of better heat flux estimates. These include (1) using a more exhaustive description of the structure parameter profile,

FIG. 2.--Heat flux vs. time of day on Sept. 7, 1978. A value of $\beta=0.2$ has been used in the calculation of these estimates. Values represent 30 min averages, calculated every 6 min. (ANL Neg. 149-79-215)

such as varying the lowest point as well as the highest, (2) employing measurements of net radiation for the estimation of β through iteration techniques, and (3) allowing variations of β with height. All these improvements can be accomplished on-line with software manipulations.

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A COMPARISON OF THREE METHODS FOR MEASURING SURFACE HEAT FLUX

R. L. Coulter and M. L. Wesely

Introduction

Surface heat flux estimates usually are derived from point measurements of temperature and wind velocity in the atmospheric surface layer. In order to have wide applicability, they must be taken over homogeneous terrain. For problems in the developing atmospheric boundary layer (pollutant dispersion, e.g.) in non-homogeneous terrain, areally averaged measurements of heat flux are more appropriate. Possible means for obtaining measurements include remote sensors such as sodars or lasers, for they are sensitive to energy (acoustic or electromagnetic) scattered during propagation over large distances through the atmosphere. Under certain atmospheric conditions, a sodar or a laser scintillation device can be used to determine the temperature structure parameter, which in turn can be related to the surface heat flux. Measurements of surface heat flux during June and July 1978, determined from sodar, laser scintillation and eddy-correlation techniques are compared in this report.

Discussion

Sodar measurements of backscattered radiation can be related to the acoustic refractive index structure parameter C_n^2 with a properly calibrated system (e.g., Clifford, 1972; Hall, 1972), and measurements of laser scintillation can be used to estimate optical C_n^2 (Clifford, 1972; Lawrence, 1972). The theoretical basis for relating these structure parameters to the temperature structure parameter C_T^2 and the surface heat flux H is described elsewhere in this report (Wesely and Coulter, 1978). We discuss here only the experimental procedure and results. The experiment was performed at ANL during seven days in early summer, 1978. The terrain is slightly rolling with trees covering about half the otherwise grassy site. The primary location for the sensors, a large field approximately 400 m square, included a 45 m meteorological tower. The sodar was located about 60 m southwest of the tower,

pointing vertically. The scintillation equipment was deployed in two ways: first, with transmitter at a height of 37.5 m on the tower and the receiver 1.7 km northeast, 15.5 m high on a building; and second, with the receiver 238 m north, 0.8 m high. The eddy correlation gear was mounted 11 m above the surface on a boom extending 4 m from the southwest corner of the tower, necessitating a choice of winds from SE to N (clockwise) to avoid unrepresentative wind measurements due to tower interference.

Simultaneous measurements of sensible heat and moment flux, line integrals of optical structure parameters (C_n^2) weighted over paths within the surface layer, and volume averages of acoustic C_n^2 in the lower planetary boundary layer (PBL) (from 40 to 500 m) were made. Both the eddy-correlation and scintillation results were continuously recorded on strip charts, while the sodar signals were digitized for analysis by computer. For comparison purposes, the data were manipulated to produce 30-min averages commencing on the hour. Additional supporting data consisted of continuous samples of wind speed and direction at 45 m, wind speeds and temperatures at 6 and 1.5 m heights, and net radiation over grass.

Results

Figure 1 depicts the heat flux estimates determined by each of the three methods for the seven days. There is general agreement in magnitude, with differences of a factor of two or more occurring rarely. More importantly, the trends agree quite well. For example, the three estimates show excellent agreement during a rapid order of magnitude change in heat flux on June 14, and all three estimates show a distinct decrease in heat flux near 0930 on July 11.

Scatter diagrams of the heat flux determined with sodar and scintillation techniques versus that from eddy correlation are shown in Figures 2a and 2b, respectively. It is immediately obvious that there is more variability in the sodar vs. eddy correlation. In both cases, however, the points vary about the 45 degree line, indicating lack of bias. One must remember that the sodar is sensitive to processes occurring above the surface layer, while

FIG. 1.--Daily variation of heat flux estimates determined via eddy correlation (●), scintillation (□.... forced convection, o.... free convection) and sodar (x) during June and July 1978.

FIG. 2.-- Scatter diagrams of sodar (a) and scintillation (b) estimates of heat flux versus eddy correlation heat flux estimates. Forced convection scintillation estimates are denoted by $\square$ and free convection estimates by $\circ$. (ANL Neg. 149-79-75 and 149-79-74)

FIG. 3.--Daily variation of the ratio of sodar (a) and scintillation (b) to eddy correlation heat flux estimates. $\bullet$ 14 June; $\times$ 16 June; $\circ$ 20 June; $+$ 21 June; $\diamond$ 22 June; $*$ 23 June; Δ 11 July. (ANL Neg. 147-79-79 and 149-79-80 Rev. 1)

the other techniques involve measurement within the surface layer. This may account for some of the variability. An estimate of the relative accuracy of the methods can be made by calculating the mean ratio of simultaneous heat flux determinations. Taking the logarithm of the ratios before averaging helps reduce large variations which can occur for small heat flux values. When this is done, the average ratio of sodar to eddy-correlation heat flux is 1.08, while the ratio of scintillation to eddy correlation heat flux is 0.93 in very unstable conditions and 1.08 in near-neutral conditions. Figure 3 shows the daily variation of this ratio. The large ratios early in the morning are due to the small values of eddy-correlation heat flux at that time of day (i.e., the possibility of a very large ratio exists even with a small absolute difference). The feature most apparent in this plot is a fairly well-defined maximum in the sodar ratio between 0900 and 1100. This does not appear in the scintillation results. The reason for these relatively large values at this time of day are as yet unknown. They may, however, be related to rapid changes occurring at higher altitudes during those hours just after the mixing layer has risen most rapidly.

A critical factor in the evaluation of surface heat flux with these methods is the Bowen ratio β , (the ratio of sensible to latent heat flux). The comparisons displayed in Figures 1 to 3 utilized values of β determined from measurements of net radiation and heat flux via eddy correlation, generally accepted to be the most accurate technique. When the remote sensors are used to estimate heat flux independently, β must be calculated using an iteration technique, with the first estimate of H established by the remote sensor. In the case of the scintillation estimates, this process works quite well. For sodar results, however, this procedure is not so successful, because over-estimates of the initial H value result in too small a correction for water vapor, while underestimates result in too large a correction. The result is that a skewed distribution in the initial observations tends to remain skewed. Thus, the average ratio for sodar with self-determined β is 1.12, slightly larger than that using β derived from the eddy correlation estimates.

In summary, the three techniques show good agreement in estimated values of surface heat flux. Large changes in heat flux are detected equally

well by each method. Estimates of H are sensitive to the value of the Bowen ratio, particularly estimates from sodar, and further work is under way in this regard.

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A NOTE ON SENSOR REQUIREMENTS FOR EDDY CORRELATION APPLICATIONS

B. B. Hicks

Recent interest in the experimental evaluation of dry deposition to natural surfaces has resulted in a number of careful examinations of the instrumental requirements and sampling constraints imposed by the micro-meteorology of the problem. Hicks and Wesely (1978) have shown that in order to measure deposition velocities to an accuracy of 0.1 cm s^{-1} by application of the familiar gradient method employing twofold height intervals, it is necessary to attain better than 0.5% accuracy in chemical concentration measurement in conditions of average winds (i.e., 30 cm s^{-1} friction velocity, see Hicks and Wesely Eq. 5). If any single determination of the concentration gradient is accurate to 20%, for example, then it is obvious that about 1600 independent evaluations of the gradient would be required in order to evaluate the deposition velocity to 0.1 cm s^{-1} accuracy. In this regard, the benefit of measuring vertical profiles by means of a single sampler, either by moving the analyzer itself or its sampling port, becomes obvious, since in this manner differences between instruments can be eliminated and effects of sensor drift can be minimized.

The eddy correlation method of turbulent flux measurement imposes a different set of instrumental requirements, most importantly regarding the response time of the sensors. Figure 1 is a schematic representation of spectra of vertical velocity fluctuations and of variations in concentration of some chemical containment at height z above a horizontally homogeneous, flat surface. Following conventional meteorological practice, frequencies are non-dimensionalized according to the wind speed and height. Also shown are typical cospectra, illustrating that very high frequency fluctuations do not contribute to the net flux of material, nor do low frequency variations. In most cases, the vertical flux is associated with frequencies of about $f = nz/u = 0.01$ to 1.0 . Thus, it is required that the sensors respond adequately to fluctuations over this range, indicating that the response time should be a

FIG. 1.--Schematic representation of the power spectra and cospectra of vertical wind speed w and some contaminant concentration c (which is assumed to behave like temperature) plotted as functions of the normalized frequency $f=nz/u$. Three stability situations are illustrated in each case. The upper curves represent the strongly unstable case, and the lower curves represent stable stratification. The middle curves are meant to show near-neutral behavior.

fraction of a second in most experimental circumstances. However, it is not required that the sensor produce a "clean" signal at high frequencies, since only signals that are correlated with vertical velocity variations will result in a net transport of material. In concept, this fundamental feature of the eddy covariance method suggests that it is permissible to relax the instrumental requirements for detection of the quantity in question, provided an accurate and uncontaminated measure of the vertical wind component can be obtained. However, noise generated by the sensors will influence the quality of the fluxes derived. It is the present intent to address this question statistically, in order to develop guidelines for instrumental signal-to-noise ratios for covariance applications.

Consider the situation in which the eddy covariance method is used to determine the flux F_s of some species s , whose concentration is measured by means of an imperfect sensor. The flux is expressed, in conventional terms,

as

$$\overline{F_s} = \overline{w'C_s'} , \quad (1)$$

where w is the vertical wind velocity and C_s is the appropriate concentration, expressed as the mass of material per unit volume of dry air. (If another measure of the concentration is used, corrections might be needed to correct for the effects of mean density fluxes, as discussed by Jones and Smith, 1977.) Following convention, primes denote departures from mean values, and the overbar indicates a time average. Thus, the covariance is constructed as the average of a large number of separate evaluations of the product $w'C_s'$, each of which is subject to error resulting from inaccuracies in the measurement of C_s' .

Suppose that the noise level of the signal derived from the concentration detector is characterized by a standard deviation σ_s , corresponding to a noise/signal ratio A . It is assumed that the noise is random. The consequences of combining "instantaneous" estimates into groups made up of n individual evaluations will be to reduce the noise-signal ratio to $A/(n)^{1/2}$. If the vertical velocity is measured with precision, then it is clear that this kind of analysis permits estimation of the possible errors in eddy fluxes evaluated from an impure concentration signal.

As an example, if individual evaluations of the product $w'C_s'$ are derived from measurements of C_s' made every half second with a noise/signal ratio of 10 at a frequency of 2 Hz, then over a one hour sampling period a total of 7200 determinations of the covariance will be combined to give an estimated error margin for the flux of about 12%.

In practice, errors always arise in the determination of the vertical velocity as well. Provided care is taken to isolate the wind and concentration sensors both physically and electronically, spurious signals arising from each should be independent of noise in the other. In this case, the net effect of noise in each of the sensors involved can be evaluated on the assumption of statistical independence, in which event variances can be taken to be additive. Thus, we can extend the example used above by now assuming that

the vertical velocity sensor is imperfect and also gives rise to a noise/signal ratio of 10, for example, for the half-second evaluations considered above. The additional error introduced as a result of a poor velocity signal is then precisely the same as that attributable to the concentration sensor performance, and the overall error margin anticipated for a flux calculated over an hour would be increased by a factor of $\sqrt{2}$ to about 17%.

It should be emphasized that the statistical methods considered above address the question of errors likely to arise in the evaluation of eddy fluxes; they do not provide a means for evaluation of the error associated with any single determination of the flux. To illustrate this point, the analysis procedure has been tested by use of a simulation involving pure sine waves, contaminated by the injection of random noise at variable levels. There is no loss of generality resulting from the use of pure sine waves in this simulation, since from first principles it is clear that the eddy flux procedure can be considered in terms of a Fourier series, in which case the present analysis can be taken to apply to each separate oscillatory component. The period of the simulated variations in w (and C_g) was taken to be 20 s, selected to be roughly at the peak of the covariance spectrum illustrated in Figure 1. Source signals were assumed to be digitized at 0.1 s intervals, and the effects of noise/signal ratios in the range 0.004 to 40 were evaluated over sampling times ranging up to 10 min. For this purpose, noise signals were injected into each channel by means of independent random number generators. Figure 2 compares the errors evaluated in these simulations with the error margins anticipated on the basis of the statistics of the problem as outlined above. It should be remembered that exact point-by-point agreement is not to be expected. Rather, the experimental points should be randomly distributed about a line of unit slope, as indeed is seen to be the case.

Having demonstrated that the matter can be addressed by use of simple and conventional statistics, it remains to apply these arguments to some practical problems of current importance. First, let us consider the case of the rapid-response particle detector used by Wesely and Hicks (1979) to evaluate eddy fluxes of small particles above grassland, crops, etc. In most

FIG. 2.--Results of a test of eddy correlation sampling theory, conducted by injecting known quantities of random noise into otherwise perfectly correlated signals. Run lengths ranged up to about 15 minutes, and data were digitized at 0.5 Hz. Perfect agreement is not expected; rather, the points should be randomly distributed about the line of equality.

practical applications, the sensor is employed in a format in which the effective response time is about 1 s, with a random noise level corresponding to about 25% of the mean level. In comparison to this RMS noise level, the particle concentration signal from which eddy fluxes must be derived is of very low amplitude. It is informative, in this regard, to write the eddy flux as

$$F_s = r_{wC} \cdot \sigma_w \sigma_C \quad (2)$$

which is in essence a definition of the relevant correlation coefficient r_{wC} . The value of r_{wC} is known to be 0.3–0.4 in most circumstances, and so we can relate the RMS concentration fluctuation σ_C to deposition velocity v_d by use of the definition

$$v_d = F_s / C_s \quad (3)$$

Furthermore, σ_w is known to be related to the friction velocity, with a constant of proportionality typically about 1.35 in near-neutral conditions. Combining these relations yields the conclusion that

$$\sigma_C / C_s \cong 2.5 (v_d / u_*) \quad (4)$$

Suppose, then, that it is intended to measure the deposition velocity of these small particles, and that it is anticipated that the result will be about

0.1 cm s^{-1} . From Eq. 4 the RMS concentration signal will be about 0.8% of the mean concentration, in average winds ($u_* = 30 \text{ cm s}^{-1}$). The noise/signal ratio is then about 30 for the concentration sensor. Working on the assumption that a sampling rate corresponding to two sensor time constants provides statistical independence of successive measurements, it should be expected that a two-hour sampling period will provide no better than 50% accuracy of the measured flux. To achieve 10% accuracy (i.e., an experimental resolution corresponding to 0.01 cm s^{-1} in v_d), data for about 30 hr would be required.

It is informative to note that the experimental evaluations of v_d over Lake Michigan, tabulated by Williams et al. (1979) indicate a standard deviation of 0.04 cm s^{-1} associated with a two-hour period of observation, using the instrument considered here. The calculations above indicate that a value of about 0.05 cm s^{-1} should have been obtained, which seems an acceptable (and encouraging) verification of the methods employed.

The case of measurement of the oceanic flux of carbon dioxide presents a second example of current interest. The problem that confronts the CO_2 research community is to measure fluxes that average about $2 \times 10^{-11} \text{ g cm}^{-2} \text{ s}^{-1}$, by use of equipment that typically has a high-frequency noise level of about $1.5 \times 10^{-10} \text{ g cm}^{-3}$ (see Jones and Smith, 1977; Figure 1) at 1 Hz. In average oceanic winds (u_* about 30 cm s^{-1}), RMS CO_2 concentration fluctuations are likely to be of the order of $1.5 \times 10^{-12} \text{ g cm}^{-3}$, indicating an experimental noise/signal ratio of about 100. On this basis, the standard deviation of daily averages of the apparent eddy fluxes would be anticipated to be about 50%.

A new CO_2 concentration sensor under development at Lawrence Livermore Laboratory (q.v. Bingham et al., 1978) promises considerably improved performance, with a high-frequency noise level below $10^{-10} \text{ g cm}^{-3}$ at 10 Hz. At an operating frequency of 1 Hz, noise/signal ratios should be about 20, resulting in at least a factor of two reduction in the random scatter of observed eddy fluxes.

In practice, fluxes measured over open ocean are rarely constant; in the case of sensible heat, for example, variations of at least an order of

magnitude are likely to be found during the course of a single day. In much the same way, fluxes of CO_2 are likely to be highly variable, and it should be expected to find that the long-term average flux results from averaging a large range of observations, some of which perhaps even change sign.

In conclusion, estimates of the errors arising in eddy flux measurement as a consequence of sensor inadequacies can be derived by applying relatively simple statistics to the problem. In the example of particle transfer, experimental results appear to confirm much of the anticipated behavior. In the more difficult case of CO_2 flux measurement at sea, the statistical considerations indicate that long averaging times will be required in order to reduce the effects of experimental error to acceptable levels.

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ESTIMATED DRY DEPOSITION VELOCITIES OF SULFUR OVER THE EASTERN UNITED STATES AND SURROUNDING REGIONS

C. M. Sheih, M. L. Wesely, and B. B. Hicks

Introduction

The Multistate Atmospheric Power Production Pollution Study (MAP3S) aims to develop an improved capability of modeling the atmospheric sulfur pollution resulting over the northeastern United States from power production. A number of modeling efforts, ranging in complexity from simple trajectory models to sophisticated dynamic grid models, are underway with the MAP3S scientific community. One of the processes rather difficult to describe in many models is dry deposition. A common approach is to estimate vertical fluxes (e.g., Bolin and Persson, 1975; Sheih, 1977) as the product of constant deposition velocities and concentrations of sulfur dioxide and particulate sulfur near the ground. Deposition velocities chosen are typically 1 cm s^{-1} for gaseous sulfur and 0.1 cm s^{-1} for particulate sulfur. Further, these values are often assumed to be appropriate for rather large regions with highly varied surfaces and for long periods of time during which large temporal variations of atmospheric stability actually occur. Depending on the surface characteristics, atmospheric conditions, and nature of the pollutant, values of deposition velocities can vary a great deal (q.v., Hicks, 1976; Sehmel and Hodgson, 1976). Such variations may significantly affect the sulfur budget over regional scales ($\sim 1000 \text{ km}$).

In order to improve the treatment of these deposition velocities, Shieh et al. (1979) have assembled the components necessary to estimate surface deposition velocities likely to be appropriate for four seasonal conditions over the eastern half of the United States and surrounding regions. The present report is a brief summary of this work.

Components for Estimating Deposition Velocities

The major equations used in the calculations of deposition velocities are given by Wesely and Hicks (1977). The associated surface resistances

are inferred from various experimental studies, some of which are listed in Table 1. The limited quantity of experimental data emphasizes the need for more field studies of dry deposition in natural circumstances and underlines the caution with which the following extrapolations and preliminary conclusions should be applied. Table 2 shows the estimated surface roughnesses and resistances, as a function of Pasquill stability categories A to F, for various land-use types for midsummer; three other seasonal conditions have been

Table 1. Summary of results of selected field experiments related to the dry deposition of sulfur compounds. Unless otherwise specified, the behavior of the vegetated surfaces is for summertime, dry, midday conditions.

Source	Pollutant	Surface	Probable r_s , s cm ⁻¹	Method of measurement
Garland et al. (1973)	SO ₂	Medium grass	0.4 ^a	Gradients
Garland et al. (1973)	SO ₂	Medium grass	0.5 ^a	Tracer ³⁵ S
Garland et al. (1974)	SO ₂	Short grass	0.2 ^a	Gradients
Owers and Powell (1974)	SO ₂	Grass	0.8 ^a	Tracer ³⁵ S
Shepherd (1974)	SO ₂	Grass	0.8	Gradients
Whelpdale and Shaw (1974)	SO ₂	Various	<< 10	Gradients
Dannevik et al. (1976)	SO ₂	Grass, winter	2.0	Gradients
Dannevik et al. (1976)	SO ₂	Soybeans	0.7	Gradients
Dovland and Eliassen (1976)	SO ₂	Snow	<< 10	Accumulation of S
Garland (1977)	SO ₂	Basic soil	0.0	Gradients
Garland (1977)	SO ₂	Pine trees	≥ 0.5	Gradients
Fowler (1978)	SO ₂	Grass	0.7 ^b	Gradients
Fowler (1978)	SO ₂	Grass, night	2.0 ^b	Gradients
Fowler (1978)	SO ₂	Grass, winter	1.5 ^b	Gradients
Fowler (1978)	SO ₂	Soil	0.0 ^b	Gradients
Fowler (1978)	SO ₂	Dew-covered	0.0 ^b	Gradients
Wesely et al. (1977)	Particles	Grass, winter	0.6 ± 0.1	Eddy-correlation
Hicks and Wesely (1978)	SO ₄ ⁼	Pine trees	1.4 ± 0.2	Eddy-correlation
Everett et al. (1979)	SO ₄ ⁼	Grass	<< 10	Gradients
Wesely and Hicks (1979)	Particles	Senescent maize	> 1.0	Eddy-correlation
Wesely and Hicks (1979)	Particles	Snow	>> 1.0	Eddy-correlation
Wesely and Hicks (1979)	Particles	Wet soil, winter	>> 1.0	Eddy-correlation

^aAs evaluated by Wesely and Hicks (1977).

^bSuggested by Fowler on the basis of direct field measurements.

Table 2. Values of surface roughnesses and surface resistances to SO₂ uptake for cases when vegetation is lush, as during midsummer.

Land-use type	z_o , cm	r_s , cm ⁻¹ s			
		For stability categories			
		A,B,C	D	E	F
1. Cropland and pasture	20	1.0	3.0	10.0	0.0
2. Cropland, woodland, and grazing land	30	1.0	3.0	10.0	0.0
3. Irrigated crops	5	1.0	3.0	10.0	0.0
4. Forest and woodland grazed	90	1.0	3.0	10.0	0.0
5. Forest and woodland ungrazed	100	1.0	3.0	10.0	0.0
6. Subhumid grassland and semiarid grazing land	10	1.0	3.0	10.0	0.0
7. Open woodland grazed	20	1.0	3.0	10.0	0.0
8. Desert shrubland	30	2.0	5.0	10.0	10.0
9. Swamp	20	0.5	0.75	1.0	0.0
A. Marshland	50	0.75	3.0	10.0	0.0
B. Metropolitan city	100	10.0	10.0	10.0	0.0
C. Lake or ocean	~ 0.01	0.0	0.0	0.0	0.0

investigated also. The midsummer case corresponds to surfaces when vegetation is lush and healthy, typical of midsummer in the Northeast and Midwest, but also other seasons in locations where frost is less common. Of the three seasons not considered fully in this report, the spring case corresponds to surfaces when green leaves are emerging on trees and field crops are small, not entirely covering the soil surface. This is a transitional situation meant primarily to apply in late spring and early summer within 1 to 2 months after the last killing frost. The autumn case represents conditions when frost is common, deciduous trees are leafless, field crops have been harvested to expose much bare soil, grass surfaces are brown, and no snow is present. Obviously this corresponds to autumn, early spring, and winter when no snow is present. The fourth condition considers snow-covered

surfaces and subfreezing temperatures. Since no single seasonal condition can apply in any given instance to the entire region depicted by Figure 1, care should be taken in choosing which condition or combination of conditions apply. A surface resistance of 1.0 s cm^{-1} was assumed for particles in all cases. The land-use map (Figure 1) was provided by T. W. Yu and W. A. Shaffer of the Techniques Development Laboratory, NWS, Silver Spring, Maryland (Long et al. 1978).

To address simulations for which meteorological data for computing atmospheric stabilities are limited, maps of surface deposition velocities have been computed and categorized in terms of Pasquill classifications. This allows selection of the most appropriate deposition velocity distributions on the basis of the rather crude technique that is conventionally employed to estimate dispersion coefficients. Sample results of deposition velocity maps for sulfur dioxide and sulfate particles corresponding to the Pasquill stability category B are shown in Figure 2. Only a quick glance at the various maps is needed to see that the conventional approach of assuming a constant deposition

FIG. 1.--Land-use map represented by alphanumeric types, as listed in Table 2. (ANL Neg. 149-79-81)

FIG. 2.--Sample results of deposition velocity maps for (a) sulfur dioxide and (b) sulfate particles. The symbols (0,1,2,---, and 9) in the maps represent deposition velocities with intervals of 0.1 cm s^{-1} ; i.e., 0 and 9 represent the ranges of deposition velocities 0 to 0.1 and 0.9 to 1.0 cm s^{-1} , respectively.

velocity in numerical models is not appropriate to regional-scale applications. Of course, sensitivity tests are needed to determine whether varying the velocities in such a range will result in important changes in the desired output, such as mean sulfate concentrations. At the least, the present maps help provide the opportunity to perform such sensitivity tests.

Conclusion

The information needed for purposes of computing sulfur dioxide and sulfate deposition velocities have been described, and results have been documented for the eastern half of the United States, southern Ontario, and adjacent oceans. The deposition velocities derived are appropriate to a height of 10 m, the level at which Pasquill stability categories are usually defined. For application to numerical models that deal with smaller heights, the assumption of a constant flux layer can be used to determine the relevant levels within the surface boundary layer.

The assembled information is readily adaptable for use in diffusion models describing sulfur pollution over an area within the region addressed. If meteorological data are sufficiently detailed to allow computations of atmospheric stability parameters such as the Obukhov scale length, workers are encouraged to compute their own deposition velocities, using the equations given by Sheikh et al. (1979), together with estimates of surface roughness (z_0) and surface resistance (r_s) such as those given here. On the other hand, for a Gaussian plume model that calls for only crude estimates of atmospheric conditions, the present work provides a set of deposition velocity maps classified according to Pasquill stability categories A to F (e.g., Turner, 1970).

The data documented in the MAP3S data bank are the land-use map of the eastern United States and surrounding areas, a table of surface roughnesses and resistances for various land-use types and for four seasonal conditions, and maps of deposition velocities of sulfur dioxide and sulfate particles for the various combinations of seasonal conditions and stability. Access to these data can be obtained by contacting the Meteorology Department, Brookhaven National Laboratory, Upton, Long Island, New York 11973.

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ON THE DEPOSITION OF SUBMICRON PARTICLES TO OPEN WATER SURFACES

B. B. Hicks and M. L. Wesely

The matter of particle deposition to open water surfaces continues to present intriguing physical and conceptual questions. Wind tunnel modeling studies have been convincing in their demonstration that exceedingly small deposition velocities are to be expected (q.v., Sehmel and Sutter, 1974), but it has also been pointed out that the average state of the ocean cannot be simulated in the confined experimental area of even the largest wind tunnel facilities. Most theoretical arguments that have been produced to this time support the wind tunnel observations. These theoretical models have been discussed in some detail by Slinn et al. (1978); most are based upon some combination of descriptions of (a) the inability of particles to respond to the high-frequency turbulent motions that transfer momentum immediately above the surface, (b) the exceedingly low rate at which they can penetrate any diffusive layer in contact with the surface, should such a layer exist, and (c) phoretic effects, in this case due mainly to the flux of water molecules away from the surface.

There seems to have been little attention paid to the role of surface tension, which typically takes values in the range 70 to 80 dynes cm^{-1} in unpolluted areas. Rule-of-thumb calculations for a one-micron unwettable particle with unity specific gravity indicate that such a falling particle will be incapable of breaking through the surface tension barrier unless the velocity of impaction is well above 10 m s^{-1} . Otherwise, particles will bounce and will eventually form a layer of particles sitting on a virtually impenetrable surface. (One remembers, in this context, the familiar sight of insects walking across the surface of calm water.) However, if the particles of interest are wettable, the possibility of passage through the surface is very high.

Another likely route for the passage of particles through the water surface is associated with the breaking of waves. In this sense, a simple physical model can be constructed as follows.

Starting with a resistive model of the transfer process, we can specify an aerodynamic term $r_a = u/u_*^2$ (assuming neutral conditions), and a surface resistance r_s which is very large when the surface is rather smooth (Williams et al., 1978) and takes some relatively small value r_b in the areas of breaking waves. These terms can be combined with the resistance in parallel, by considering the proportion of the surface which is covered by breaking waves. In oceanographic studies, this property has received considerable attention, and so we can use it in the present application with some confidence. Kraus (1972), for example, presents data attributed to Blanchard (1963), from which the values listed in Table 1 have been extracted. These evaluations of the fraction of white water emphasize the limited applicability of most of the air-water pollutant exchange studies performed to date; no breaking waves are expected to occur when the wind is less than about 4 m s^{-1} . The global average wind speed is about 8 m s^{-1} , and in the "roaring forties" about 20% of the surface is white water, even as an annual average.

Table 1 also lists estimates of the drag coefficient ($C_d = u_*^2/u^2$) from which r_a for near-neutral conditions can readily be evaluated. Although discussion about the magnitude of the drag coefficient and its wind speed dependency continues, the values tabulated here are certainly close enough for the present purposes. The last three columns in Table 1 give the final product of these considerations: estimates of the spatially averaged deposition velocity appropriate for the parameterization of particle fluxes to open water, evaluated on the assumption that the surface resistance in areas of white water takes a value of either 0, 1, or 2 s cm^{-1} . The tabulated values are obtained on the assumption that observations are made at the standard height of 10 m, but this is not critical.

The major feature of Table 1 is the importance of the property r_b . If it is assumed that r_b is near zero, then relatively high deposition velocities should be expected in strong winds. But fluxes will be greatly affected even if r_b takes the seemingly low value of 1 s cm^{-1} . In the highly turbulent mixture of spray and water immediately above breaking waves, very low values of r_b can easily be rationalized, but as yet experimental support is lacking.

Table 1. Tabulation of values of the wind speed at 10 m height (u), the fraction of the surface covered by breaking waves (A), and the drag coefficient (C_d), which are used to evaluate the aerodynamic resistance (r_a) and the net effective deposition velocity (v_d) of small particles to open water surfaces, for a range of values of the white-water surface resistance (r_b , in $s\ cm^{-1}$).

u , m/s	A , %	C_d	r_a , s/cm	v_d , cm/s		
				$r_b = 0$	$r_b = 1$	$r_b = 2$
1	0	0.0010	10.0	~ 0	~ 0	~ 0
2	0	0.0010	5.0	~ 0	~ 0	~ 0
3	0	0.0010	3.3	~ 0	~ 0	~ 0
4	0	0.0010	2.5	~ 0	~ 0	~ 0
5	1.0	0.0010	2.0	0.005	0.003	0.003
10	4.5	0.0017	0.59	0.08	0.03	0.02
15	10.5	0.0020	0.33	0.32	0.08	0.05
20	18.7	0.0024	0.21	0.90	0.15	0.09
25	27.0 ^a	0.0028	0.15	1.9	0.24	0.13
30	36.0 ^a	0.0031	0.11	3.3	0.33	0.17

^aEstimated value.

In fact, the basic tenets of the model put forward here can only be tested by direct measurements of particle fluxes to open water surfaces, in wind speeds above $5\ m\ s^{-1}$. The open-water evaluations of deposition velocities presented in a companion report (Williams et al., 1978) were obtained when no whitecaps were present, but have succeeded in demonstrating the feasibility of obtaining relevant experimental evidence. It should be anticipated, however, that the present model might be a considerable oversimplification. For example, the simple resistive electrical analogy fails to take into account the flux of particles to the air layer immediately above a smooth water surface accompanied by the possible periodic entrapment of particles when waves break. This process can be visualized as a capacitor which is being charged at some steady, slow rate, and is being periodically short-circuited. At this time, it seems best to await the results of experiments presently being

planned before proceeding with further conjecture along these lines.

There is nothing in any of the basic considerations presented here which would suggest a size dependency of the deposition velocity, save for that imposed by gravity. Further, the matter of surface contamination is not addressed here, even though some effects must be anticipated. Perhaps, however, surface-film effects will occur mostly with light winds, when particle fluxes will be small anyway.

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ATMOSPHERIC DIFFUSION PARAMETERS OVER WATER NEAR SHORE

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Introduction

Methods for computing dispersion coefficients should be consistent with Taylor's theory (1921) or, equivalently, with Kampe de Fariet's equation (1939),

$$\sigma_{\xi}^2(T) = 2\sigma_{\eta}^2 T \int_0^T (1 - \tau/T) R_{\eta}(\tau) d\tau, \quad (1)$$

even though homogeneous, steady-state conditions required by the theory are, at best, only partially realized in nature. Here ξ and its corresponding η are the general spatial and velocity coordinates, respectively, T is travel time, σ_{ξ} and σ_{η} are dispersion coefficients, and R is the Lagrangian autocorrelation function. With the Lagrangian integral time scale ($\mathcal{T}_{\eta}$) defined as

$$\mathcal{T}_{\eta} = \int_0^{\infty} R_{\eta}(\tau) d\tau, \quad (2)$$

Eq. 1 can be integrated for $T \ll \mathcal{T}_{\eta}$ and $T \gg \mathcal{T}_{\eta}$ because the corresponding values of the autocorrelation are 1 and 0 (e.g., Taylor, 1921). However, determination of the autocorrelation function for intermediate travel times presents difficulty in routine use of these equations.

In order to provide a simple method for estimating the dispersion coefficient over intermediate travel times, Pasquill (1975) expressed the autocorrelation function in terms of the Lagrangian spectral density function and showed that for a given spectral density function, Eq. 1 can be restated as

$$\sigma_{\xi}^2 / (\sigma_{\eta}^2 T) = F_{\eta}(T / \mathcal{T}_{\eta}), \quad (3)$$

where $F_{\eta}(T / \mathcal{T}_{\eta})$ will be called the PT (Pasquill-Taylor) dispersion function. To be consistent with the behavior of the autocorrelation function, the PT function must have the general properties $F=1$ and $F=(2\mathcal{T}_{\eta}/T)^{1/2}$ for $T \ll \mathcal{T}_{\eta}$ and $T \gg \mathcal{T}_{\eta}$, respectively. However, the PT function depends upon the shape of the wind velocity spectrum for intermediate travel times.

Draxler (1976) computed the PT dispersion function for dispersion over land by analyzing various data reported by other investigators. However, since the integral time scale was unavailable in most cases, he normalized travel time by a time constant, indirectly related to the integral time scale, derived from a plot of $\sigma_{\xi}/(\sigma_{\eta} T)$ vs. T . Most recently, Irwin (1979) also recommended the use of such a function for estimating dispersion coefficients.

In the present study, the PT dispersion function is evaluated for the experiments over water near shore described in detail earlier (Hart and Frenzen, 1976). In particular, the Lagrangian integral time scale is computed directly from neutrally-buoyant balloon trajectories. Fourier analysis has been used to compute the dispersion parameters.

A Lagrangian time series of balloon coordinates $\xi(t)$ can be represented by a discrete Fourier series as follows:

$$\xi(p \cdot \Delta t) = \sum_{q=0}^{N-1} A_{\xi}(q \cdot \Delta f) \exp(-i2\pi pq/N), \text{ for } p=0, \dots, N-1 \quad (4)$$

with

$$\Delta f = (N \cdot \Delta t)^{-1},$$

$$\Delta t = T_s/(N-1),$$

$$t = q \cdot \Delta f,$$

and

$$t = p \cdot \Delta t,$$

where N is the total number of samples taken during the total sampling duration T_s with discrete sampling interval Δt , f is frequency, t is time, and $i = (-1)^{1/2}$. The Fourier coefficient $A_{\xi}(f)$ can be found from the inverse transformation of Eq. 4:

$$A_{\xi}(q \cdot \Delta f) = \sum_{p=0}^{N-1} \xi(p \cdot \Delta t) \exp(i2\pi pq/N), \text{ for } q=0, \dots, N-1. \quad (5)$$

The corresponding balloon velocities can then be obtained simply by taking a time derivative of Eq. 4, thus

$$n(p \cdot \Delta t) = N^{-1} \sum_{q=0}^{N-1} A_{\xi}(q \cdot \Delta f) (-i2\pi q/N) \Delta t^{-1} \exp(-2\pi p q/N) ,$$

for $p = 0, \dots, N-1$. (6)

The Lagrangian autocorrelations for these velocity components are

$$R_{\eta}(\gamma \cdot \Delta t) = 4\pi^2 N^{-4} \Delta t^{-2} \sigma_{\eta}^{-2} \sum_{q=0}^{N-1} q^2 A_{\xi}(q \cdot \Delta f) A_{\xi}^*(q \cdot \Delta f) \\ \times \exp(-i2\pi \gamma q/N) , \text{ for } \gamma = 0, \dots, N-1 ,$$

(7)

where the velocity variances are

$$\sigma_{\eta}^2 = 4\pi^2 N^{-4} \Delta t^{-2} \sum_{q=0}^{N-1} q^2 A_{\xi}(q \cdot \Delta f) A_{\xi}^*(q \cdot \Delta f) ,$$

(8)

and the asterisk indicates the complex conjugate. The dispersion coefficient is computed from

$$\sigma_{\xi}^2(p \cdot \Delta t) = 8\pi^2 N^{-4} \sum_{\gamma=0}^{p-1} \sum_{k=0}^{\gamma-1} \sum_{q=0}^{N-1} q^2 A_{\xi}(q \cdot \Delta f) A_{\xi}^*(q \cdot \Delta f) \\ \times \exp(-i2\pi q_k/N) , \text{ for } p = 0, \dots, N-1 .$$

(9)

Finally, the measured Lagrangian integral time scale is computed from

$$\sigma'_{\eta} = \Delta t \sum_{\gamma=0}^{N-1} R_{\eta}(\gamma \cdot \Delta t) ,$$

(10)

where the superscript prime indicates an observed value.

Discussion

Table 1 extends the analysis presented earlier (Sheih et al. 1979) of field data collected over Lake Michigan. The Obukhov stability parameter (z/L , with L being Obukhov length) is computed using the bulk aerodynamic procedure described by Hicks (1975) and mean meteorological measurements taken at the boat. To classify the atmospheric stabilities for other applications where only limited meteorological observations are available, the

Table 1. Dispersion parameters and Lagrangian statistics of balloon observations over water near shore.

Pasquill category	z/L	N	$\bar{u}$ $m\ s^{-1}$	u_* $m\ s^{-1}$	T_1 s	T_2 s	σ_v $m\ s^{-1}$	σ'_v s	σ_{y1} m	σ_{y2} m	α_v	σ_w $m\ s^{-1}$	σ'_w s	σ_{z1} m	σ_{z2} m	α_w
C	$\begin{Bmatrix} -2.8 \\ -2.4 \\ -1.5 \\ -1.3 \end{Bmatrix}$	6	4.0	0.14	25	50	0.52	9.0	12.1	19.5	0.11	0.31	2.3	4.2	5.7	0.33
D	$\begin{Bmatrix} -0.2 \\ 1.5 \end{Bmatrix}$	27	6.3	0.19	15.9	31.7	0.38	7.3	6.0	11.1	0.03	0.27	5.3	2.9	5.0	0.29
E	$\begin{Bmatrix} 1.7 \\ 2.1 \\ 4.0 \end{Bmatrix}$	7	5.7	0.16	17.5	35.1	0.38	8.1	5.1	8.7	0.24	0.19	6.6	2.1	3.5	0.39

Obukhov stability parameter is related to Pasquill stability categories by use of the nomogram provided by Golder (1972). Further description of the above computations is given by Sheih et al. (1979). M is the number of trajectories considered in each stability category, $\bar{u}$ is the mean wind speed, u_* is the friction velocity, and T_1 and T_2 are travel times at which dispersion coefficients are computed. For the PT function, Draxler (1976) proposed

$$F_{\eta} (T/\mathcal{T}'_{\eta}) = [1 + \alpha_{\eta} (T/\mathcal{T}'_{\eta})^{0.5}]^{-1} , \quad (11)$$

where α_{η} is a constant. One can determine α_{η} as a function of $T/\mathcal{T}'_{\eta}$ by using regression analysis. The analysis tabulated in Table 1 indicates that α_v for lateral dispersion is smaller than α_w for vertical dispersion, and the minimum values for both occur at near-neutral stability.

It can be shown that, as T approaches infinity, Taylor's theory gives

$$F_{\eta} = (2\mathcal{T}'_{\eta}/T)^{0.5} , \quad (12)$$

while Eq. 11 becomes

$$F_{\eta} = (\alpha_{\eta}^{-2} \mathcal{T}'_{\eta}/T)^{0.5} . \quad (13)$$

Combining Eqs. 12 and 13, one obtains

$$\mathcal{T}'_{\eta} = 0.5 \alpha_{\eta}^{-2} \mathcal{T}'_{\eta} . \quad (14)$$

Equation 14 indicates that the observed integral time scale should be multiplied by $0.5 \alpha_{\eta}^{-2}$ to make it equivalent to the "effective" integral time scale implied in Taylor's theory. The effective integral time scales computed from α_{η} and $\mathcal{T}'_{\eta}$ (Table 1) indicate that lateral dispersion is dominated by scales much larger than those of vertical dispersion. The ratios of lateral to vertical effective integral time scales are 35, 129, and 3 for the stability categories C, D, and E, respectively, and are much larger than the corresponding ratios of observed integral time scales. The large variation in the value of α_v with respect to atmospheric stability is probably due to lake breeze and land breeze circulations. Multiplying mean wind speed by twice the effective integral time scale [according to Tennekes and Lumley (1972), the period of energy-

containing eddies in a turbulent flow is about twice the integral time scale], we obtain lateral length scales 3.0, 51, and 0.8 km for stability categories C, D, and E, respectively. It should be noted that the Lagrangian integral time scales used in the above calculation should be replaced by the corresponding Eulerian time scales. However, since large variations in the ratio of Lagrangian to Eulerian time scales have been reported (e.g., by Lumley and Panofsky, 1964), no attempt has been made to derive the Eulerian integral time scales from the Lagrangian measurements. For the present rather qualitative analysis, no discrimination between Lagrangian and Eulerian time scales has been made. Typical horizontal length scales of the most energetic cells of lake-breeze circulations are found to be less than 4 km (e.g., Lyons and Olsson, 1972; Sheih and Moroz, 1975). The present estimates of lateral length scales for stable and unstable conditions are consistent with these results. The lateral length scale for the near-neutral condition appears to be unexpectedly large.

Draxler (1976) excluded cases of vertical diffusion during stable conditions and represented with Eq. 11 the normalized dispersion coefficients of lateral and vertical dispersion from elevated sources over land for various field experiments reported in the literature. His value of α_η , 0.7, is more than twice the value found here for over water. The PT dispersion function of Draxler and the theoretical dispersion function given by Pasquill (1975) for an assumed exponential autocorrelation function $R_\eta(t) = \exp(-t/\sigma'_\eta)$ are compared (Figure 1) with the results (solid lines) of the Lake Michigan field experiments. Vertical dispersion results obtained over Lake Michigan are represented by a single curve (labelled III), since there is little apparent variation with stability. The PT dispersion function appears to take much larger values over water than over land. The figure also shows that vertical dispersion over water behaves similarly to the exponential autocorrelation function in the present range of travel times. Lateral dispersion over water during stability category D has only about 9% variation for travel time $T=0$ to $12\sigma'_\eta$ and for practical applications perhaps can be considered as unity over this range.

FIG. 1.--Comparison of the Pasquill-Taylor dispersion functions for (I) an exponential correlation function, (II) Draxler's result over land, (III) vertical dispersion over water, and (IV) lateral dispersion over water as a function of atmospheric stabilities C, D, and E. The solid line indicates the results of the Lake Michigan field experiments.

Since balloons were tracked to only about 200 s downstream from release, the results should be extended to greater downstream distances with caution, primarily because some large-scale motions, previously filtered by the finite sampling duration, might then enter into the statistics. However, vertical dispersion should not be greatly affected because apparently little energy is contained in scales larger than 200 s.

Conclusion

The concept of multiplying the Pasquill-Taylor function by the travel time and the standard deviation of wind speed in order to compute a dispersion coefficient (as suggested by Pasquill and applied to dispersion over land by Draxler) has been used to analyze Lagrangian measurements taken over water near shore. The results for both lateral and vertical dispersion show that values of the PT function over water are generally higher than over land. The characteristic length scales estimated for lateral dispersion over the water are 3.0, 51.1, and 0.8 km for unstable, near-neutral, and stable atmospheres, respectively. The PT dispersion function of vertical dispersion over water is found to be similar to that for an exponential autocorrelation function.

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MEASUREMENTS OF ATMOSPHERIC DISPERSION OVER WATER NEAR SHORELINE ENVIRONMENTS

C. M. Sheih, P. Frenzen, and R. L. Hart

Introduction

Improved methods of estimating atmospheric dispersion over water in onshore flow are needed because of expected offshore construction of power plants. Relatively little is known of the unique dispersive properties of the coastal marine atmosphere, not only because flow over water has been studied far less than flow over land, but also because small-scale dispersion over water is often combined with large-scale meandering due to sea-breeze or lake-breeze circulations. These two scales of motion must be identified and separated in order to analyze observations logically.

A comprehensive survey of earlier work on dispersion experiments over oceans and shorelines (Prophet, 1961) primarily discusses the changes of dispersive characteristics of flow passing from land to water. In a somewhat different manner, Van der Hoven (1967) has summarized dispersion during on-shore air flow. He categorizes flow regimes as overwater, overland, and transitional from overwater to overland. Relatively few diffusion studies, however, have been conducted over water, perhaps because of the difficulties encountered. Slade (1962) made a series of measurements of air-water temperatures, mean wind speed and wind-direction fluctuations on the east and west shores of Chesapeake Bay during general westerly wind conditions to obtain estimates of the relative importance of diffusion over land and over water. More recently, the results of an extensive series of studies of smoke plume diffusion over water have been reported by Raynor et al. (1975, 1978).

In the present study additional physical insight into the problem has been attained by measuring turbulence statistics of balloon trajectories in a Lagrangian frame. Field experiments, described by Hart and Frenzen (1976), were conducted over open water. Fourier analysis has been used to separate large- and small-scale components of the trajectories, to filter high frequency noise from the data, and to compute dispersion coefficients in terms of

atmospheric stabilities.

Analysis of the Data

Frequencies that separate large-scale motion, small-scale motion, and instrumentation noise are identified by computing a velocity spectrum for each balloon trajectory. A section of the Lagrangian spectrum in a logarithmic plot (Figure 1) appears to have a slope of -2 as suggested by Corrsin (1963). This power-law behavior of the spectrum can thus be used as a guide to screen bad data. The corresponding energy-conserving plot of the spectrum (Figure 1b) also reveals contamination by high frequency noise above 0.3 Hz. This part of the spectrum would therefore be discarded from further consideration. Further, Figure 1 shows that a spectral gap near 0.03 to 0.05 Hz separates large-scale and small-scale components of the Lagrangian velocity field; this suggests that the wind fluctuations are caused by two different mechanisms and that their spectral components can be treated independently.

FIG. 1.--A sample spectrum of normal Lagrangian wind velocities in (a) a log-log plot and (b) an energy-conserving, semi-logarithmic plot.
(ANL Neg. 149-78-384)

In order to classify atmospheric stabilities for these experiments with only limited meteorological observations, Obukhov length scales (L) are obtained from estimates of momentum and heat fluxes computed by bulk aerodynamic methods described by Hicks (1975). An average balloon height of $z = 25$ m is assumed throughout the analysis. Further, by assuming that the aerodynamic roughness length of the water was 0.2 mm, Pasquill stability categories can be estimated from the roughness and Obukhov length with an extrapolated version of the nomogram provided by Golder (1972). The resulting approximate relationship between Obukhov length and Pasquill category is shown in Table 1. Lagrangian correlation coefficients and variances for the various scales and velocity components are computed for each trajectory. The trajectories are classified by Pasquill category, and average correlation coefficients and variances are computed for each category. Integral time scales, defined as the total integrated area under the correlation coefficient curve, are found for each combination of scale size, velocity component, and category. Dispersion coefficients are then obtained by use of the Taylor dispersion relation.

Discussion of the Results

The average correlation coefficients for the lateral velocity component are shown in Figure 2. The total correlation coefficients for various stability categories in Figure 2 are similar. Each coefficient shows a sharp decrease in the first 2 seconds followed by a slow roll-off for larger time lags, a behavior which implies the existence of two distinct scales of lateral motion. The integral time scales are 9.0, 7.3, and 8.1 s for categories C, D, and E, respectively, indicating a somewhat higher ratio of large-scale to small-scale fluctuations for categories C and E. The correlation coefficients for the unstable case are smaller than those for other categories at lags less than 6 s, but become larger at greater lags. When the lateral velocities are separated into large-scale and small-scale components, the correlation coefficients, shown in Figure 2b, exhibit the conventional monotonic exponential form. The small-scale component correlation coefficients for all the stability categories

Stability category	z/L	N	$\bar{u}_{-1}$ $m\ s^{-1}$	$u_{*,-1}$ $m\ s^{-1}$	τ_{vs}	τ_{vl} (s)	τ_{vT}	σ'_{vs}	σ'_{vl}	σ'_{vT}	τ_{ws}	τ_{wl} (s)	τ_{wT}	σ'_{ws}	σ'_{wl}	σ'_{wT}
C	$\left\{ \begin{array}{l} -2.8 \\ -2.4 \\ -1.5 \\ -1.3 \end{array} \right\}$	6	4.0	0.14	1.3	19.5	9.0	2.9	1.8	3.7	1.3	11.5	2.3	2.1	0.5	2.2
D	$\left\{ \begin{array}{l} -0.2 \\ 1.5 \end{array} \right\}$	27	6.5	0.19	1.5	13.8	7.3	1.4	1.1	1.8	1.5	12.9	5.3	1.0	0.8	1.3
E	$\left\{ \begin{array}{l} 1.7 \\ 2.1 \\ 4.0 \end{array} \right\}$	7	5.7	0.16	1.4	17.4	8.1	1.4	1.8	2.4	1.3	14.5	6.6	0.8	0.7	1.2

considered are almost identical and give an essentially constant integral time scale of about 1.4 s. The large-scale components appear to be more stability dependent with integral time scales of 19.5, 13.8, and 17.4 s for Pasquill categories C, D, and E, respectively. The minimum value near neutral stability may indicate that neither lake nor land breeze circulations are acting in this case to increase lateral integral time scales, as indeed they might be expected to do in stable or unstable conditions due to temperature differences between the land and water surfaces.

Correlation coefficients of the vertical velocity are shown in Figure 3. The total correlation coefficients (Figure 3a) decrease faster and have smaller amplitude at large time lags than the corresponding total correlation coefficients

FIG. 2.--The Lagrangian correlation coefficients of lateral horizontal velocity components for Pasquill stability categories C ($\square$), D ($\circ$), and E ($\diamond$): (a) total signal (unshaded), and (b) large-scale component (partially shaded) and small-scale component (solid).
(ANL Neg. 149-78-383)

FIG. 3.--The Lagrangian correlation coefficients of vertical velocity components for Pasquill stability categories C ($\square$), D ($\circ$), and E ($\diamond$): (a) total signal (unshaded), and (b) large-scale component (partially shaded) and small-scale component (solid).
(ANL Neg. 149-78-381)

shown earlier for the horizontal velocity; this indicates that the ratio of large-scale to small-scale components is smaller for vertical motion than for horizontal motion. In particular, the category C value decreases dramatically to below 0.1 at only a 3 s time lag, which suggests that at elevations of about 25 m in unstable conditions over water, the vertical motion contains relatively little energy from time scales greater than 3 s. The integral time scales for the total signal are 2.3, 5.3, and 6.6 s for categories C, D, and E, respectively.

The correlation coefficients for small-scale components of vertical velocity (Figure 3b) also appear to be nearly independent of stability and give integral time scales about the same as those for horizontal motion. As the atmospheric stability changes, the correlation coefficients of the large-scale component of vertical motion vary over a somewhat smaller range than do those of lateral motion. The integral time scales of the large-scale component for vertical motion are 11.5 s and 12.9 s for categories C and D, respectively, and 14.5 s for category E.

Averaged RMS values of velocity variances for various spatial components and scales, normalized by friction velocities, are given in Table 1, along with a summary of the integral time scales. The normalized standard deviations are quite similar to (Eulerian) bivane data reported by Merry and Panofsky (1976) and Hart and Frenzen (1976), for stability categories C and D, but are much smaller for category E.

The average values of the dispersion coefficients obtained for each stability category are compared in Figure 4 to those obtained from conventional Pasquill curves for flow over gently rolling grassland. The vertical dispersion coefficient σ_z for flow over water is about a factor of 2 smaller (i.e., shifted about 2 categories toward the stable side) than is indicated for the same Pasquill stability category. For stability category C, the small-scale component (solid squares) contributes slightly more to the total dispersion coefficient (open squares). However, for categories D and E, σ_z appears to be the result primarily of large-scale motion. Lateral dispersion coefficients (σ_y) over water are also smaller (shifted about one category toward the stable

FIG. 4.--Comparison of dispersion coefficients of Pasquill (represented by curves) with the present analysis for stability categories C ($\square$), D ($\circ$), and E ($\diamond$) and components total (unshaded), large-scale (half shaded); and small-scale (solid). (ANL Neg. 149-78-382)

side) compared to the standard Pasquill curves for estimating dispersion over land. As expected, the computed results indicate that lateral dispersion is due primarily to large-scale meandering. It should be noted that although direct comparison of the present results with the plume measurements by Raynor et al. (1975 and 1978) is not possible (primarily because their results are grouped rather loosely into stable, neutral, and unstable classes), their observation of dispersion coefficients over water consistently below category E of the "standard" Pasquill curves for both vertical and lateral dispersion is quite consistent with these results.

Conclusion

It appears that there are two major scales of motion in overwater dispersion processes. While small-scale components are due to thermal and mechanical turbulence production, we speculate that, in the near-shore environment of these experiments, the large-scale component may be in part an effect of the temperature difference between land and water. The two components have been successfully separated with Fourier analysis and their individual statistics have been used in interpreting results. However, because the available data are insufficient, no further analysis of large-scale dispersion as a function of the temperature difference between land and water has been made.

In comparison to the Pasquill curves appropriate to plume dispersion over land, the offshore dispersion coefficients computed from neutral density balloon trajectories are about 50% lower (or shifted about two categories toward the stable side) for vertical dispersion, and are about 30% lower (or shifted about one category toward the stable side) for horizontal dispersion.

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ISENTROPIC TRANSPORT OF SO_2 AND O_3 DURING DACSE-I

D. L. Sisterson and J. D. Shannon

Introduction

The diurnal cycle of stability in the lowest several kilometers of the atmosphere significantly affects long-range transport and dispersion of pollutants. According to boundary-layer theory, pollutants emitted into the mixed layer during the daytime are dispersed more or less uniformly. As surface cooling begins, stable layers form and vertical diffusion of pollutants within these layers decreases. More importantly, the layers "decouple" from the surface, and all surface-layer turbulent exchange processes, including dry deposition, are either cut off or greatly reduced (Wesely and Hicks, 1977). Pollutants aloft are then transported along isentropic surfaces without appreciable dilution. If sufficient surface heating occurs on the following day, the top of the boundary layer can rise and incorporate yesterday's pollutants into the new mixed layer. If surface heating is weak, the top of the boundary layer may remain below some of the polluted layers from the previous day, with those layers left essentially intact.

An opportunity to investigate the diurnal variation of pollutant transport arose in August, 1977. In a study termed the Diagnostic Atmospheric Cross-Section Experiment (DACSE-I), air-quality profiles obtained with a research aircraft were combined with upper-air soundings made with rawinsondes. This paper summarizes the investigation of O_3 and SO_2 transport during DACSE-I.

Data Acquisition and Analysis

By special arrangement with the National Weather Service (NWS), rawinsonde stations at Salem, IL, Dayton, O, and Pittsburgh, PA, made three additional soundings daily during the period of investigation, 1-10 August 1977. These stations are located in a WSW-ENE line, which is 100 to 150 km north of the Ohio River for most of its length. Profiles of SO_2 , O_3 , b_{scat} (a measure of aerosol concentration in terms of the back-scattering of light), w (mixing ratio), and θ (potential temperature) were measured with the

Battelle Pacific Northwest Laboratory DC-3 research aircraft operated along the cross-section line. Details of the experiment, experimental procedures, and data acquisition and analysis are presented elsewhere (Sisterson and Shannon, 1977, 1979).

Data Discussion; 5 August

On 5 August, a day without significant convective activity, the maximum aircraft-measured concentrations of O_3 and SO_2 were 80 and 45 ppb, respectively. The b_{scat} cross sections, discussed elsewhere (Sisterson and Shannon, 1979), showed evidence of isentropic transport of particles in the lower troposphere. The O_3 cross sections, however, did not provide definite evidence of isentropic transport. In a fresh industrial plume, NO quickly destroys O_3 and therefore oxidant concentrations may fall rapidly below background concentrations in a thin layer (Figure 1). In a cross-section analysis based on relatively few air-quality profiles, such local effects may mask any evidence of long-range transport of O_3 . Individual air-quality profiles, however, are consistent with a regional O_3 maximum at the 305 K isentrope. Figures 2 and 3, taken one hour apart, clearly show O_3 maxima at the same 305 K isentrope.

An important feature of these profiles and others not shown is the positive correlation between O_3 and b_{scat} . This can be interpreted as evidence that the O_3 originated from anthropogenic sources and not stratospheric intrusions (Hoffman et al., 1975). Therefore, long-range transport of O_3 (barring major sinks) appears similar to the transport of particles described by Sisterson et al. (1979). Ozone aloft can also become decoupled from the surface and surface sinks at night by the formation of the nocturnal inversion. The resulting O_3 layer can then be transported along isentropic surfaces until sufficient vertical mixing is established to incorporate the O_3 into a new mixed layer.

Relatively large O_3 concentrations above the nocturnal inversion have been observed overnight by Gloria et al. (1974), and for much longer periods by Samson and Ragland (1977). In the latter study, a region of reduced

FIG. 1.--Profiles of SO₂, O₃, b_{scat}, w and θ at 0833 EST on 5 August, west of Dayton, O. Dashed line is mixing ratio.

FIG. 2.--Profiles of SO₂, O₃, b_{scat}, θ, and w at 0803 EST on 5 August, east of Dayton, O. Dashed line is mixing ratio.

FIG. 3.--Profiles of SO₂, O₃, b_{scat}, θ , and w at 0856 EST on 5 August, east of Dayton, O. Dashed line is mixing ratio.

visibility and high O₃ concentrations was associated with a high pressure system. Trajectory analysis implied a long period of pollutant loading into a relatively small volume of air as the airmass moved through the Midwest. These episodes are similar to those found by Husar et al. (1976) and Wolff et al. (1976). The explanation by Samson and Ragland of O₃ transport, particularly at night, is similar to that outlined here; i.e., the thermal decoupling and increased wind speed aloft at night (Sisterson and Frenzen, 1978) could be expected to produce an efficient isentropic transport mechanism. While the frequency of large-scale episodes of decreased visibility and high O₃ concentrations has not been determined, this study suggests that long-range transport of O₃ may not be uncommon, especially at night. This could explain observations of elevated O₃ concentrations in and above rural mixed layers hundreds of kilometers downwind of urban area sources of industrially-produced precursors of O₃.

The cross sections of SO₂, like those of O₃, did not show definite evidence of isentropic transport, possibly because of masking by local sources.

However, individual SO_2 plumes in stable stratification were identifiable at much the same isentropic surfaces throughout the day, as expected on the basis of the hypothesized isentropic transport mechanism.

Industrially emitted SO_2 oxidizes in the air and forms sulfate particles, which in turn make up much of the aerosol measured by nephelometry (q.v., Rodhe et al., 1972; Charlson et al., 1974; Waggoner et al., 1976). Hence, while it is difficult to determine SO_2 transport directly, secondary effects of SO_2 , i.e., sulfate particles, may be the primary constituent of the aerosols (represented by b_{scat}) that exhibit long-range transport.

Conclusion

Although vertical profiles of SO_2 and O_3 concentrations suggest that local sources are dominant, selected pairs of profiles give indications of possible isentropic transport. In some cases, behavior of SO_2 , O_3 , and b_{scat} characteristic of plumes can be traced over several hundred kilometers. While there is some sampling uncertainty, similar air-quality characteristics are observed at successive times at different locations along the same isentropic surface and thus are consistent with isentropic transport of pollutants.

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FORMULATION OF THE BOUNDARY CONDITIONS FOR DYNAMIC MODELS USING THE HEAT ENERGY BALANCE AT THE SURFACE

T. Yamada

Vertical variations of wind and temperature are greatest within the first 10 m or so above the surface of the earth. These changes may not be adequately represented by the vertical resolution in numerical models. One solution to this problem is to use the observed surface temperature and adopt empirical formulae (Businger, 1966; Dyer and Hicks, 1970) for the mean wind and temperature gradients in the surface layer over horizontally homogeneous terrain. However, observed temperatures may not always be easily adapted to a three-dimensional numerical model, since the observational network and sampling frequency can differ significantly from the grid density and integration time steps of the model, and thus require considerable spatial and temporal interpolation of observations before use in the model. An alternative approach is to compute the surface temperature from the heat energy balance equation at the surface.

In conditions in which vegetation does not play an active part, several methods of modeling the ground heat transfer by considering transfer through thin layers have been proposed. For one layer (slab), a time dependent equation for the surface temperature can be obtained. A review of these methods and a comparison of the predicted surface temperatures with those obtained by a multiple layer method have been given by Deardorff (1978). Here we use a multiple layer method due to its greater accuracy and more general applicability.

The temperature T_s in the soil layer is obtained by solving the heat conduction equation

$$\frac{\partial T_s}{\partial t} = \frac{\partial}{\partial z_s} \left(k_s \frac{\partial T_s}{\partial z_s} \right) \quad (1)$$

where z_s is positive downward, and soil diffusivity k_s could be a function of soil moisture content. Appropriate boundary conditions normally used to solve Eq. 1 are the heat energy balance at the soil surface and specification of the

soil temperature or soil heat flux at a certain depth (dependent on the duration of the integration). The heat energy balance at the surface is given by

$$R_s + R_L \downarrow - R_L \uparrow = H + LE + G \quad , \quad (2)$$

where R_s is the incoming solar radiation absorbed by the ground surface, $R_L \downarrow$ is the incoming longwave radiation, and $R_L \uparrow$ is the outgoing longwave radiation. The basic inputs for this model, R_s and $R_L \downarrow$, are discussed in detail by Yamada (1978a).

The surface heat flux H , latent heat flux LE and ground heat flux G are given by

$$H = -\rho_a c_p u_* T_* \quad , \quad (3)$$

$$LE = -\rho_a L u_* Q_* \quad , \quad (4)$$

and

$$G = -k_s \frac{\partial T_s}{\partial z_s} \Big|_G \quad , \quad (5)$$

where ρ_a is the air density, u_* is the friction velocity, T_* is the temperature scale, Q_* is the water vapor scale, and the subscript G denotes the value at the ground surface. Substituting Eqs. 3 to 5 into Eq. 2, we obtain

$$R_s = \epsilon R_L \downarrow - \epsilon \sigma T_G^4 = -\rho_a c_p u_* T_* \left(1 + \frac{1}{B}\right) - k_s \frac{\partial T_s}{\partial z_s} \Big|_G \quad , \quad (6)$$

where the relation

$$R_L \uparrow = \epsilon \sigma T_G^4 + (1 - \epsilon) R_L \downarrow \quad (7)$$

and Bowen ratio

$$B = H/LE \quad (8)$$

are used; ϵ is the emissivity of the surface and σ is the Stefan-Boltzman constant. Garratt and Hicks (1973) obtained a relationship between the surface temperature and air temperature at z_1 (in the surface layer):

$$\frac{\Theta(z_1) - \Theta_G}{T_*} = \frac{P_r}{k} \left[\ln \frac{z_1 + z_{0t}}{z_{0t}} + 2 - \psi \right] \quad , \quad (9)$$

where Θ is the potential temperature, P_r is the turbulence Prandtl number at neutral stability, k is the von Karman constant, z_{ot} is the roughness length for temperature and ψ is the stability correction term of Panofsky (1963). Using Eq. 9, we can eliminate T_* from Eq. 6 to obtain

$$R_s + \epsilon R_L \downarrow - \epsilon \sigma T_G^4 + m [\Theta(z_1) - T_G (P_o/P_G)^{R/c_p}] + k \frac{\partial T_s}{\partial z_s} \Big|_G = 0 \quad (10)$$

where

$$m = \frac{k \rho_a c_p u_* (1 + \frac{1}{B})}{P_r [\ln \frac{z_1 + z_{ot}}{z_{ot}} + 2 - \psi]} \quad (11)$$

P_o is a reference pressure (1000 mb) and P_G is the pressure at the surface. Equation 10 may be linearized by noting that $(T_G^{n+1} - T_G^n)/T_G^n \ll 1$, where the superscripts n and $n+1$ denote the n and $(n+1)$ th time steps of integration (a typical time increment used in integration is 10 min). After substitution into Eq. 10 of the approximation,

$$(T_G^{n+1})^4 \approx 4 (T_G^n)^3 T_G^{n+1} - 3 (T_G^n)^4 \quad (12)$$

we obtain

$$\begin{aligned} [4\epsilon \sigma (T_G^n)^3 + m(P_o/P_G)^{R/c_p} + R_s/\Delta z_s] T_G^{n+1} \\ = \frac{k_s}{\Delta z_s} T_s^{n+1}(1) + R_s + \epsilon R_L \downarrow + 3\epsilon \sigma (T_G^n)^4 + m\Theta^n(z_1) \end{aligned} \quad (13)$$

where the derivative $\partial T_s / \partial z_s \Big|_G$ is replaced by a forward finite-difference approximation $(T_s^{n+1}(1) - T_G^{n+1})/\Delta z_s$, $T_s(1)$ is the soil temperature at the first grid level from the surface, and Δz_s is the distance between the surface and the first grid level in the soil layer.

Equation 1 is solved numerically in finite-difference form by Laasonen (Richtmyer and Morton, 1967, p. 189). By this method, Eq. 1 reduces to $\underline{A} \underline{T}_a = \underline{B}$ where $\underline{A}$ is a tridiagonal matrix and $\underline{B}$ is a column vector. The solution is conveniently obtained by using the relation (Richtmyer and Morton, 1967, p. 198)

$$(T_s)_\ell = E_\ell (T_s)_{\ell+1} + F_\ell, \quad (14)$$

where $(T_s)_\ell$ is the soil temperature at the ℓ th grid level from the surface. Expressions for E and F when $\ell > 1$ are determined from the finite-difference form of Eq. 1, and Eq. 13 determines E_1 and F_1 . Comparing Eq. 13 with Eq. 14, we obtain

$$E_1 = \frac{k_s}{\Delta z_s} / [4\epsilon\sigma (T_G^n)^3 + m(P_o/P_G)^{R/c_p} + k_s/\Delta z_s], \quad (15)$$

and

$$F_1 = [k_s + \epsilon R_L + 3\epsilon\sigma (T_G^n)^4 + m\Theta^n(z_1)] / [4\epsilon\sigma (T_G^n)^3 + m(P_o/P_G)^{R/c_p} + (k_s/\Delta z_s)]. \quad (16)$$

Numerical integration of Eq. 1 by use of Eqs. 14–16 is rapid, since no iteration is required. Examples of diurnal variations of the surface temperature computed by this method are given by Yamada (1978b). It should be emphasized, however, that vegetation will play an active part in the apportionment of available heat energy between convective (sensible and latent) and conductive (into the soil) components. The technique discussed here is intended to address only the case of bare soil; the complexity introduced by biological factors remains under study.

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NUMERICAL SIMULATIONS OF AIRFLOW OVER THE NORTHEASTERN UNITED STATES: PRELIMINARY RESULTS

T. Yamada

A preliminary version of a three-dimensional dynamic model has been used previously to simulate airflow over Gaussian-shaped mountains (Yamada, 1978a; henceforth Y78). Simulations with this model for a stable atmosphere produced accelerations and decelerations of winds due to the presence of the mountains; strong subsidence ($\sim 2 \text{ m s}^{-1}$) was confined to the downward slope of the mountains, while weak upward motion ($\sim 0.2 \text{ m s}^{-1}$) occurred over much larger regions surrounding the mountains. Air directly above the mountains was much warmer and drier than the surrounding air at the same level. These solutions were obtained by specifying a constant surface temperature as a boundary condition.

The dynamic model has been recently improved in order to simulate more realistically the diurnal variations of airflow over mountains. The surface temperature is now computed from a heat energy balance at the ground. In order to compute soil heat flux at the surface, a heat conduction equation is solved for a soil layer 1 m deep (Yamada, 1978b). Furthermore, the amount of direct solar radiation received at an inclined surface is computed according to the formulas reviewed by Yamada (1978c).

Since the main features of the model are described in Y78, only a brief review of the model is necessary here. Governing equations are 1) the equation of motion for the horizontal winds, 2) the hydrostatic approximation, 3) the continuity equation from which the vertical wind is computed, 4) the conservation equations for liquid water potential temperature and mixing ratio of the total water content (vapor and liquid), and 5) the conservation equations for turbulence energy and a master-length scale of turbulence. It is noted that the liquid water potential temperature and the mixing ratio of total water content are conservative quantities even when phase change occurs. A terrain-following coordinate system is used to simplify the treatment of the surface boundary conditions. Boundary conditions at the top of the computational domain

and at the surface are given by Yamada (1978d).

A perspective view of the topography of the study area is shown in Figure 1. The Atlantic ocean is artificially lowered so that the coast can be seen clearly. The most distinguishable mountain ranges in the area are the Appalachian Mountains in the east and the Ozark Mountains in the southwest.

FIG. 1.--Perspective view of the topography of the study area.

Initially, a horizontally homogeneous field is assumed, and topography is gradually introduced into the model in order to reduce numerical disturbances. A uniform westerly wind of 10 m s^{-1} (except in the surface layer where the wind profile is logarithmic) is initially assumed. Initial temperatures (288 K in the mixed layer) increase at a rate of approximately 4 K km^{-1} in the inversion layer. Initial temperature in the soil layer is 291 K; the ocean temperature is kept constant at 293 K. Integration is continued for 5 hr with a time increment of 10 min. Computed horizontal wind vectors at 1100 EST at 0.2, 12, 340, and 3200 m above the surface are shown in Figures 2, 3, 4, and 5, respectively. Horizontal winds computed are, as expected, greatly affected by topography, particularly by the Appalachian Mountains. Another interesting feature is the acceleration of wind speed in the surface layer (Figure 2) over the Atlantic ocean due to roughness lengths being much smaller over water (typically $2 \times 10^{-5} \text{ m}$) than over land (10 cm is assumed). As expected, computed horizontal wind vectors become more uniform with increasing height. Although the simulated wind vectors shown here appear reasonable for the boundary conditions, further improvements are underway; for example, long wave radiation cooling in the atmosphere will be incorporated in order to simulate the nocturnal boundary layer.

FIG. 2.--Simulated horizontal wind vectors at 0.2 m above the ground. Grid points are located at the centers of the wind vectors. Wind speed is proportional to the length of the arrow, with an arrow of one grid spacing representing 10 m s^{-1} .

FIG. 3.--As in Figure 2 but at 12 m above the ground.

FIG. 4.--As in Figure 2 but at 340 m above the ground.

FIG. 5.--As in Figure 2 but at 3200 m above the ground.

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A SPECTRAL METHOD FOR SEPARATING LARGE-SCALE COMPONENTS OF A TIME SERIES

C. M. Sheih

Introduction

Proper separation of the values of a variable in turbulent flow into mean (large scale) and fluctuating (small scale) components is not always a straightforward matter. This problem is particularly acute in transient flows when mean conditions change within the sampling duration. For example, meteorological variables in the atmosphere are subject to changing mean flow conditions, and consequently the measured mean value of a variable depends upon the duration of the averaging time. Often, the averaging time is determined by intuition or by trial and error until one finds a range of times over which the averaged value changes little. Obviously, averaging times determined in this fashion are quite subjective and can cause discrepancies between the results of two independent analyses of identical data.

Since the process of time averaging effectively filters fluctuations with time scales smaller than the averaging time (or with frequencies larger than the inverse of the averaging time), it is logical to determine an appropriate averaging time by investigation of the spectrum, particularly if a frequency exists that separates large scale and small-scale components. The time duration associated with this frequency is then an appropriate averaging time, and one can reconstruct large-scale and small-scale components by use of Fourier series. This technique will be demonstrated below for an ideal signal consisting of a linear combination of high-frequency and low-frequency sine waves, and then will be applied to the analysis of a balloon trajectory.

Description of the Technique

1. A Sinusoidal Signal

Let an idealized signal consist of two sine wave components as follows,

$$y(t) = C_1 \sin 2\pi f_1 t + C_2 \sin 2\pi f_2 t , \quad (1)$$

where $y(t)$ is the signal to be analyzed, C_1 and C_2 are constants, f_1 and f_2 are frequencies, and subscripts 1 and 2 represent large-scale and small-scale components, respectively.

As an example, if $C_1 = 1$ m, $C_2 = 2$ m, $f_1 = 20$ s⁻¹, and $f_2 = 200$ s⁻¹, $y(t)$ will be as shown in Figure 1a. Let us suppose that $y(t)$ is given and the task is to find the large-scale and small-scale components of $y(t)$. This can be done by digitizing the signal and computing the Fourier coefficients of the time series of the signal with the following equation:

$$A(q\Delta f) = \sum_{p=0}^{N-1} y(p\Delta t) \exp(i2\pi pq/N), \quad \text{for } q=0, 1, \dots, N-1, \quad (2)$$

with $\Delta f = (N\Delta t)^{-1}$

and $f = q\Delta f$,

where A is a Fourier coefficient of y , $i = (-1)^{1/2}$, and N is the total number of data points in the time series of y . The spectral density of y can be computed from

$$S(f) = N^{-2} A(f) A^*(f) \quad (3)$$

with

$$\overline{y^2} = N^{-2} \sum_{q=0}^{N-1} S(q\Delta f), \quad (4)$$

where S is spectral density and A^* is the complex conjugate of A . In order to represent in graphical form the contribution to the total variance associated with each frequency range, the general practice is to plot the spectral density in an "area-conserving" relationship, i.e., $fS(f)$ vs. $\log f$. The analysis of the sinusoidal signal (Figure 1c) shows that any frequency within the range $3 \leq q \leq 10$ between the peaks of the spectrum will separate the large-scale and small-scale components. If this frequency is denoted by $f = M\Delta f$, then the large-scale and small-scale components constructed from the Fourier coefficients are

$$y_1(p\Delta t) = N^{-1} \sum_{q=0}^M A(q\Delta f) \exp(-i2\pi pq/N) \quad (5)$$

and

$$y_2(p\Delta t) = N^{-1} \sum_{q=M+1}^N A(q\Delta f) \exp(-i2\pi pq/N) \text{ for } p=0, \dots, N-1 \quad (6)$$

with

$$f_1 < M\Delta f < f_2 .$$

The sinusoidal signals computed from Eqs. 5 and 6 recover the original sinusoidal components satisfactorily (Figure 1b).

2. A Constant-Level Balloon Trajectory

In order to demonstrate the applicability of the technique to practical problems, the above scheme is used to analyze one of the constant-level balloon trajectories of Hart and Frenzen (1976). Figure 2a shows the lateral horizontal coordinate of the balloon as a function of time. The corresponding spectral density plotted in area-conserving coordinates is shown in Fig. 2c. It appears that the spectral density has a gap over the frequency range of $6 \leq q \leq 10$, which separates the trajectory into large-scale meandering and small-scale turbulence. Using $M=7$ in Eqs. 5 and 6, one can decompose the signal into large-scale and small-scale components (Figure 2b).

Conclusion

The present paper suggests an objective approach to the separation of mean and fluctuating parts of a variable in flows where mean conditions are changing with time. The method involves computing Fourier coefficients for the discrete time series of the variable. The Fourier coefficients enable us to compute the spectral density, which can then be plotted in an area-conserving curve. If there is a gap in the curve, any frequency within the gap can be used to separate the mean and fluctuating parts of the signal. If there is no gap in the curve, the mean value of the variable is not easily defined and depends upon the choice of the averaging time. Once a frequency suitable for scale separation is determined, the mean and fluctuating components of the variable can be constructed by use of Fourier series.

Although the present analysis is carried out with time as the independent variable, similar calculations could be made with a spatial coordinate.

FIG. 1.--Spectral decomposition of a sinusoidal signal: (a) original signal, (b) decomposed large-scale and small-scale components of the original signal, and (c) variance spectrum of the original signal.

FIG. 2.--Spectral decomposition of a constant-level balloon trajectory: (a) horizontal balloon trajectory, (b) decomposed large-scale and small-scale components of the balloon trajectory, and (c) variance spectrum of the original balloon trajectory.

Furthermore, the transformation from Fourier coefficients to time series is not always necessary, because many statistical quantities, such as the autocorrelation coefficient and the variance, can be obtained by directly integrating the Fourier coefficients. Computations with the present method should be rather inexpensive since the FFT (Fast Fourier Transform) computer subroutine is widely available today.

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COMPUTATIONAL STABILITY OF THE GAUSSIAN MOMENT-CONSERVATION TECHNIQUE OF NUMERICAL MODELING OF DIFFUSION

J. D. Shannon

Numerical modelers of advection and diffusion should use the longest integration time step that still results in stable and accurate solutions, since this minimizes computation costs. It thus becomes important to investigate the limits of computational stability in numerical schemes. The behavior of the Gaussian Moment-Conservation (GMC) technique (Shannon, 1979) as the time increment varies is examined below.

If a method is a mathematically "pure" scheme of numerical dispersion (i.e., equations are solved and the solution is used directly without adjustment), then the analytical solution of a test distribution, such as a harmonic wave, can be analyzed to determine the exact Fourier number ($F = K\Delta t/(\Delta x)^2$) at which computational instability begins. In the GMC model, however, the post-solution adjustment of pollutant masses to avoid negative concentrations requires evaluation to be more subjective.

A review of the equations in the GMC model (simplified from Shannon (1979) by assumption of the uniform Δx) is useful here:

$$M_{i,i-1} + M_{i,i} + M_{i,i+1} = 1 \quad , \quad (1)$$

$$-M_{i,i-1}\Delta x + M_{i,i+1}\Delta x = \mu_i \quad , \quad (2)$$

$$M_{i,i-1} \left(\frac{5(\Delta x)^2}{4} \right) + M_{i,i} \frac{(\Delta x)^2}{4} + M_{i,i+1} \left(\frac{5(\Delta x)^2}{4} \right) = \mu_i^2 + \sigma_i^2 \quad , \quad (3)$$

where

$$\mu_i = (\xi_i^+ + \xi_i^-) \Delta t/2 + u_i \Delta t \quad (4)$$

$$\sigma_i = \Delta x/2 + (\xi_i^+ - \xi_i^-) \Delta t/2 \quad , \quad (5)$$

and

$$\xi_i = \frac{K}{c} \frac{\partial c}{\partial x} \quad . \quad (6)$$

The equations refer to a normalized pollutant mass originally within cell i and assumed to have an original mean of 0 (relative to the center of cell i) and variance of $(\Delta x)^2/4$. $M_{i,i-1}$, $M_{i,i}$, and $M_{i,i+1}$ are the masses to be allotted to cells $i-1$, i , and $i+1$, respectively, after a time step of diffusion. The symbols μ_i and σ_i^2 refer to the mean and variance of the single cell puff after diffusion but before allotment to the three cell puffs. Symbols ξ_i^+ and ξ_i^- represent artificial diffusion velocities calculated at the boundaries between i and $i+1$ and between i and $i-1$, respectively; other variables are defined according to standard practice.

A simple distribution for analysis, but one providing a rigorous test, is initial zero concentration except for one grid cell (a spike). Pure diffusion ($u_i = 0$) will be examined. Because of the symmetry of the test distribution, $\xi_i^+ = -\xi_i^-$ and $M_{i,i-1} = M_{i,i+1}$. Rearrangement of terms leads to

$$M_{i,i} = 5/4 - \sigma_i^2 / \Delta x^2 \quad (7)$$

or

$$\Delta t = \frac{(\Delta x)^2}{2K} (\sqrt{5/4 - M_{i,i}} - 1/2) \quad (8)$$

The test examples have $\Delta x = 100$ m and $K = 10 \text{ m}^2 \text{ s}^{-1}$. If Δt is sufficiently large, $M_{i,i} < 1/3$ and the initial peak becomes a relative valley, at least for the initial time step. A time increment of $\Delta t = 230$ s ($F = 0.23$) leads to $M_{i,i} < 1/3$. As shown in Table 1, however, the solution becomes stable with succeeding time steps. If $M_{i,i} \leq 0$, the solution will alternate between positive values and zeros, because of the post-computation adjustment to avoid negative masses. A time increment $\Delta t = 320$ s ($F = 0.32$) leads to a non-positive $M_{i,i}$, as shown in Table 2. It can be noted that averaging of the n th and $(n+1)$ th iterations, where n is even, produces a reasonable solution field. However, since any time increment greater than 320 s produces the same result in this case ($M_{i,i-1} = M_{i,i+1} = 0.50$, $M_{i,i} = 0.0$ after adjustment), the reasonableness (i.e., Gaussian symmetry) does not necessarily imply accuracy.

When the initial distribution is smoother, the GMC technique is stable for larger Fourier numbers, but the critical values must be determined essentially by trial and error for particular cases. A Gaussian distribution

Table 1. Example of temporary computational instability for an initial spike distribution with Fourier number = 0.23.

Initial	1000									
Iteration 1	0	0	0	336	328	336	0	0	0	0
Iteration 2	0	0	95	221	367	221	95	0	0	0
Iteration 3	0	28	93	229	301	229	93	28	0	0
Iteration 4	8	35	109	212	271	221	109	35	8	8

Table 2. Example of computational instability for an initial spike distribution with Fourier number = 0.32.

Initial	1000							
Iteration 1			500	0	500			
Iteration 2		250	0	500	0	250		
Iteration 3	125	0	375	0	375	0	125	
$\frac{2+3}{2}$	63	125	188	250	188	125	63	

Table 3. Sensitivity of the GMC technique to the time increment after 7200 s. ($K, \Delta x$) = (10 m² s⁻¹, 100 m); ends of distributions are truncated.

$\Delta t = 60$ s	45	60	76	89	99	102	99	89	76	60	45
$\Delta t = 120$ s	45	60	76	89	98	102	98	89	76	60	45
$\Delta t = 240$ s	45	60	75	89	98	101	98	89	75	60	45
$\Delta t = 360$ s	45	60	75	88	97	100	97	88	75	60	45

across the grid, with a system standard deviation (as opposed to a cell puff σ) of Δx , can be diffused in a computationally stable manner by the GMC technique until the time step is increased to 400 s ($F=0.40$). The effect of varying the time step within the stable Fourier region can be seen in Table 3; a variation of a factor of six in the time step changes the solution after two hours of modeled dispersion by about 1 to 2%.

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A REVIEW OF DIRECT SOLAR RADIATION TO INCLINED SURFACES

T. Yamada

Over complex terrain the surface heat energy balance is quite different from that over horizontally homogeneous terrain. For example, during winter in the northern hemisphere, surfaces facing south receive much more direct solar radiation than those facing north; likewise, surfaces facing east receive more direct solar radiation in the morning and those facing west receive more in the afternoon. Differences in solar radiation received by the surface become one of the most important factors producing heterogeneous surface temperature distributions over complex terrain, the cause of many interesting local circulations. This article reviews the formulation of solar and long-wave radiation received by inclined surfaces. The results are incorporated in the heat energy balance at the surface (Yamada, 1978).

According to Kondrat'yev (1977, p. 1) the direct solar radiation S_s received by an inclined surface is

$$S_s = S_m \cos I , \quad (1)$$

where S_m is the near surface, direct solar radiation, and I is the angle of incidence of solar rays to the surface. The cosine of the angle of incidence can be expressed as

$$\cos I = \cos \alpha \cos Z + \sin \alpha \sin Z \cos(\gamma - \beta) , \quad (2)$$

where Z is the zenith angle of the sun, γ is the solar azimuth; and α and β are the angle of inclination of the surface relative to the horizontal plane and the azimuth of the projection of the normal to the surface on the horizontal plane, respectively. The azimuth angle is in reference to south and positive clockwise. Note that Eq. 1 reduces to the more familiar form for flat terrain when $\alpha=0$. The zenith angle and azimuth are determined from the following formulae:

$$\cos Z = \sin \phi \sin \delta + \cos \phi \cos \delta \cos H , \quad (3)$$

$$\sin \gamma = \frac{\sin H \cos \delta}{\sin Z} \quad \text{and} \quad \cos \gamma = \frac{\cos Z \sin \phi - \sin \delta}{\sin Z \cos \phi} , \quad (4a,b)$$

where ϕ is the latitude, δ is the declination of the sun, and H is the solar hour angle, positive clockwise from apparent noon. Equation 2 can then be rewritten as (Kondrat'yev, 1977, p. 3)

$$\cos I = C_1 + C_2 \cos H + C_3 \sin H , \quad (5)$$

where

$$C_1 = \sin \delta (\cos \alpha \sin \phi - \sin \alpha \cos \phi \cos \beta) , \quad (6a)$$

$$C_2 = \cos \delta (\cos \alpha \cos \phi + \sin \alpha \sin \phi \cos \beta) , \quad (6b)$$

$$C_3 = \sin \alpha \cos \delta \sin \beta . \quad (6c)$$

Since the maximum change in the solar declination δ in 24 hr is less than 0.5 deg, δ is assumed to be constant during a given day. Spencer (1971, quoted in Paltridge and Platt, 1976, p. 63) provides a formula to compute δ in radians,

$$\begin{aligned} \delta = & 0.006918 - 0.399912 \cos \theta_0 + 0.070257 \sin \theta_0 \\ & - 0.006758 \cos 2\theta_0 + 0.000907 \sin 2\theta_0 \\ & - 0.002697 \cos 3\theta_0 + 0.00148 \sin 3\theta_0 , \end{aligned} \quad (7)$$

where the angle θ_0 in radians is related to the Julian date J_d by

$$\theta_0 = 2\pi(J_d - 1)/365 . \quad (8)$$

Equation 7 estimates δ with a maximum error of 0.0006 radians. Solar hour H can be obtained if the longitude, clocktime and the equation of time are known. The equation of time is the difference between the local apparent time and a fixed mean solar time, which is derived from the motion of an imaginary body (the "mean sun") moving at uniform angular speed along the celestial equator at a rate equal to the average movement of the sun. The solar hour angle H is given in radians by

$$H = \pi(t_s - 12)/12 , \quad (9)$$

where t_s is the true solar time (local apparent time) in hours. The true solar

time is obtained from

$$t_s = t_{c.t.} + \Delta t_{long} + t_{eq} , \quad (10)$$

where $t_{c.t.}$, t_{long} , and t_{eq} are the clocktime, the longitude correction, and the equation of time, respectively. The longitude correction accounts for the difference between the local meridian and a standard meridian, and is positive if the local meridian is east of the standard. The equation of time is provided by Spencer (quoted in Paltridge and Platt, 1976, p. 63) as follows:

$$t_{eq} = \frac{12}{\pi} [0.000075 + 0.001868 \cos \theta_0 - 0.032077 \sin \theta_0 - 0.014615 \cos 2\theta_0 - 0.040849 \sin 2\theta_0] , \quad (11)$$

where t_{eq} is in hours and θ_0 is defined by Eq. 8. Equation 11 has a maximum error, compared with values tabulated in the National Almanac, of 35 s in time.

The angle of inclination α and azimuth β of a sloped surface are obtained from the geometric considerations:

$$\alpha = \tan^{-1} [(\partial z_g / \partial x)^2 + (\partial z_g / \partial y)^2]^{1/2} , \quad (12)$$

and

$$\beta = \frac{\pi}{2} - \tan^{-1} [\partial z_g / \partial y / \partial z_g / \partial x] , \quad (13)$$

where $z_g(x, y)$ is the ground elevation. Azimuth of a sloped surface β is defined clockwise from the south.

The amount of solar radiation reaching the surface is much less than that at the top of the atmosphere due to many factors, including molecular scattering and absorption by permanent gases such as oxygen, ozone, and carbon dioxide. This effect is parameterized by Atwater and Brown (1974), who modified the original form by Kondrat'yev (1969) to include the effect of the forward Rayleigh scattering. The expression is

$$G = 0.485 + 0.515 [1.041 - 0.16 \left(\frac{0.000949p + 0.051}{\cos Z} \right)^{1/2}] , \quad (14)$$

where p is pressure in mb. Other important factors that also modify the

amount of incoming solar radiation include water vapor, clouds, and airborne particles. Parameterizations for these factors will be included in the model in the near future.

Currently S_m is calculated from

$$S_m = S_o G , \quad (15)$$

where S_o is the incoming radiation flux at the top of the atmosphere and G is given by Eq. 14. Thus, the amount of incoming radiation absorbed by a sloped surface is

$$\begin{aligned} R_s &= S_s (1 - A) \\ &= S_o G (1 - A) \cos I , \end{aligned} \quad (16)$$

where A is the albedo of the surface.

Finally, the long-wave incoming radiation at the surface is parameterized according to Swinbank's empirical formula (quoted in Platt and Platt, 1976, p. 139) ,

$$R_L \downarrow = 5.31 \times 10^{-13} T^6 \cos \alpha , \quad (17)$$

where $R_L \downarrow$ is in $W m^{-2}$ and T is the screen temperature. Swinbank's regression Eq. 17 is based on nighttime data and is therefore reduced by about $20 W m^{-2}$ to compute daytime values (Platridge and Platt, 1976, p. 139). Long-wave radiation transfer in the atmosphere will be included in the model in the near future. Examples of diurnal variations of the surface temperature computed by using the radiation fluxes discussed here are given in an accompanying article in this report.

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ON THE USE OF A KYTOON SYSTEM TO MONITOR THE NOCTURNAL PBL

R. L. Hart

A realistic model of regional-scale dispersion of pollutants should include diurnal variations in the structure of the PBL, particularly during the stable, nighttime hours, since variations in the height of nocturnal inversions and in the strength of nocturnal, low-level wind maxima (Sisterson and Frenzen, 1978) can have significant effects in diffusion and transport of pollutants (Shannon, 1979). While the structure of the well-mixed layer during daytime hours is adequately measured with radiosondes, sodars, and lidars, these methods do not have sufficient resolution to measure the complex temperature and wind variations occurring near the surface during the onset and during the collapse of the nocturnal PBL.

The ANL radiosonde system (Hart et al., 1978) in conjunction with a double-theodolite, balloon-tracking system (the WHAT system, for wind, height, and temperature, as reported by Frenzen and Prucha, 1973) has been successfully used for the past several years to study the detailed vertical structure of temperature and wind above 30 m in the nocturnal PBL (e.g., see Hess et al., 1976). However, the lack of accurate wind data below 30 m from the WHAT system has prompted trials of alternate systems. During the 1976 Sangamon experiment (Sisterson et al., 1976a) an instrument package suspended from the tether of a kytoon, a tethered balloon with aerodynamic lift (Figure 1) was used to obtain profiles of temperature, wind speed, and wind direction in the lowest 250 m. This report compares kytoon and radiosonde data obtained during the 1976 Sangamon experiment.

Equipment

The instrument package consisted of batteries, temperature and wind speed sensors, signal conditioners, and a two-channel telemetry transmitter. The output of the Yellow Springs Instrument Company thermoliner thermistor was converted by the telemetry system to a frequency linear with temperature.

FIG. 1.--The kytoon with the temperature and wind telemetry package attached to the tether line.

A single baseline check of temperature versus frequency before each run was the only necessary calibration of this portion of the system. A photo-chopper anemometer using an R. M. Young two-bladed propellor and a modified bivanne tail was suspended in a manner allowing orientation in the prevailing wind direction. The pulse train produced by the anemometer was transmitted to the ground station.

Accurate data on height, package motion, and wind direction were obtained by tracking the package with the double-theodolite system. Instantaneous elevation and azimuth angles from the two theodolites and integrated averages of the temperature and wind speed signals were recorded on digital magnetic tape at 1 s intervals.

Analysis

Wind speeds are obtained from the kytoon data at 10, 20, and 30 m by averaging the anemometer measurements within 5 m either side of the respective heights; this corresponds to time averaging over about 13 s intervals at typical ascent or descent rates of 0.75 m s^{-1} . The average wind speeds are corrected for motion of the package by removing the mean package speed during the averaging interval. Lateral motion of the package due to turbulence or instability of the kytoon cannot always be completely removed from the averages, but lateral motion is usually small. Wind speeds at 50, 75, and 100 m are obtained similarly by averaging over 20 m (approximately 27 s) intervals. The averaging interval above 100 m for wind speeds is increased to 25 m, which corresponds to a time interval of about 33 s.

Wind velocity profiles measured by the ANL radiosonde-WHAT system are obtained from 9 s differences of balloon position below 100 m and from 18 s differences above 100 m. The velocities are then linearly interpolated at standard heights of 30 m, 50 m, 75 m, 100 m, and every 50 m thereafter (Sisterson et al., 1976b). A comparison between radiosonde and kytoon wind speed profiles taken at approximately the same time (usually within 10 min of one another) shows differences as large as 20%, but this is not unexpected, since the averaging processes are not identical. However, these differences

should average out over a large number of runs. Averages and standard errors of 37 runs in which the kytoon reached a height of 150 m or more and in which kytoon and radiosonde profiles were taken within 15 min of one another are shown in Figure 2. At every level the average wind speed obtained from radiosonde runs and that obtained from kytoon runs are within the standard error of the estimate of the other. Additional wind speed averages obtained from the kytoon at 10, 20, and 125 m and from radiosondes above 150 m are included to complete the profile.

Wind directions corresponding to each wind speed obtained from the kytoon data are estimated with the azimuth angle between the mean package position during the averaging interval and the tether point at the ground. No correction is made for the fact that this angle is determined by the wind direction at the height of the kytoon, which was usually 5 to 10 m above the instrument package. Furthermore, the simple estimation method may have significant error over the short averaging intervals when there is marked directional shear. While wind directions obtained from the kytoon data agree reasonably well with those from radiosonde data, a wind direction sensor on the kytoon package is necessary for improved accuracy.

Highly detailed temperature profiles can be produced from the 1 s average temperatures measured by the kytoon and corresponding heights calculated from the theodolite data (Figure 3). The temperature values are accurate to $\pm 0.15^\circ\text{C}$ with a resolution of 0.03°C . Temperatures obtained from radiosondes have a resolution of 0.05°C , but are accurate only to $\pm 0.4^\circ\text{C}$. Since the radiosonde temperature sensor has a 2.7 s time constant and rise rates are normally 3 to 4 m s^{-1} , the sensor rises about 10 m during one time constant. On the other hand, while the thermoliner thermistor used in the kytoon package has a 4.5 s time constant, the lower ascent or descent rate results in the sensor rising or falling only about 3.4 m during one time constant. The averaged temperature values recorded by the WHAT system are corrected for the "exponential" smoothing imposed by sensor lag as follows:

$$x_t = \bar{x}_t + \frac{\tau}{\Delta t} (\bar{x}_t - \bar{x}_{t-1}) ,$$

FIG. 2.--Average of 37 kytoon (▽) and radiosonde (Δ) wind profiles taken within 15 min of one another. Standard errors of the estimate are shown for each point.

FIG. 3.--A typical temperature profile obtained with the kytoon system.

where X_t is the true value of the temperature at time t , $\bar{X}_t$ and $\bar{X}_{t-1}$ are the smoothed and averaged temperature values recorded at time t and $t-1$, respectively, τ is the sensor time constant, and Δt is the averaging interval (Panofsky and Brier, 1968). The corrected profiles from the two systems agree quite well (Figure 4).

Conclusions

Highly detailed profiles of wind speed and temperature in the lowest several hundred meters of the atmosphere, which are especially useful in monitoring the evolution of the surface-based inversion during the onset and during the collapse of the nocturnal PBL, have been obtained with a kytoon system. To thoroughly document the diurnal cycle of the PBL, supplementary profiles obtained from other techniques such as radiosondes are necessary, especially when the height of the well-mixed layer is 1 to 2 km or when wind speeds are too high for stable kytoon operation.

FIG. 4.--A kytoon temperature profile (solid line) and a radiosonde temperature profile (dashed line) taken within 15 min of one another.

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CORRELATIONS BETWEEN OBSERVED WINDS AND COMPUTED SURFACE GEOSTROPHIC VELOCITIES FOR THE 1975 SANGAMON EXPERIMENT

T. Yamada

The surface geostrophic winds calculated for the 1975 Sangamon experiment have been reported elsewhere (Yamada et al., 1977). The geostrophic winds were calculated from the pressure records collected at a network consisting of three microbarograph stations (see Lipschutz, 1975) and 14 National Weather Service (NWS) stations, and from the three microbarograph stations alone. In order to check the reliability of the geostrophic wind calculations, correlation coefficients between these geostrophic winds and horizontal winds measured by the double-theodolite balloon-tracking "WHAT" system (Frenzen and Prucha, 1974) have been computed (Figures 1 and 2) at 30, 50, 75, and 100 m, then at 50 m increments up to 2000 m above the surface. Results have been separated according to atmospheric stability. For the unstable case (Figures 1a and 2a) the computed geostrophic winds are highly correlated with observations, particularly below 100 m; a maximum correlation coefficient of 0.94 is obtained. However, the correlations for the stable case (Figures 1b and 2b) are much lower, perhaps as a result of the nocturnal inertial oscillation of winds around the geostrophic winds (see Blackadar, 1957).

The correlations for all cases combined are shown in Figures 1c and 2c. The correlations involving geostrophic winds computed from the microbarograph network alone (Figure 2c) are slightly lower than those involving the combined NWS/microbarograph network (Figure 1c). This may be due to the fact that the microbarograph network correlations are computed with a greater portion of stable cases than are correlations for the NWS/microbarograph network. The average number of data points at each level are shown in Table 1. Correlations, particularly for the east-west component, become lower as the height increases, possibly as a consequence of the change of geostrophic wind with height (the thermal wind). In the present case, thermal winds have been computed (Figure 3) from the screen height temperatures of the NWS network by a method similar to that for geostrophic winds; the surface pressures in

FIG. 1.--Correlation coefficients between observed wind components and the geostrophic wind components computed from the pressure records at the network consisting of the three microbarograph stations and 14 NWS stations for the unstable (a), stable (b), and combined (c) cases. Correlation coefficients for the east-west and north-south components are plotted as solid lines and dashed lines, respectively.

FIG. 2.--As in Figure 1 for the geostrophic winds obtained from the microbarograph stations alone.

Table 1. Average number of observations used in computation of correlation coefficients.

	Unstable	Stable	Total
NWS/microbarograph	80	45	125
Microbarograph	35	55	90

Eq. 2 of Yamada et al. (1977) are replaced by the surface temperatures. Values of $|\partial u_g / \partial z|$ and $|\partial v_g / \partial z|$ of the order of $2 \times 10^{-3} \text{ s}^{-1}$ appear to be typical, indicating the importance of thermal wind components in comparisons between observed and geostrophic winds.

FIG. 3.--Calculated thermal winds for the 1975 Sangamon experiment.

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GEOSTROPHIC WIND COMPONENTS NEAR THE BOULDER ATMOSPHERIC OBSERVATORY DURING APRIL 1978

B. B. Hicks, T. Yamada, F. C. Kulhanek, and R. G. Everett

The Boulder Atmospheric Observatory (BAO), located at Erie, Colorado, is a 300 m instrumented tower erected and operated by the Wave Propagation Laboratory of NOAA as a means for testing remote probing systems. The location is some 30 km east of Boulder, which is at the foothills of the Rocky Mountains. It is likely, therefore, that winds in the vicinity of the tower will be dominated by surface characteristics much of the time. In April 1978, a cooperative study of wind patterns in the area was conducted in order to determine the circumstances under which winds are orographically controlled and, subsequently, to identify those occasions in which the tower can be used as a platform for studies of the planetary boundary layer. As a contribution to the overall endeavor, this laboratory conducted an investigation of geostrophic wind components, and compared these with velocities measured at the tower. In this way, it was anticipated that upper-level wind directions suitable for investigation of the relationship between flow patterns in the lower atmosphere and geostrophic components would be identified, if indeed such situations exist.

In the same manner as in previous experiments (q.v., Lipschutz, 1975; Yamada et al., 1977), carefully exposed pressure sensors were set up at selected sites in the area of interest. These sites (Figure 1) were at Kimball, Nebraska ($41^{\circ}14'13''$ N; $103^{\circ}40'01''$ W; elevation 1439 m above mean sea level), Limon, Colorado ($39^{\circ}11'06''$ N; $103^{\circ}41'02''$ W; 1696 m MSL), and Erie, Colorado ($40^{\circ}02'46''$ N; $105^{\circ}00'29''$ W; 1585 m MSL). The pressure sensors were marine microbarographs, each vented to the air through a small hole at the center of a flat, circular disk mounted horizontally at the top of a 1.5 m pole. This method of exposure was intended to minimize dynamic pressure effects. A fourth, digital barometer was hand-carried to each site in sequence throughout the experiment, as a continuing verification of the readings of the individual sensors.

FIG. 1.--A sketch of the area surrounding the BAO, showing the sites used for the ANL pressure network at Kimball (1), Limon (2), and Erie (3). The BAO tower is located at Erie.

On twenty occasions it was possible to compare velocities measured by anemometers mounted on the BAO tower with geostrophic wind components evaluated from the pressure data by application of the methods described by Lipschutz (1975). As in the case of previous experiments, the results reported here are based on analyses of the observed pressure field, after reduction to a standard level (sea level). Figure 2 illustrates the results. Points plotted represent the origins of corresponding velocity vectors; thus a point plotted in the upper right quadrant indicates a wind from the northeast. The lines joining the pairs of symbols represent velocity differences; the average speed difference calculated from these data is 13.5 m s^{-1} , with a standard deviation of 8.2 m s^{-1} . This speed differential exceeds the average geostrophic wind speed (12.5 m s^{-1}).

Figure 2 is based on observations made at a variety of times and in a range of conditions. Good agreement is not necessarily expected, especially in the case of observations made at night (when strongly stratified flow can occur, sometimes leading to a pronounced inertial effect; see Yamada, 1978). It is informative to compare the data of Figure 2 with similar results obtained over more uniform terrain. Figure 3 shows that the 1975 Sangamon data (see Hess and Hicks, 1975) provide much better agreement between observed winds and geostrophic components. In this case, the average speed difference is 3.8 m s^{-1} ($\pm 2.8 \text{ m s}^{-1}$), compared with an average geostrophic speed of about 7.0 m s^{-1} .

FIG. 2.--Comparisons between winds measured at 300 m height at the BAO (solid circles) and surface geostrophic winds (open circles) derived from the network of Figure 1. Dashed lines indicate observations made at night.

FIG. 3.--Comparisons between winds measured by the WHAT system (double theodolites and pilot balloons) during the 1975 Sangamon experiment in central Illinois (solid circles) and geostrophic winds determined from the ANL pressure network (open circles). As in Figure 2, measured winds are at 300 m height. Dashed lines indicate observations made at night.

A further difference between the observations contributing to Figures 2 and 3 should be explained. Whereas in Figure 2 the measured velocities are obtained as hourly averages, those used in Figure 3 are derived from individual radiosonde ascents and accordingly are susceptible to considerably more error. Even so, on almost every occasion the wind speed differences evident in Figure 2 exceed those of Figure 3.

The intent of this study was to identify circumstances in which the Boulder Atmospheric Observatory could be used in fundamental studies of the nature and evolution of the planetary boundary layer, especially in relationship to the geostrophic driving mechanism. From the data obtained during April 1978, it is not possible to conclude that suitable wind directions exist.

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THE ARGONNE WIND ENERGY STUDY

D. L. Sisterson and B. B. Hicks

A study of the nature, frequency, and distribution of nocturnal, low-level wind maxima in the Midwest and Great Plains is making use of wind data from an instrumented television tower in central Oklahoma (Carter, 1970) and the ANL meteorology tower in northern Illinois. Since the use of power-law wind profiles is common in engineering design studies, power laws have been fitted to actual wind profiles, and the frequency distribution of the power-law exponent has then been examined according to hourly, seasonal, and annual time intervals. This report summarizes the results obtained from the ANL tower data; it is basically an extension of earlier analyses presented by Moses and Bogner (1967).

The present studies were initiated following an analysis of serial wind-profile measurements in central Illinois (Sisterson and Frenzen, 1978) which indicated that decoupled, low-level wind maxima form more frequently and at significantly lower altitudes than had previously been supposed. From the examination of other investigations (Table 1), the heights of such wind maxima appear to be lowest in the Great Plains and in the Midwest, an area which is largely unrecognized as a potential source for wind energy.

Analyses of the Argonne Data

A four-year period of wind data at Argonne was analyzed by Moses and Bogner (1967) in a study of the frequency distribution of the power-law exponent (p), based on wind data between heights of 5 and 50 m. Figure 1 has been constructed to show that there is a wide distribution of the best-fitting value with a clear seasonal variation; median values range from 0.14 in winter to more than 0.20 in summer. The annual median value is 0.17. Furthermore, Figure 2 shows a large diurnal variation of p -values with the largest values (indicating that decoupling is more likely) occurring at night. In a previous study (Sisterson and Frenzen, 1978), a power-law exponents greater than 0.4 were found to be associated with nocturnal, low-level wind maxima. A value

Table 1. Table of nocturnal, low-level wind observations found in the literature.

Investigator	Location	Method of investigation	Height of max., m	Time of max., (L.T.)	Seasonal dependence	Period of observation
Barad (1961)	O'Neill, Nebraska 42°N	Loeser balloon technique	250-600	0300	not available	6 weeks in the fall of 1953
Sisterson and Frenzen (1978)	Springfield, Illinois 40°N	WHAT system-pilot balloons	150	0500	not available	July-August 1975; July-August 1976
Crawford and Hudson (1970)	Oklahoma City, Oklahoma 35°N	1602 ft tower	355-455	0000-0100	not available	June 1966-May 1967
Jehn and Durie (1963)	Dallas, Texas 33°N	1400 ft tower	91-435	0300	most frequent in July (80%); least frequent in winter (40%)	606 days in January 1961-December 1962
Ridger and Armendariz (1971)	White Sands Missile Range, New Mex. 33°N	radar-tracked balloons	100-600	1900-0700	most frequent in March-June; secondary in November; least in July-September	1969-1970; 91 jets occurred on 154 nights
Farquharson (1939)	Khartoum, Africa 14°N	single theodolite pilot balloons	300	0600-0900	most frequent January-March; secondary in July	315 days over 11 months
Goualt (1938)	Fort Lamy, Africa 12°N	single theodolite pilot balloons	500	0500	most frequent in winter	April 1936-March 1937

FIG. 1.--Probability of the value of the power-law exponent $\leq p$, based on the frequency distribution of the power-law exponent for the Argonne site. (ANL Neg. 149-79-55)

of $p \geq 0.4$ tends to be associated with an Obukhov length $0 < L < 10$ m, a value signaling decoupling of winds aloft from the restraining influence of surface drag. During these very stable conditions, low-level wind maxima are most likely to occur. Since the Argonne tower is not high enough to detect the maximum wind speed, Figure 3 should not be considered to be a definitive evaluation of the frequency of these phenomena; however, it does provide a representation of power-law wind profile estimates in stable flow.

It is clear from the figures presented here that a constant-power law will underestimate wind energy potential unless due allowance is made for the stability dependence of p . This is particularly obvious in the short-term diurnal case. For long-term studies of wind energy potential, however, a simple power law with a constant $p = 1/7$ has often been assumed to provide a fairly good but conservative estimate. The magnitude of error involved has been tested with the ANL data. On the average, velocities of winds at 50 m estimated by applying a $1/7$ power law to velocities measured at 5 m were 20% too low. The wind energy potential (the cube of the wind speed) was underestimated by about 40%. Since the roughness length of the Argonne site (10 to 15 cm) is not unlike that of the farmlands of the Midwest and Great Plains, particularly in summer and fall when crops are tall, the error in calculating wind energy with $p = 1/7$ may be significant in these regions.

Nocturnal, low-level wind maxima are associated with near-zero wind velocities close to the surface and unusually large vertical gradients of horizontal winds in the first 30 to 150 m above the surface, as can be seen in the data of Sisterson and Frenzen. Although more complex methods to estimate low-level wind profiles characterizing the momentum-decoupled regime exist, a simple power law can be used to approximate the resulting wind shears in these conditions, provided an appropriate p -value is used. The nomogram presented in Figure 4 is intended for use in evaluating wind shear for profiles characterized by a power law of the form

$$\frac{u_t}{u_b} = \left(\frac{z_t}{z_b} \right)^p,$$

FIG. 2.--Median power-law exponent as a function of the time of day for the months of February (F), May (M), August (A), and November (N).
(ANL Neg. 149-79-52)

FIG. 3.--Probability of occurrence of the power-law exponent $p \geq 0.4$ as a function of the time of day for the months of February (F), May (M), August (A), and November (N). The power-law exponent $p = 0.4$ is taken to signal decoupled conditions, when nocturnal wind maxima are strongest.
(ANL Neg. 149-79-54)

FIG. 4.--Wind shear nomogram based on the power-law exponent (p) and the diameter (D) of a wind machine for a ground/blade clearance of 10 m. The wind shear is expressed in terms of multiples of wind speed experienced at 10 m, the standard climatological reference height.
(ANL Neg. 149-79-53)

where p is the power-law exponent and u_t and u_b represent the wind speeds experienced by rotor blades of large wind machines at the highest point (z_t) and lowest point (z_b) during rotation. In Figure 4, $z_b = 10$ m (the clearance between blade and ground) and $z_t = D + 10$ m, where D is the diameter of the wind machine. Since the frequency distribution of p determined for the Argonne tower is expected to be similar to that for the Midwest and Great Plains, reliable shears for these regions should be obtained.

Conclusions

From the ANL data, the hourly, seasonal, and annual frequency distribution of the power-law exponent based on actual wind profiles has been assessed. The data show that stable wind profiles with p -values in excess of 0.4, indicative of decoupled conditions, occur most frequently in summer. Wind energy potential from actual wind data at 50 m is shown to be underestimated by 40% when lower level winds are extrapolated by a $1/7$ power law. These results are expected to be representative of general conditions in the Midwest and Great Plains. From the frequency distribution of the power-law exponent, wind shears associated with nocturnal wind maxima are readily estimated.

Since wind profiles depend upon stability and surface stress, a simple power law cannot accurately describe a wind profile, particularly in stable conditions. For this reason, we do not recommend its general use. There are better, although more complicated, methods for describing wind profiles. However, for areas where stability and surface stress measurements are too costly or not available, a power law may be used as a conservative estimate of wind energy potential and wind shear, once the distribution of the power-law exponent has been determined.

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WINDMILL IMBALANCE MOMENT INDUCED BY VERTICAL GRADIENTS OF WIND SPEED AND DIRECTION

C. M. Sheih

Introduction

An important factor to consider in designing a horizontal-axis windmill is the imbalance moment caused by the windmill blades experiencing vertical gradients of horizontal wind speed and direction. Rotating windmill blades are subject to lift forces in the direction of rotation and drag forces in the direction of the wind. Lift forces are the primary sources of the net turning moment; drag forces do not contribute to the extraction of wind energy and may cause structural design problems. If the wind is constant in space and time, then the drag will appear only as a pressure or tension force between the blades and the supporting structure. However, if the wind velocity is not uniform, as is usually the case, different drag forces will act on different blades and will result in an imbalance moment with respect to the axis of rotation, in addition to the pressure force mentioned earlier. Since the radius of the moment exerted by the wind on the blades could be two orders of magnitude larger than that of the bearing assembly supporting the blades, the corresponding imbalance drag force could be amplified by two orders of magnitude when transmitted to the bearing. Further, this difference in drag force oscillates sinusoidally as the blades rotate through regions of higher and lower wind speed, placing even more strain on the structure. The present paper investigates the imbalance moment as a function of vertical gradients of wind speed and direction for various atmospheric stabilities and windmill dimensions.

Theoretical Considerations

Let us consider a simplified double-bladed windmill as shown in Figure 1a, where the blades of radial length R and width b projected on the plane of rotation are at angle θ with respect to the horizontal plane, and the axis of rotation is at height z_c from the ground. Further, the wind velocity

(u) is assumed to be

$$u = u_c \left(1 + \frac{r}{z_c} \sin \theta\right)^p \cos\left(\alpha \frac{r}{100} \sin \theta\right) , \quad (1)$$

where u_c is the wind speed at hub-height z_c , r is the radial distance from the axis of rotation, p is the exponent of the power-law velocity profile and α , in degrees per 100 m, is the vertical gradient of horizontal wind direction. Both p and α are functions of atmospheric stability, surface roughness, etc. The drag forces (see Figure 1b) acting at the radial distance r on the radial segment dr of the upper (F_1) and lower (F_2) blades are, respectively,

$$F_1 = \rho C_D u_c^2 \left(1 + \frac{r}{z_c} \sin \theta\right)^{2p} \cos^2\left(\alpha \frac{r}{100} \sin \theta\right) b dr , \quad (2)$$

$$F_2 = \rho C_D u_c^2 \left(1 - \frac{r}{100} \sin \theta\right)^{2p} \cos^2\left(\alpha \frac{r}{100} \sin \theta\right) b dr , \quad (3)$$

where ρ is the air density and C_D is the drag coefficient. The imbalance moment (M) is

$$M = \rho b C_D u_c^2 \int_0^R \left[\left(1 + \frac{r}{z_c} \sin \theta\right)^{2p} - \left(1 - \frac{r}{z_c} \sin \theta\right)^{2p} \right] \cdot \cos^2\left(\alpha \frac{r}{100} \sin \theta\right) r dr . \quad (4)$$

FIG. 1.--The configuration of an idealized windmill.

If $\xi = r/R$, Eq. 4 can be transformed and written as follows,

$$\begin{aligned}\tilde{M} &= M(\rho C_D b R^2 u_c^2)^{-1} \\ &= \int_0^1 \left[\left(1 + \frac{R}{z_c} \sin \theta\right)^{2p} - \left(1 - \frac{R}{z_c} \sin \theta\right)^{2p} \right] \cos^2 \left(\alpha \frac{R}{100} \xi \sin \theta \right) \xi d\xi .\end{aligned}\quad (5)$$

Obviously, the imbalance moment becomes zero when $\theta=0$ or π and a maximum when $\theta=\pi/2$, and the imbalance moment oscillates between these two values during blade rotation.

For structure design, maximum ($\tilde{M}_{\max}$) and average ($\tilde{M}_{\text{av}}$) values of the moment are of greatest interest. From Eq. 5 we can derive

$$\tilde{M}_{\max} = \int_0^1 \left[\left(1 + \frac{R}{z_c}\right)^{2p} - \left(1 - \frac{R}{z_c}\right)^{2p} \right] \cos^2 \left(\alpha \frac{R}{100} \xi \right) \xi d\xi , \quad (6)$$

$$\begin{aligned}\tilde{M}_{\text{av}} &= \frac{1}{\pi} \int_0^\pi \int_0^1 \left[\left(1 + \frac{R}{z_c} \sin \theta\right)^{2p} - \left(1 - \frac{R}{z_c} \sin \theta\right)^{2p} \right] \\ &\quad \times \cos^2 \left(\alpha \frac{R}{100} \xi \sin \theta \right) \xi d\xi d\theta .\end{aligned}\quad (7)$$

Equation 5 can be further simplified by the constraint $R/z_c < 1$, since the blades do not touch the ground. Retaining terms through second-order in a Taylor's expansion for the exponential terms and integrating, one can simplify Eq. 5 to

$$\tilde{M} \approx \frac{2pR}{z_c} \sin \beta \left(\frac{1}{3} + \frac{1}{\beta} \sin \beta + \frac{2}{\beta^2} \cos \beta - \frac{1}{\beta^3} \sin \beta \right) , \quad (8)$$

where $\beta = \frac{\alpha R}{50} \sin \theta$.

Results and Discussion

In order to investigate the stability dependence of the imbalance moment with the conventional power-law approach, which neglects the vertical gradient of horizontal wind direction, Eqs. 6 and 7 have been integrated numerically with $\alpha=0$, even though α depends upon atmospheric stabilities. The values

used for the exponent p (Table 1) are 0.1 for very unstable, $1/7$ for neutral and 0.8 for very stable conditions (Haltiner and Martin, 1957). The correspondence of these stabilities to the well-known Pasquill stabilities is also shown in the table. The windmill is assumed to be at the height $z_c = 100$ m.

Table 1. Exponent (p) of power-law wind velocity profile and vertical gradient (α) of horizontal wind direction as functions of atmospheric stability. The values of α are likely to apply over a "perfect" site and should not be considered to be representative of any particular location until site-specific information is obtained.

Pasquill stability classification	p	α , deg. 100/m
A (very unstable)	0.1	5
B		10
C		15
D (neutral)	0.141 (or $1/7$)	20
E		25
F (very stable)	0.8	35

The results of the computation are shown in Figure 2, where the subscript 0 denoted $\alpha = 0$. However, the difference between very unstable and neutral conditions is small. The values of $(\tilde{M}_{\max})_0$ are about 1.6 times their corresponding $(\tilde{M}_{av})_0$. The relative magnitudes of the moments, for both $(\tilde{M}_{\max})_0$ and $(\tilde{M}_{av})_0$ are 1, 1.4, and 8 for very unstable, neutral, and very stable conditions, respectively (note that these values are proportional to the assumed value of p). The normalized moments increase almost linearly with respect to the ratio R/z_c . From Eq. 5, this implies that the moment is proportional to R^3 , indicating that reducing the blade length might be more effective in preventing structural damage than the conventional method of increasing the pitch angle of the blades.

FIG. 2.--Maximum and average imbalance moments (normalized) as functions of the ratio of windmill blade radial length and the height of the rotating axis for very unstable (O), neutral (Δ), and very stable ($\square$) atmospheric stabilities.

To investigate the effect of the vertical gradient of horizontal wind direction, Eqs. 6 and 7 are integrated numerically for various values of α . Table 1 shows the values of α summarized by Irwin (1979) for Pasquill stability classification A to F. However, only the stabilities, A, D, and F, which correspond to the cases investigated earlier, are computed. The values of $\tilde{M}_{\max}$ and $\tilde{M}_{\text{av}}$ for $\alpha \neq 0$ are divided by the corresponding values computed earlier for $\alpha = 0$. These ratios for various stabilities and blade sizes are shown in Figure 3. Since all these ratios are smaller than unity, the vertical gradient of wind direction reduces the imbalance moment. The drag force is affected only by the wind velocity component parallel to the rotating axis; although the wind speed increases with height, the deviation of the wind direction from the rotating axis reduces the total effective wind velocity component experienced by the windmill. The effect of the wind direction variation of the moment is not dramatic; there is practically no effect on the unstable case throughout the range of blade sizes considered. The maximum attenuations due to wind direction variation are 6% for the neutral condition and 17% for the stable condition.

FIG. 3.--The ratios of the imbalance moments with and without consideration of vertical gradients of horizontal wind direction for very unstable (O), neutral (Δ), and very stable (□) atmospheric conditions.

Conclusion

Investigation of the imbalance moment on a windmill as a function of atmospheric stability and windmill blade size leads to a simple, analytical equation for estimating the imbalance moment for a stationary windmill. Computation of imbalance moments from power-law wind velocity profiles and vertical gradients of horizontal wind direction for various atmospheric stabilities reported in the literature indicates that the vertical wind speed variation affects the imbalance moment significantly with relative magnitudes of 1, 1.4, and 8 for very stable, neutral, and very unstable conditions, respectively. However, the effect of the vertical gradient of horizontal wind direction is quite moderate. For windmills with rotating axes at 100 m and blade radial lengths from 10 to 90 m, the reduction in the imbalance moment from the wind directional variation is at most 17% of the imbalance moment in corresponding cases with no horizontal wind directional variation.

Since many uncertainties remain, such as the wind velocity power-law exponent and the vertical gradient of the horizontal wind direction, the above results should be tested with field experiments in which the profiles of wind speed and horizontal direction are directly measured and the resulting

imbalance moments are calculated. Although the above analysis has been made for a double-bladed windmill, extension to other multiple blades should be rather straightforward.

Furthermore, the instantaneous velocity profile rather than the mean profile should be considered; organized vortices, not seen in mean velocity profiles, can substantially increase the vertical gradient of the instantaneous wind profiles. Since windmills are immune from the effects of certain high frequency components of the dynamic wind forces because of inertia, spectral densities of the imbalance moment should be investigated also.

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THE USE OF THE TOTAL TOTALS INDEX AS AN AID TO FORECASTING SEVERE WEATHER ACTIVITY IN NORTHERN ILLINOIS

D. L. Sisterson

Introduction

On 3 August 1978, a small tornado touched down west of ANL during working hours and travelled through the Laboratory site. Although damage was light, the occurrence of a tornado on site serves as a reminder of the susceptibility of northern Illinois to violent weather. This note summarizes the operational value of the Total Totals Index (TTI) as a predictor of severe weather within a 100 km radius of ANL (which includes the Chicago area) during the 1978 tornado season. The results are very similar to the results from a pilot study (Sisterson, 1977).

The use of the TTI is not original to this study, as the index is a well-established method of forecasting thunderstorm activity (Miller et al., 1965). While other thunderstorm prediction schemes are more sophisticated and presumably more accurate, the simplicity of the TTI makes it attractive for use, for it requires no complicated analysis or computations.

Data Acquisition

The TTI is defined as:

$$TTI = (T_{850} - T_{500}) + (Td_{850} - T_{500}) ,$$

where T_{850} represents the temperature at 850 mb, Td_{850} is the 850 mb dew-point temperature, and T_{500} is the 500 mb temperature. The upper-air data used to compute TTI values were obtained from the 0600 Central Standard Time (CST) sounding at Peoria, Illinois, supplied by the National Weather Service office in Chicago. Peoria is the rawinsonde station nearest ANL, lying 150 km to the southwest; it provides the best measure of the airmass characteristics of severe weather in northeastern Illinois, since those air-masses typically move southwest to northeast. The TTI is computed daily by 1030 local time.

The critical value for thunderstorm development during 1978 was empirically found to be near 42 (the critical value in 1977 was 43). Data were insufficient for a critical value for severe weather development to be determined. Therefore, severe weather events were counted as thunderstorm events for statistical analysis of the TTI as a predictor of thunderstorm activity.

An event was defined as either a thunderstorm or a severe weather occurrence within 100 km of ANL, verified by television, radio, National Weather Service radio, or direct observation at ANL. A storm was considered severe if wind gusts over 25 m s^{-1} , half-inch hail, or a tornado or funnel cloud occurred.

During the 101-day study period, southwesterly winds ($180-270^\circ$) at 850 mb occurred on 40 days. The study in 1977 used the 500 mb wind direction to determine airmass movement, but use of the Peoria 850 mb wind direction in 1978 allowed comparison with the 850 mb wind from the Chicago Midway 0600 CST sounding. Occurrence of southwesterly winds at both Peoria and Midway was selected as an indication that the Peoria airmass reached the Chicago area, and that the Peoria sounding was thus representative for this area. Unfortunately, the Midway data were not available until after the study period. In 5 cases, stationary fronts separated Peoria and Chicago airmasses, and these cases were eliminated. However, these cases could not be eliminated a priori and the TTI over-predicted thunderstorm potential for the Chicago area on those days.

Data Analysis

A comparison of the TTI forecasts and observations is shown in Table 1. A chi-square test indicates significance at the 0.1% level, i.e., there is less than a 0.1% probability that the distribution in Table 1 could have occurred by chance. The skill score S , defined by $S = (R-E)/(T-E)$, where R is the number of correct forecasts, T is the total number of forecasts, and E is the corresponding number of correct forecasts that could have occurred by chance, is 0.56.

Table 2 shows the frequency of events observed in southwesterly winds

Table 1. A comparison of the TTI forecasts and observations.

Forecasts	Observations	
TTI	No event	Thunderstorm
< 42	14	2
> 42	7	17
	Total 40	

Table 2. The distribution of 1978 events by wind direction and storm type.

Wind direction	No event	Thunderstorm	Severe	Total days
Southwesterly winds	21	13	6	40
All other wind directions	58	1	2	61
Total	79	14	8	101

compared to the frequency of events observed for all other wind directions. The flow was southwesterly on 40% of the study days, but 86% of all thunderstorms and 75% of all weather occurred when winds were southwesterly. Events occurred on 47% of the days with southwesterly winds, but only on 5% of other days.

Figure 1 contains a plot of the components of the TTI, convective instability ($T_{d850} - T_{500}$) versus static stability ($T_{850} - T_{500}$). The 45° line through the origin represents saturation at 850 mb, with the short lines parallel to it representing 850 mb dew-point depressions. Curves of constant TTI are orthogonal to the dew-point depression lines. Sixty-three percent of the cases contained in empirically-derived boundaries formed by the saturation and TTI = 42 lines represent occurrences of days with thunderstorm activity, sometimes severe. In a few cases, events occurred with low values of TTI,

FIG. 1.--Verification of the total totals index plotted as convective instability versus static instability. Three no-event points with negative abscissas are not plotted. No event, O; thunderstorm, ●; severe, ■.

and in several others no events occurred for values of TTI within these boundaries. There is no immediately obvious explanation or rationalization for these exceptions.

Conclusions

In a second year of testing, the Total Totals Index (TTI) has again provided an early means for assessing potential thunderstorm activity within a 100 km radius of ANL. Thunderstorm activity occurred 63% of the time it was forecast in this study, with severe weather on one thunderstorm day in three. More importantly, a forecast of no thunderstorms with southwesterly winds was correct 88% of the time, a much better result than during 1977. For winds from the other three quadrants, no thunderstorm activity occurred on 95% of the days, regardless of the TTI. Thus, when special events requiring favorable weather are considered, the TTI provides a simple, objective assessment of the conditions to be expected later in the day.

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SOME LIMITATIONS OF DIMENSIONAL ANALYSIS AND POWER LAWS^{*}

B. B. Hicks

An artificial correlation introduced as a direct consequence of constructing dimensionless quantities from sets of unrelated data is investigated. The analysis results in expressions for the induced correlation coefficient and power-law exponent that permit workers to examine the objectivity of results obtained with this method of data analysis.

^{*} Abstract of a paper published in 1978: *Boundary-Layer Meteorol.* 14, 567-569.

X

AN EXAMINATION OF SOME MICROMETEOROLOGICAL METHODS FOR MEASURING DRY DEPOSITION*

B. B. Hicks and M. L. Wesely

Dry deposition on natural surfaces is one of the major removal pathways for air pollutants. In order to develop mathematical descriptions for the numerical simulation of the transport, removal, and ecological impact of pollutant gases and aerosols, the dependence of dry deposition rates on physical, chemical, and biological parameters must be understood. Such relationships can be studied by using several experimental methods to determine the vertical fluxes of pollutants over natural surfaces. The possible experimental methods include aerodynamic, modified Bowen ratio, eddy correlation, variance, and eddy accumulation. The relative advantages and disadvantages of these methods are discussed, with consideration being given to the sensor response time and accuracy. The roles of atmospheric stability and the zero plane, site and instrumental requirements, and averaging time are discussed for flux measurements.

* Abstract of a 1978 Interagency Energy/Environment R & D Program Report, EPA-600/7-78-116.

A PUFF POLLUTANT DISPERSION MODEL WITH WIND SHEAR AND DYNAMIC PLUME RISE*

C. M. Sheih

A puff diffusion model, which includes wind shear and dynamic plume rise, is developed for numerical prediction of pollutant concentrations under steady and non-uniform flow conditions. The plume from a continuous source is treated as a series of puffs emitted successively from the source. Each puff is represented by a set of six tracer particles, which define the size, shape, and location of the puff. Initially these particles are located at the surface of the source, on arbitrarily chosen orthogonal axes. The location of the particles is computed at each time step by taking into account advection, eddy diffusion, wind shear and entrainment of ambient air during plume rise. The concentration distribution of each puff is determined by fitting an ellipsoid to the cluster of six particles and assuming a three-dimensional Gaussian distribution, with standard deviations equal to the half-lengths of the principal axes of the ellipsoid. The concentration at a point of interest is obtained by summing the contribution from nearby puffs. The effect of wind shear on the pollutant is investigated by use of a typical wind shear encountered in the atmosphere. The results show that, at 600 m downstream from the source, the present model gives concentrations a factor of 2 higher and lower at one standard deviation below and above the plume center, respectively, than conventional models in which no wind shear is considered. The plume-rise formulation is calibrated against the observations compiled by Briggs, and the model is used to predict the trajectory of a plume observed by Slawson and Csanady. Excellent agreement between the prediction and the observation can be achieved if an appropriate eddy diffusivity is chosen.

* Abstract of a paper published in 1978: Atmos. Environ. 12, 1933-1938.

X

A THREE-DIMENSIONAL, SECOND-ORDER CLOSURE NUMERICAL MODEL OF
MESOSCALE CIRCULATIONS IN THE LOWER ATMOSPHERE: DESCRIPTION OF
THE BASIC MODEL AND AN APPLICATION TO THE SIMULATION OF THE ENVIRON-
MENTAL EFFECTS OF A LARGE COOLING POND*

T. Yamada

A detailed description of a three-dimensional, mesoscale model of circulations in the lowest few kilometers of the atmosphere, the planetary boundary layer, is presented. The model is termed a turbulence second-moment closure model since it is based on various second-moment turbulence equations involving wind velocity, temperature, water vapor, and liquid water.

A simplified version of the model is recommended, in which only time-dependent, partial-differential equations for the turbulence energy balance and a master length scale are solved. A Gaussian probability function for cloud cover statistics recently proposed by Sommeria and Deardorff and by Mellos is used in order to model condensation processes. Thus not only are the effects of thermal and moisture stratification included, but the effects of condensation are considered as well. As one example of an environmental assessment application, the model is used to simulate formation of clouds over a cooling pond, and the results obtained are qualitatively compared with an observation of this inadvertent modification of local weather.

* Abstract of Argonne National Laboratory Radiological and Environmental Research Division Report ANL/RER-78-1.

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