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R. E. Rowland, Division Director

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This volume, Part IV of the Radiological and Environmental Research Division Annual Report, contains results of investigations conducted by the Atmospheric Physics Section during 1972. Part I, on radiation physics and Part II on human radiobiology, have appeared earlier, while Part III, reporting the results of ecological studies, is being issued concurrently with the present volume.

The reader is reminded that, because of seasonal restrictions imposed upon many environmental research programs, Parts III and IV of this series are produced on a calendar rather than a fiscal year basis.

FOREWORD

During 1972 the Atmospheric Physics Section operated with reduced support and a somewhat smaller professional staff. Research activities were, therefore, primarily concerned with completion of work in hand. Preliminary steps in the form of equipment development were initiated in anticipation of one future program, an experimental investigation of the planetary boundary layer in depth in order to improve our knowledge of those circulations which control pollutant dispersion over mesoscale distances.

Analysis of the extraordinarily large number of double-theodolite pilot-balloon observations obtained over St. Louis during the 1971 METROMEX field program continued. An account of some preliminary results of this investigation is included in these pages, and a catalog of the wind profile data obtained is in preparation for a topical report. The statistical studies of light-wind, low-dispersion phenomena are represented in this report by several papers on the nature and frequency of occurrence of these conditions in the Chicago metropolitan area. Further studies of this nature are to be summarized in an additional topical report.

Four articles report further results obtained in the continuing micro-meteorological studies over Lake Michigan. Additional measurements of the exchange of thermal energy between the lake surface and the overlying atmosphere are presented, while a note on indirect evaporation measurements describes a procedure to be employed in future studies of cooling ponds. A matter of some importance to safety analyses of nuclear power installations offshore is discussed in a note describing the estimation of plume dispersion over open water in terms of turbulence measurements, the basic data used originating from the Lake Michigan experiments. Finally, two short papers describe the development of new, semi-automatic pilot-balloon tracking and radiosonde equipment, anticipating the extension of the Atmospheric Physics Section's boundary layer research to greater elevations.

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SOME MEASUREMENTS OF NOCTURNAL WIND FLOW OVER
ST. LOUIS: METROMEX 1971

Richard E. Albert, Harry Moses, and William R. Robinson*

Introduction

Long term average dispersion in the lower atmosphere over meso-scale distances is largely controlled by the mean wind; small scale turbulent diffusive contributions are of lesser importance. Wind direction obviously determines the general orientation of pollutant transport, while wind speed affects pollutant concentration by spreading downwind materials emitted from ground sources. It follows that an important consideration in the development of mesoscale pollutant dispersion models must be the fact that systematic differences in mean wind distributions exist over urban and rural terrain, the physical and thermal structure of cities themselves being the principal cause. An investigation of the nature of this disturbed wind flow above a large metropolitan area was begun by this Laboratory in the summer of 1971 as a contribution to the Metropolitan Meteorological Experiment (METROMEX).⁽¹⁾

Steady-state wind flow in the planetary boundary layer is governed by a balance of three forces: the pressure gradient, the frictional stress, and a virtual force associated with Coriolis acceleration. Changes in the characteristics of the lower boundary can affect the flow aloft by altering this balance. For example, the relatively greater surface roughness of urban terrain increases the frictional force; winds in lower levels typically respond by decelerating and turning toward lower pressure. In addition, a city frequently acts as an area heat source, notably during evening hours when the central urban region loses heat less rapidly than does the surrounding countryside. Under conditions of clear skies and light winds, large urban-rural temperature contrasts develop, creating what has come to be known as an

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"urban heat island;" low-level wind flow over the city is modified through associated changes in the local pressure field.

The August 1971 Experiment

To measure these effects, a network of pilot-balloon launching stations was operated for an extended period in and near St. Louis, Mo. during the summer of 1971. Figure 1 shows the location of the sites selected to obtain measurements representative of wind flow over the central urban area, the suburbs, and surrounding rural regions. This study uses profiles observed between 2100 and 2400 LST on six nights to investigate the spatial and temporal variability of the low level winds during the hours when urban heat effects are often strong. During the observation periods, balloons were released at 20-min intervals; measurements obtained were interpolated to derive wind velocities at height intervals of 100 m up to 1000 m. Field operations for the program were carried out by personnel from Argonne National Laboratory and from the United States Air Force Air Weather Service,* under the direction of B. Ackerman.**

The observation program produced double-theodolite data in the form of azimuth and elevation angles of successive apparent balloon positions as seen simultaneously from two points separated by baselines between 472 and 610 m. The raw information was recorded in the field verbally on small tape recorders. After being transcribed from the audio tapes, the data were reduced to successive balloon positions in space, and then to successive balloon displacements or wind speeds and directions averaged over the 20-sec interval between observations, using a technique developed by Thyer.⁽²⁾ This method recognizes that as a result of small individual tracking errors, the two optical rays designated by any set of simultaneous theodolite readings usually do not intersect; for this reason, the program calculates a most

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FIG. 1.--Locations of Pibal sites in the METROMEX research area

probable balloon position in terms of the expected resolution error of the angular observations.

The measurements (often in the form of preliminary wind profiles) were screened to eliminate the spurious contributions of observer errors, instrumental defects, and mistakes in site surveys. Examples of these include:

1. Human errors: baseline distance or height difference between theodolite pads measured incorrectly; theodolite not leveled properly;

angles read incorrectly; angles from both instruments not read simultaneously; and verbal audio tape records not transcribed correctly during playback.

2. Instrumental defects: misalignment of short or long range optics; slippage of azimuth or elevation scales; and inaccurate positioning of cross hairs.

3. Orientation errors: incorrect survey of baseline orientation relative to true north; and incorrect orientation of theodolite relative to local direction references.

Synoptic weather conditions prevailing during the evenings of August 12, 13, 16, and 17 were basically similar. High pressure systems centered east and north of the city resulted in relatively weak pressure gradients and generally clear skies. On August 23 and 24 the St. Louis area was under the influence of a low pressure system centered to the north. Associated with this cyclone was a weak frontal system that was nearly stationary during the measurement period of the 23rd, but which began to advance during the early morning hours of the 24th. Note the evidence of a change in advection aloft in the wind direction profile of August 23 (Figure 2f). The other five profiles show that the wind steadily veered with height as would be expected of deep warm air advection.

Wind Profiles

Effects of the city upon flow in the first 1000 m above the ground were evaluated by separately considering pilot-balloon data recorded over rural (sites 1, 2, 3, and 4) and urban (sites 7, 9, 16, 25, and 30; see Figure 1) areas. Arithmetic average speeds and directions determined from the profiles observed on each of the six days over the two types of terrain are plotted in Figure 2, each point on the profile representing an average of between five (one site only) and 37 (five sites) individual launch readings. The complete data set used consists of 319 wind profiles. Characteristic features of the observed wind distributions include the following:

1. On each of the six nights a low level wind speed maximum is present between 200 and 300 m above the rural site and, with one exception,

over the urban sites; the exception, August 17, exhibited the lightest winds of this set of observations.

2. On five of the six nights the wind speed average maxima over urban terrain were displaced upward approximately 100 m relative to simultaneous elevations of the faster moving layer over rural areas. Again the exception was August 17.

3. On four nights the maximum wind speed over the rural sites exceeded the maximum over the urban sites by values ranging from 0.2 to 2.5 mps.

4. Below some elevation between 200 and 700 m, wind speeds over urban areas were generally less than corresponding rural values; above these elevations the reverse is true in five of the six cases.

5. At elevations of 100 m decelerated flow over the city is deflected to the left (toward lower pressure) relative to the undisturbed rural flow. Two factors contributed to this deviation from geostrophic flow: 1) surface friction, and 2) small scale circulations arising from convection, urban-rural heat differences, and other local effects.

The occurrence of a low-level wind speed maximum (not strictly a low-level jet, but rather an extensive quasi-horizontal layer of faster moving air) is related to the formation of inversions near the ground. In the evening, rapid cooling of the air near natural vegetation surfaces creates a rural, ground-based inversion, which restricts vertical mixing and consequently reduces the downward transport of momentum. Released from frictional coupling with the surface, air in the lower layers tends to accelerate, beginning an inertial oscillation which soon results in the wind speed overshooting the equilibrium value.⁽³⁾

Wind Variability as Denoted by Ranges of Speed and Direction

Frequently one uses wind data measured several miles away from the point of interest with the hope that they are sufficiently representative. As an example, in response to citizens' complaints, a state or local air pollution agency, for lack of better measurements, may be forced to determine the

FIG. 2a.--Wind speed profiles for August 12th and August 13th averaged over a three-hour period for the total number of rural and urban sites indicated.

FIG. 2b.--Wind direction profiles for August 12th and August 13th averaged over a three-hour period for the total number of rural and urban sites indicated.

FIG. 2c.--Wind speed profiles for August 16th and August 17th averaged over a three-hour period for the total number of rural and urban sites indicated. (For convenience in plotting, the higher wind speeds of the 16th are shown on the right, reversing the chronological order of the other diagrams.)

FIG. 2d.--Wind direction profiles for August 16th and August 17th averaged over a three-hour period for the total number of rural and urban sites indicated.

FIG. 2e.--Wind speed profiles for August 23rd and August 24th averaged over a three-hour period for the total number of rural and urban sites indicated.

FIG. 2f.--Wind direction profiles for August 23rd and August 24th averaged over a three-hour period for the total number of rural and urban sites indicated.

trajectory of a toxic emission using winds measured 5 to 10 miles from the point of release. In fact, the winds are often so variable that the distance over which particular wind measurement may be considered representative may be limited to less than one mile from the point of measurement. Factors which give rise to local wind variability are:

1. orographic features which modify surface roughness, such as bodies of water, stands of trees, or hill-valley complexes;
2. cloudiness and time of the day which determines the degree of incoming and outgoing radiation and in turn control convective circulations or katabatic activity such as a valley wind; and
3. the size and structure of local and mesoscale pressure patterns which determine the magnitude and direction of gradients in the wind field.

The degree of wind variability over St. Louis and its surrounding rural area up to elevations of 1000 m above ground are examined below. As indicated previously, the measurements are limited to nighttime (2100-2400 LST) conditions during clear skies with their attendant strong radiational cooling.

Although there are many ways to describe variability in the wind field, the observed "spatial range" of simultaneous wind speed and direction values is used here. The spatial range of speed, for example, is the difference between the highest and lowest speeds measured simultaneously over either the urban or rural area; an analogous definition applies to wind direction. To clarify the techniques and assumptions made in the analysis which follows, a sample calculation is presented based on the data of Table 1.

This table shows the wind speed values at 100 m for each launch on August 13, 1971 at both the rural and urban sites. At 2140, the range of wind speeds for the rural sites was 0.7 mps (see column 5). The corresponding range for the urban sites was 3.0 mps (see column 11). Only those range values were calculated for which data were available at all sites used; therefore, no range value for the rural area appears for 2120.

TABLE 1. Wind Speeds and Wind Speed Ranges at 100 m on August 13, 1971

Launch time, (a) CST	Wind speed at rural site number, mps			Range, mps	Wind speed at urban site number, mps					Range, mps
	1	3	4		7	9	16	25	30	
2120	7.2	6.9	9.5	---	4.9	6.4	---	4.6	5.8	---
2140	7.9	7.2	7.7	0.7	4.4	7.1	7.4	5.3	4.4	3.0
2200	7.8	7.9	8.2	---	5.9	5.0	6.4	4.6	---	---
2220	6.7	7.5	7.9	1.2	4.0	6.9	6.6	5.9	4.7	2.9
2300	7.4	6.5	6.8	0.9	3.6	5.7	4.7	3.2	4.7	2.5
2320	5.2	6.3	7.3	---	4.2	---	5.1	3.2	4.2	---
2340	6.4	---	6.9	---	4.0	5.2	7.1	4.1	3.8	---
2400	6.6	7.1	7.3	---	4.9	4.8	---	2.9	4.0	---
Average range				0.9	Average range					2.8

(a) The range value for any launch was not computed if an entry was missing for either rural or urban station.

The average range given in the last row of the table represents the average of the range values of column 5 for the rural data and the average of the range values of column 11 for the urban data. Similar tables of wind speed and direction range were generated for each of the six night periods and for every 100 m above ground to 1000 m. Results are summarized in Figures 3 and 4, as discussed in detail below.

Ranges of Wind Speed

Wind speed range values for four elevations above ground, 100, 300, 500, and 700 m, for the six nighttime periods, are given in Figure 3a. It is evident from this figure and the summary shown in Table 2 that, in general, the urban ranges are larger than the rural ranges. For example, at 100 m the rural mean of the speed ranges is 1.2 mps compared to 2.5 mps for the urban sites. An application of the Student t test shows that the difference between the urban and rural mean ranges is significant to at least the 5% level for elevations up to 400 m.

As shown by Table 2, the mean ranges of wind speed for urban data are lowest at the 300-m level, which approximates the elevation of the maximum speed of the low level jet. The difference between the urban and rural average wind speed ranges appears to be largest at the 100-m level, the measurements nearest the surfaces in these observations.

Average wind speed ranges as a function of height for the individual days of August 12, 13, and 17 are presented in Figure 4; these days were selected because the data were most nearly complete on these dates. Range bars show the minimum range as the left extremity and the maximum as the right extremity. When no range bars are given, only one set of data was available for that height.

Figure 4 shows that on August 12, 13, and 17, the rural mean wind speed range varied little with height. To illustrate, on August 17 all of the points fell within a 0.7-mps band. Similarly, the speed ranges for August 12 and 13 fell within 0.9- and 1.1-mps bands, respectively. On the other hand, the magnitudes of the urban mean wind speed ranges exhibited appreciably

FIG. 3a.--Range of wind speed among rural and urban sites for individual nighttime launches

FIG. 3b.--Range of wind direction among rural and urban sites for individual nighttime launches

TABLE 2. Variation with Height of the Mean Wind Speed and Direction Ranges and Their Standard deviations^(a)

Mean wind speed ranges					
Height, m	Rural mean, mps	Urban mean, mps	Rural standard deviation, mps	Urban standard deviation, mps	Student t value
100	1.2	2.5	0.4	0.3	-6.0
200	1.0	2.4	0.6	0.4	-4.5
300	0.9	1.8	0.2	0.8	-2.4
400	0.9	1.8	0.3	0.7	-2.5
500	1.2	2.2	1.1	1.1	-1.4
600	1.4	1.9	1.3	0.7	-0.8
700	1.6	2.4	1.9	1.3	-0.8
800	1.6	1.7	1.9	0.7	-0.1
900	1.6	1.6	1.9	1.0	-0.0
1000	1.1	1.6	0.9	0.8	-0.9

Mean wind direction ranges					
Height, m	Rural mean, deg	Urban mean, deg	Rural standard deviation, deg	Urban standard deviation, deg	Student t value
100	8.3	43.8	6.9	36.3	-2.2
200	7.0	41.5	4.9	33.1	-2.3
300	12.3	33.5	15.2	17.1	-2.1
400	24.5	30.0	17.3	11.8	-3.8
500	9.2	30.2	5.2	12.7	-3.4
600	11.8	25.8	5.5	6.2	-3.8
700	13.8	24.7	11.5	7.9	-1.7
800	10.3	23.0	4.9	5.4	-3.9
900	18.0	26.7	10.4	9.1	-1.4
1000	23.7	31.7	16.0	7.7	-1.0

(a) This table is based on the measurements taken during the nights of August 12, 13, 16, 17, 23, and 24, 1971.

greater variation, falling within about a 2-mps band. The urban wind field differed from that of the rural in two additional respects: 1) the speed range bars (figure 4) for the urban data were longer, and 2) the urban data for all three days suggest a minimum speed range at the approximate height of the low level jet maximum, but the rural data do not.

Ranges of Wind Direction

The mean urban and rural direction ranges and the standard deviations for every 100 m to 1000 m are given in Table 2. Mean rural direction ranges extend from only 7 to 24 degrees, while mean urban direction ranges extend from 23 to 44 degrees. The actual distributions of the urban and rural wind direction ranges upon which the data of Table 2 are based are shown in Figure 3b for 100, 300, 500, and 700 m. A Student t test indicates the urban-rural circulation to be significant to better than the 10% level to 600 m. No minimum in wind direction range appears at the level of the jet maximum, but Table 2 shows that the mean wind direction range decreases up to 800 m in the urban but not in the rural areas.

Figure 4 shows the mean direction ranges as a function of height for the rural and urban data. Mean direction ranges appear to vary but little with height with the exception of the urban values for August 17. On this day the ranges in lower layers were quite high, attaining about 120° at 100 m; they fell to approximately 20° at 900 m, appearing to increase above that level. When the pressure gradients are weak local air flows arising from convection or urban-rural calculations predominate, and large direction differences result, especially when the scale size of these patterns is smaller than the distance between sites.

Summary and Comments

A low level jet, generally at 200 to 300 m, appeared during all six nighttime periods. On five of the six nights, the maximum speeds over the city occurred at levels of about 100 m higher than those in the country; on the sixth night, August 17, the series with the lightest winds, the rural

FIG. 4.--a) Maximum, minimum and average ranges in wind speed and wind direction for August 12th over a three hour period for the total number of rural and urban sites indicated; b) Maximum, minimum and average ranges in wind speed and wind direction for August 13th over a three hour period for the total number of rural and urban sites indicated.

FIG. 4.--c) Maximum, minimum and average ranges in wind speed and wind direction for August 17th over a three hour period for the total number of rural and urban sites indicated.

maximum occurred at 300 m, but no real urban peak occurred.

Invariably the wind speeds at 100 m in the rural area exceeded those over the city, but a crossover in the speed profile occurred at a level between 200 and 700 m. Associated with the lower urban wind speed at 100 m above ground, the average wind direction was directed toward lower pressure, i.e., counterclockwise with respect to the rural wind direction, responding to the deceleration effects of surface friction which upset the balance of forces and permit the pressure gradient to deflect the flow. Additional effects on the wind direction are exerted by urban-rural circulation patterns superimposed upon air flows arising from the synoptic highs and lows.

The wind variability as measured by the ranges of the speed and direction was significantly greater over the city than over the rural areas. In fact, insofar as wind variability is concerned, the influence of the city extended up to at least 1000 m — the highest level examined, although the urban effect was diminished above about 600 m. Values of direction range over the city were generally high, averaging about 30° at all levels compared with about 10° for the rural area. However, urban direction ranges decreased with height to about 20° at 800 m (increasing again above that level), while rural direction ranges tended to remain low (10 to 20°) at all levels.

Wind variability appears to vary inversely with wind speed. For example, on the day with the lightest winds, direction ranges were greater than the average value for all days (see Table 2); further, the wind speed ranges showed a minimum at the level of the low level jet maximum.

The assistance of Paul E. Hess, Frank C. Kulhanek, and Gunther A. Zerbe is gratefully acknowledged. Valuable counsel in analyzing the data was provided by Dr. Bernice Ackerman.

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COMPARISON OF THE SURFACE AND GEOSTROPHIC WINDS

Harry Moses and Paul E. Hess

Introduction

Wind speed and direction are the most important meteorological variables controlling concentration levels of any atmospheric pollutant. Although air pollution models generally consider other variables as well, one relatively satisfactory model uses only wind speed.⁽¹⁾ For application in cities or residential areas, any model requires information on the behavior of the in situ wind field, knowledge of which is often lacking and prediction of which is undoubtedly difficult. To predict occasions which may result in high levels of pollution, it is tempting to employ geostrophic winds, scaled in some manner appropriate to the particular situation, since these can be obtained from conventional synoptic charts and prognoses. The present purpose is to study the relationship between geostrophic and surface winds and to emphasize the limitations of this approach.

The wind is determined by the ambient pressure field, which may arise both from the large scale pressure patterns and the local pressure differences which produce local convective or katabatic flow. To a fairly good approximation, the wind speed, U , in a given area without appreciable convective force is given by

$$1/\rho \frac{\Delta P}{\Delta n} = f U + \frac{U^2}{R} + F \quad (\text{flow around a low pressure area}) \quad (1a)$$

and

$$1/\rho \frac{\Delta P}{\Delta n} = f U - \frac{U^2}{R} + F \quad (\text{flow around a high pressure area}) \quad (1b)$$

where ρ is air density, $\Delta P/\Delta n$ is the pressure gradient, f is the Coriolis parameter ($9.717 \times 10^{-5} \text{ sec}^{-1}$ for the Chicago Midway Airport), R is the radius of curvature of the isobars, and F is the retarding acceleration due to frictional forces. Any consistent system of units may be used,

Since our main interest is in light winds, conditions under which R

is generally large, the centrifugal term U^2/R may be neglected. For $U = 5 \text{ m sec}^{-1}$ and $R = 1000 \text{ km}$, the centrifugal term is an order of magnitude smaller than the Coriolis term. In rare cases with small radii of curvature such as 100 km and winds of 10 m sec^{-1} or greater, the two terms may be comparable. When dealing with averages which do not distinguish between cyclonic and anticyclonic flow, errors resulting from neglect of the centrifugal term are minimized owing to the opposite signs of the centrifugal term, U^2/R , in cyclonic flow and anticyclonic flow [see Eqs. (1a) and (1b)]. The neglect of the centrifugal term is not considered a serious shortcoming for this study.

The Determination of the Geostrophic Wind

The technique for calculating the geostrophic wind is given below. Although the procedures are conventional, it is desirable that they be given so that users of this method or those who apply the results of this study be informed of the assumptions made and of the differences from more precise methods, such as taking into account curvature effects of the isobaric surfaces.

The geostrophic wind equation is given as

$$V_g = \frac{1}{\rho f} \frac{\Delta P}{\Delta n} \quad (2)$$

The value for f is given above, and the value for density, ρ , can be determined from the gas law equation,

$$P = \rho \cdot \frac{R}{m} \cdot T \quad (3)$$

in which P = pressure in dynes/cm² at the Chicago Midway Airport,

$\frac{R}{m}$ = gas constant $\approx 2.8704 \times 10^6 \text{ ergs/g/}^\circ\text{K}$ for air, and

T = temperature in $^\circ\text{K}$.

The pressure gradient is determined as follows: Pressure values for every 3-hour period of the day for the 5-year period, January 1, 1966 to December 31, 1970, were obtained from the National Weather Service Stations in Madison, Wisconsin; Peoria, Illinois; Muskegon, Michigan; South Bend

and Indianapolis, Indiana. These stations surround the Chicago area.

The latitudes, longitudes, and elevations of these stations are given in Table 1. Pressures at each station are adjusted to that on a horizontal datum plane having an elevation equal to that of Chicago Midway Airport, the station with the lowest elevation. The relation to accomplish this adjustment is given by

$$P_i = \rho_i g (Z_r - Z_i) + P_r, \quad (4)$$

where P_i = pressure in dynes/cm² at given station,
 P_r = pressure in dynes/cm² at height of reference station, Chicago Midway Airport,
 ρ_i = density in g/cm³ at given station,
 Z_r = height above sea level of reference station = 18,994 cm,
 Z_i = height above sea level of given station in cm, and
 g = acceleration of gravity = 980 cm/sec².

ρ_i is given as

$$\rho_i = \frac{P_i}{\frac{R}{m} \left[T_i - \frac{6 \times 10^{-5} (Z_r - Z_i)}{2} \right]}, \quad (5)$$

where subscript i refers to a given station. The number 6×10^{-5} represents the mean lapse rate of 6°C/km or 6×10^{-5} C/cm. Since an average density is desired, the last term in the bracket of Eq. (5) is divided by 2. Values for ρ and f are now known, only $\Delta P/\Delta n$ needs to be determined.

To determine the pressure gradient $\Delta P/\Delta n$, a conventional multiple regression technique is used to specify a plane of best fit expressed as

$$P = C + AX + BY, \quad (6)$$

where A, B, and C are constants. From elementary calculus

$$\frac{\partial P}{\partial n} = \left[\left(\frac{\partial P}{\partial X} \right)^2 + \left(\frac{\partial P}{\partial Y} \right)^2 \right]^{1/2}, \quad (7)$$

which, together with the relationships $\partial P/\partial X = A$ and $\partial P/\partial Y = B$ from

TABLE 1. Latitude, Longitude, and Elevation of Stations used for Determining the Regression Plane

	Latitude	Longitude	Elevation above sea level, cm $\times 10^4$
Indianapolis, Ind.	39° 44'	86° 17'	2.4634
Madison, Wis.	43° 08'	89° 20'	2.6402
Muskegon, Mich.	43° 10'	86° 14'	1.9299
Peoria, Ill.	40° 40'	89° 41'	2.0183
South Bend, Ind.	41° 42'	86° 19'	2.3567
Chicago, Ill. (a) (Midway Airprt)	41° 47'	87° 45'	1.8994

(a) The Chicago Midway pressures are not used in determining the regression plane.

Eq. (6), leads to a tractable form for the geostrophic wind, which may be computed from

$$V_g = \frac{1}{\rho f} \frac{\partial P}{\partial n} = \frac{1}{\rho f} \sqrt{\left(\frac{\partial P}{\partial X}\right)^2 + \left(\frac{\partial P}{\partial Y}\right)^2} = \frac{1}{\rho f} \sqrt{A^2 + B^2}. \quad (8)$$

The geostrophic wind V_g is parallel to the isobars with highest pressure to the right. Assuming a polar coordinate system with 0° as north, 90° as east, 180° as south, and 270° as west, the direction of the geostrophic wind is determined as follows.

First the u and v components of the geostrophic wind were calculated. The direction of u is east-west, with positive in the east direction; and v is in north-south, with positive in the north direction.

$$u = -\frac{1}{\rho f} \frac{\partial P}{\partial Y} \quad (9)$$

and

$$v = \frac{1}{\rho f} \frac{\partial P}{\partial X} \quad (10)$$

The direction of the geostrophic wind (that from which it is blowing), θ , can be determined from $\phi = \arctan u/v$:

$$\theta = \phi \text{ for } u < 0$$

and

$$\theta = \phi + 180^\circ \text{ for } u > 0 ,$$

where ϕ is confined to the range 0 to 180° . When $u = 0$, $\theta = 0^\circ$ if $v < 0$ and $\theta = 180^\circ$ if $v > 0$; when $v = 0$, $\theta = 90$ when $u < 0$ and $\theta = 270$ when $u > 0$.

Comparison of Geostrophic and Surface Wind Speeds

In comparing the near-surface wind speed, U , with the geostrophic wind speed, V_g , the ratio U/V_g , of these variables, was computed for each third hour of five years of data (1966 to 1970) from four different locations: 1) the Argonne Meteorology Site, 2) the Chicago Midway Airport, 3) TAM Station 1, and 4) TAM station 6. The latter two are part of the Telemetering Automatic Monitoring (TAM) Network of the Chicago Department of Environmental Control. Station locations are shown in Figure 1 and anemometer elevations in Table 2.

The ratio, U/V_g , is a function of several variables beside U and V_g itself. Of special concern are wind direction, θ (which is related to surface roughness and terrain features), and hour of the day, h (which may serve as an indicator of atmospheric stability). We consider here

$$U/V_g = F(U, V_g, \theta, h, X) \quad (11)$$

where X represents variables, not treated in this paper, which exert sufficient influence to provide "noise" in the data.

One may study the effect of U/V_g of the independent variables in $F(U, V_g, \theta, h, X)$ either singly or in any combination. Multivariate contingency tables [tabulation prediction technique⁽²⁾] were constructed using a selection of combinations of independent variables. These tables, which provide probability distributions of U/V_g values and other statistics, such as the mean and standard deviation, were used as the bases for the charts presented here. Only those combinations of independent variables were selected for study which bring into sharper focus the processes which govern the U/V_g ratio.

FIG. 1.--Map showing location of TAM Stations 1 and 6, Argonne Meteorology Station, and Chicago Midway Airport.

TABLE 2. Anemometer Elevations

Station	Anemometer height, ft
Argonne (Lower)	19
Argonne (Upper)	150
Midway	48
TAM 1 - Wm. H. Taft High School	49
TAM 6 - Christian Fenger High School	54

The relation between mean values of U/V_g and V_g itself is given in Figure 2, where it is seen that the U/V_g values are largest at Midway and lowest at Argonne. The curves resemble hyperbolae. If they really were hyperbolae of the form

$$(U/V_g)(V_g) = k,$$

then surface and geostrophic wind would be uncorrelated. A detailed examination of the data shows that this is not the case.

FIG. 2.--Mean values of surface-geostrophic wind speed ratio vs. geostrophic wind speed.

From Figure 2, when V_g is less than 5 mph, the ratio U/V_g is generally much greater than unity ranging from 1.02 at Argonne to 1.73 at Midway. As the wind speed is increased to 20 mph or greater, an asymptotic value of 0.3 to 0.4 appears to be reached at all but the Midway station. The supergeostrophic values for the lighter geostrophic winds may be explained by either katabatic or convective forces exceeding those of the pressure gradients. The convective and katabatic circulations may arise from either land-water temperature differences or differential heating over the land surface itself. As the geostrophic wind increases, the U/V_g ratio decreases, although quite slowly. Evidently the downward transport of momentum, which would tend to increase U , is not enough to prevent a decrease in U/V_g . The surface drag, which varies as U^2 , dominates the decrease in U/V_g . As the geostrophic wind increases beyond 25 mph the U/V_g ratio changes very slowly.

The effect of atmospheric stability is shown in Figure 3 in which the U/V_g ratio is plotted as a function of V_g for the stable period at midnight and for the unstable period around noon. The ratios, U/V_g , are substantially lower during the dark hours when the atmosphere is stable than during the daytime when appreciable convection is expected. This is true not only for light winds but for the equilibrium U/V_g values associated with higher winds. Equilibrium values for U/V_g range from 0.3 to 0.4 at night for Argonne, TAM 1, and TAM 6, but do not appear for Midway. During the day the values are 0.3 to 0.5 for Argonne, TAM 1, and TAM 6; an equilibrium value does not appear for Midway over the speed range considered. The U/V_g values for Midway run from 0.4 to 0.8. The stability effect is quite marked at all of the stations.

The effect of surface roughness and terrain features is shown in Figure 4 when the V_g values are held constant but the surface wind direction is varied. A trajectory over Lake Michigan with its low surface roughness would produce higher wind speed values, U , for a given geostrophic wind speed, V_g . A maximum in the mean U/V_g ratio is to be expected with easterly and northeasterly winds, and indeed it appears. The minor peaks observed probably arise from differences in the land surface roughness and will differ from site to site. A minimum U/V_g value occurs with southeasterly winds at Argonne, TAM 1, and TAM 6 stations, but is quite weak at Midway. These effects are evident for all of the geostrophic wind speeds but are most pronounced with light winds.

The effect of any of the three independent variables (V_g , stability, or wind direction) when the other two are held constant, is shown in Figure 5 for the Chicago Midway Airport. These charts, giving values of U/V_g for all combinations of hour of the day and wind direction to eight points, were drawn for the geostrophic wind speed intervals of 0 to 5, 6 to 10, and 16 to 26 mph. With weak pressure gradients the effectiveness of external pressure forces is marked. Note the pronounced maximum in the afternoon with winds from the northeasterly quadrant. Higher U/V_g values are observed in the

FIG. 3.--Surface-geostrophic wind speed ratio vs. geostrophic wind speed.

FIG. 4.--Surface-geostrophic wind speed ratio vs. wind direction for the indicated wind speeds.

a) Argonne National Laboratory 19-ft level; b) Argonne 150-ft level; c) Chicago Midway Airport; d) TAM 1; e) TAM 6.

4b

4c

4d

4e

FIG. 5a.--Values of U/V_g for all combinations of hour and wind direction to eight points at Chicago Midway Airport. Wind speed interval 0-5 mph.

FIG. 5b.--Values of U/V_g for all combinations of hour and wind direction to eight points at Chicago Midway Airport. Wind speed interval 6-10 mph.

FIG. 5c.--Values of U/V_g for all combinations of hour and wind direction to eight points at Chicago Midway Airport. Wind speed interval 16-25 mph.

afternoon with winds from any direction. The effectiveness of wind direction and probably (but not necessarily) surface roughness is substantially diminished with wind speeds exceeding 6 mph since Figures 5b and 5c show a column maximum instead of the "dome" of Figure 5a. According to these charts supergeostrophic winds prevail in the afternoon for wind speeds in the 6- to 10-mph category but are subgeostrophic in the morning hours. The direction appears to have little influence. For winds in the 0- to 5-mph category all of the values of U/V_g exceed unity, and the maximum exceeds 4. Wind speeds are subgeostrophic in Figure 5c, with values ranging from 0.40 to 0.74. These figures provide a useful tool for wind speed forecasts.

Comparison of Geostrophic and Surface Wind Directions

Winds blow parallel to the isobars when the pressure gradient force just balances the Coriolis force as illustrated in Figure 6a. When surface friction stresses are present (see Figure 6b) the wind blows toward lower pressure. The surface friction stress vector is opposite to the velocity vector.

The deviation of the actual wind from that of the geostrophic wind (parallel to the isobars) is examined in Figure 7 and in Table 3. This deviation is caused by surface frictional stresses as well as other factors such as local stresses arising from surface temperature differences, or a lag in the wind flow adjustment to the pressure field or frictional stress. This is especially true with rapidly changing pressures or wind directions.

Figure 7 shows the relation between the geostrophic wind direction deviation $\Delta\theta_g$ and the geostrophic wind speed V_g . Separate curves are drawn for selected percentiles and for the mean. The mean values are about 30 to 40° at the four stations for speeds exceeding about 20 mph; with lower wind speeds the mean value appears to be smaller. This is misleading since at low speeds the variability is very large, as is evident from the divergence of the curves in Figure 7. In fact, for a larger fraction of the light wind observations, the deviations from the geostrophic direction exceed those for

FIG. 6.--Balance of forces to produce wind; a) without surface friction, b) with surface friction.

the strong wind conditions. This is an example of the limitations of the mean as a statistic.

With light speeds it is possible for winds to blow in a direction opposite to that of geostrophic balance. The greatest variability (dispersion about the $\overline{\Delta\theta_g}$) occurs with the lightest winds and the least with strong winds.

Table 3 presents values of $\overline{\Delta\theta_g}$, the mean deviation of the observed wind direction from the geostrophic direction and the standard deviation of $\Delta\theta_g$, $\sigma_{\Delta\theta_g}$, as functions of wind direction and time of the day. Separate sections are presented for each geostrophic wind speed grouping for each of the four stations. From Table 3 one may observe how the values of $\overline{\Delta\theta_g}$ or $\sigma_{\Delta\theta_g}$ vary with any one of the independent variables when the others are held constant. The deviation from geostrophic flow appears to vary directly with the geostrophic wind speeds, i.e., high wind speeds are associated with greater deviations. On the other hand, for the 26 to 35 mph category, values of $\sigma_{\Delta\theta_g}$ decrease to approximately 20 to 40° with few exceptions. The

TABLE 3. Geostrophic Wind Direction Mean Difference $\overline{\Delta\theta}_g$ and Standard Deviation σ_{θ_g} in Tens of Degrees between January 1, 1966 and December 31, 1970

Hours CST	$V_g = 0-5$ mph								$V_g = 6-10$ mph								$V_g = 26-35$ mph							
	N		E		S		W		N		E		S		W		N		E		S		W	
	$\overline{\Delta\theta}_g$	σ_{θ_g}	$\overline{\Delta\theta}_g$	σ_{θ_g}	$\overline{\Delta\theta}_g$	σ_{θ_g}	$\overline{\Delta\theta}_g$	σ_{θ_g}	$\overline{\Delta\theta}_g$	σ_{θ_g}	$\overline{\Delta\theta}_g$	σ_{θ_g}	$\overline{\Delta\theta}_g$	σ_{θ_g}	$\overline{\Delta\theta}_g$	σ_{θ_g}	$\overline{\Delta\theta}_g$	σ_{θ_g}	$\overline{\Delta\theta}_g$	σ_{θ_g}	$\overline{\Delta\theta}_g$	σ_{θ_g}	$\overline{\Delta\theta}_g$	σ_{θ_g}
Chicago Midway Airport																								
00-02	-1	11	-1	7	4	9	-3	8	4	6	6	8	4	5	3	9	4	2	5	3	4	2	5	2
03-05	7	7	4	5	3	8	2	11	5	5	6	6	3	5	3	8	3	2	4	2	4	2	4	2
06-08	0	9	4	2	1	4	-2	12	6	5	6	4	4	6	3	8	4	2	4	2	4	2	5	1
09-11	2	6	7	8	2	8	4	6	2	6	2	5	3	4	4	6	3	2	3	3	4	2	4	1
12-14	3	9	1	9	0	8	3	1	-1	6	0	6	3	5	3	7	2	3	4	2	3	2	4	1
15-17	-17	1	5	8	3	5	4	8	1	5	2	7	3	3	4	5	3	2	5	5	3	2	4	2
18-20	1	9	0	10	-1	9	3	8	1	6	4	8	4	5	4	4	4	2	5	3	4	2	4	2
21-23	0	11	3	6	3	11	2	6	3	7	5	7	5	5	1	4	4	3	5	3	4	2	4	1
Argonne National Laboratory 19 ft																								
00-02	3	7	5	7	2	6	3	4	5	5	4	6	3	5	4	4	2	2	5	4	3	1	3	2
03-05	2	7	5	6	-4	8	0	9	4	5	4	6	2	3	1	6	2	2	4	3	3	2	3	2
06-08	3	6	3	7	-3	7	0	7	2	6	6	5	3	5	1	5	2	2	4	2	3	2	4	2
09-11	-7	6	2	9	2	7	-2	7	-3	5	0	6	2	6	0	4	1	2	2	2	2	2	2	2
12-14	0	6	-2	7	-4	7	1	10	0	6	1	4	1	6	2	5	2	3	3	3	2	1	2	2
15-17	0	10	4	9	-1	6	2	8	2	6	2	4	0	6	1	6	1	3	5	3	2	2	3	2
18-20	1	12	0	9	5	7	2	9	-2	8	3	7	3	4	0	4	1	4	6	2	3	1	3	2
21-23	1	8	2	9	6	11	0	5	1	7	4	7	4	7	2	3	2	2	5	3	3	1	3	2
TAM Station 1																								
00-02	4	8	3	10	0	8	3	11	5	6	4	7	3	4	3	6	3	3	4	3	3	2	4	2
03-05	3	11	8	6	0	4	1	9	5	5	3	7	0	5	0	8	2	3	3	3	2	1	4	2
06-08	3	9	-5	16	-3	9	-1	9	2	8	6	4	1	4	3	6	3	2	4	4	2	1	4	1
09-11	3	8	1	9	3	7	0	5	0	6	0	6	1	5	2	6	2	3	3	3	2	1	4	2
12-14	-2	4	0	8	-1	10	3	9	2	7	2	8	3	5	1	4	2	2	3	3	2	2	3	2
15-17	2	6	3	9	1	0	4	7	1	5	4	8	2	3	2	4	3	2	3	4	2	2	4	2
18-20	2	11	0	10	4	5	1	7	0	6	5	8	3	7	2	4	1	5	6	2	3	2	3	2
21-23	-3	6	2	8	5	8	2	5	3	5	5	7	3	7	1	4	3	4	5	2	2	2	4	2
TAM Station 6																								
00-02	2	9	3	11	0	9	3	5	2	6	6	6	3	5	3	6	2	3	4	4	3	2	3	3
03-05	-1	13	5	3	0	7	1	11	3	7	5	4	1	4	4	5	3	3	3	2	4	2	3	3
06-08	7	6	0	10	-1	7	4	9	8	3	5	4	2	7	1	5	3	2	3	2	3	2	4	3
09-11	-1	8	-1	12	-2	8	-1	8	0	6	1	7	1	5	2	6	3	2	3	3	3	2	2	4
12-14	4	9	-2	8	-1	6	3	8	0	5	1	7	3	2	0	5	2	3	4	3	2	2	2	3
15-17	-2	5	2	10	4	2	2	7	4	4	2	8	2	5	0	6	2	2	4	4	2	2	2	3
18-20	13	4	0	12	6	5	0	8	6	5	2	9	4	3	1	6	1	5	3	5	3	2	3	3
21-23	-1	6	1	8	1	12	1	10	4	4	5	6	5	3	1	5	2	1	4	2	3	2	2	2

FIG. 7.--Difference in direction between geostrophic and surface wind vs. geostrophic wind speed for indicated percentiles and the mean. a) Chicago Midway Airport; b) Argonne; c) TAM 1; d) TAM 6.

value of the mean $\Delta\theta_g$ was generally 20 to 40° for speeds above those in the 6- to 10-mph interval. As pointed out earlier, the low $\Delta\theta_g$ readings for the 0- to 5-mph grouping are misleading. There appears to be a minimum in $\Delta\theta_g$ values in the afternoon which is associated with maxima in U/V_g . On the other hand, there is evidence for an increase in the $\Delta\theta_g$ values with easterly winds, yet easterly winds are associated with U/V_g maxima.

Summary and Conclusions

The results of this study bring to light a number of interesting facts. These may be summarized as follows:

1. With light geostrophic winds, ≤ 5 mph, relatively large mean U/V_g ratios are observed, i.e., values appreciably greater than unity. Mean values exceeding 4 have been observed.
2. As the geostrophic wind increases, the U/V_g approaches a nearly constant value of 0.3 to 0.4. There are appreciable differences among stations with Argonne Station providing the lowest values and Midway the largest.
3. U/V_g values are substantially greater during the day, with the ratio reaching a maximum in the afternoon, even with winds exceeding 35 mph.
4. The variation of U/V_g is direction-dependent, with substantially larger values occurring with northeasterly and easterly winds, i.e., trajectories over Lake Michigan. Lake breeze winds show larger U/V_g values than winds from other directions having the same V_g values.
5. There appear to be two distinct regimes in the wind flow, corresponding to light wind conditions ≤ 10 mph and to winds exceeding 20 mph, in which the distributions of both U/V_g and $\Delta\theta_g$ are quite different.
6. The mean deviation of the actual wind direction from the geostrophic direction $\Delta\theta_g$ ranges from about 30° at Argonne to 40° at Midway. This value appears to remain nearly constant above 18 mph.
7. The standard deviation of the $\Delta\theta_g$ value is large for light winds, exceeding 100°, but falls off with increasing wind speeds to values of

20 to 40°.

The results presented in this paper, especially in the charts of Figure 5 and Table 3, provide a basis for applying correction factors to the geostrophic wind to yield an improved value of the surface wind. These corrections take into account the influence of stability and surface roughness when time of day and wind direction are used as independent variables.

A number of additional refinements are possible. For example, although the hour of the day and stability are related, the correlation is far from perfect. An improvement would result from considering the actual stability. Further, each station should be calibrated to determine how to apply the roughness correction. As seen from the study, there are urban-rural differences shown by the data from Argonne, a rural site, and the three city sites. In fact, there were appreciable differences among the city sites which need to be taken into account.

Of further importance is the effect of a lag between the indicated pressure patterns and the observed wind. Information on rates at which the wind field adjusts to a changing pressure field should be applied to improve the determination of the surface wind speed and direction from the geostrophic wind.

The assistance provided by Frank C. Kulhanek and Gunther A. Zerbe is hereby acknowledged.

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OCCURRENCE OF LIGHT WINDS AT CHICAGO MIDWAY AIRPORT

Harry Moses and Paul E. Hess

Of the several meteorological variables affecting local pollutant concentrations in the lower layers, wind speed is the most important. During light wind conditions (surface wind speeds of 5 mph or less), levels of pollutant concentration over an urban area are likely to increase since the rate of ventilation of the dilution volume over the city is low. It is therefore useful to examine the frequency with which such potentially hazardous conditions can be expected to occur on both hourly and seasonal bases. This paper describes one such study for a site on the edge of a major metropolitan area, namely, Chicago's Midway Airport, located 10 miles southwest of the city center.

Wind speed near the surface is a function of atmospheric stability, as well as horizontal pressure gradient. During unstable conditions, as might prevail, for example, in a comparatively cool air mass on a sunny afternoon, active convective mixing transports momentum downward from normally faster moving air layers aloft, a process which is able to maintain relatively higher surface-wind speeds against friction losses. For this reason, light wind conditions can be expected to occur less frequently in the afternoon than at other times of the day. On the other hand, radiation cooling of the surface layers during dark hours will often generate ground-based inversions which suppress vertical mixing; in these circumstances air layers near the ground become decoupled from the source of momentum aloft, and light surface winds result.

Seasonal variations in the occurrence of light wind conditions are controlled in part by variations in the strength of the upper-level momentum source. In winter and spring, intense low pressure systems frequently move through the midwestern states, giving rise to periods of higher wind speeds near Chicago. In summer the pressure systems are generally weaker; as a result, light wind conditions occur with greater frequency.

Observed relative occurrence frequencies of surface wind speeds less

than several specified values are plotted in Figure 1. Data for this graph and those following have been derived from 5 years of hourly observations taken at Chicago Midway Airport over the period 1966 to 1970. As shown in the figure, light wind conditions (less than 5 mph) characterize 10% of the observations, while speeds less than 10 mph were observed nearly half the time. The marked change in slope between 4 and 5 mph suggests the onset of a shift in the balance of influences dominating the wind circulation. (See the paper "Comparison of Surface and Geostrophic Winds," elsewhere in this report.)

Figure 2 shows the effects of diurnal stability variation on light wind frequency. As expected, light winds occur least often during the afternoon (1200-1800) when convective mixing is most effective; similarly, the expected maximum frequency of light wind conditions is found during the dark hours (between 2100 and 0600). The peak values appearing in the curves near 0500 are significant; these indicate a maximum probability for light wind conditions at, or slightly before, sunrise. At this time of day, light wind conditions occur 20% of the time throughout the year, compared to 10% for the overall daily average and to less than 3% for the late afternoon.

Seasonal variations in these conditions are shown in Figure 3. During midsummer and early fall (July-September) light winds occur more than 15% of the time, just about twice the 7% frequency recorded during the winter months (December-March). The graph also shows that during July and August 65% of the hourly wind speed observations were less than 10 mph, somewhat more than the annual relative frequency of 50%. Two opposing effects influence summer surface wind conditions; whereas the seasonally increased convective mixing acts to produce higher surface winds, the presence of weaker pressure systems supplies less momentum to be mixed downward. The net result is shown in Figure 3, where the maximum frequency of light winds is seen to occur in the summer months in spite of the increased mixing. Pronounced nocturnal inversions in summer probably also contribute to the increased frequency of light winds at this time of year.

FIG. 1.--Percent of observations showing wind speeds less than or equal to indicated amount at Chicago Midway Airport from January 1, 1966-December 31, 1970.

FIG. 2.--Percent of observations during each hour of the day with wind speeds of indicated amount at Chicago Midway Airport from January 1, 1966-December 31, 1970.

FIG. 3.--Percent of observations during each month of the year with wind speeds of indicated amount at Chicago Midway Airport from January 1, 1966-December 31, 1970.

FIG. 4.--Percent frequency of wind speeds during each hour of the month less than or equal to 3, 5, and 10 mph at Chicago Midway Airport, January 1, 1966-December 31, 1970. a) January; b) April; c) July; d) October.

Additional insight is provided by Figure 4, where the Midway observations are stratified by both time of day and time of year. As expected, light wind conditions are particularly frequent during the early morning hours of summer, totaling 37% of the observations at 0500 in July as compared to only 11% on those recorded at the same hour in January. Diurnal variations will be seen to be much stronger in July than in the other months, although this feature is somewhat concealed by the logarithmic scale of the plots.

This brief study, part of a larger series of investigations of light wind statistics being prepared for a topical report, points out the frequency of occurrence of light winds and how this frequency varies with time of day and season. It also brings to light the errors which may arise when one deals with a variable which is a function of several independent variables and all of these are not considered. This may be true especially when one uses annual wind roses for evaluating the air quality over a region without properly considering the effects of all of the relevant variables.

The assistance provided by Frank C. Kulhanek and Gunther A. Zerbe in preparing this report is hereby acknowledged.

VARIATIONS OF MEAN SURFACE WIND DIRECTION AT CHICAGO MIDWAY AIRPORT

Harry Moses and Paul E. Hess

When mean ground-level pollutant concentrations are calculated for regions downwind of a point source for a specified period, e.g., several hours, due regard is paid to the important effects of mean wind speed; but the mean wind direction is often tacitly assumed to remain constant. In reality, because of the movement and development of both large scale and local pressure systems, the direction of the surface wind changes significantly. Obviously, even small shifts in mean wind direction can result in significant changes of pollutant concentrations observed downstream. To document this important effect, this paper examines the degree of variation in mean surface wind direction that can be expected in periods of 2, 3, 6, 9, 12, 18, and 24 hours. The study has been carried out for a site on the edge of a major metropolitan area, namely Chicago's Midway Airport, located 10 miles southwest of the city center. The basic data used are taken from hourly observations recorded for the 5 years between 1966 and 1970 by the U.S. National Weather Service. In these records, wind directions are reported to the nearest 10° , the observer estimating an average reading of a fluctuating indicator dial over a 1-minute period.⁽¹⁾

The term "range of wind direction change" used here is defined as the sum of the maximum clockwise (CW) and maximum counterclockwise (CCW) departures of wind direction from a given initial value reported during a given subsequent period. The calculation may be illustrated with the data in Table 1. Here the maximum clockwise change in wind direction (veering) reported during a particular interval of 6 hours was $170^\circ - 90^\circ = 80^\circ$, while the maximum counterclockwise change (backing) was $90^\circ - 70^\circ = 20^\circ$; the resultant total range of direction change is, therefore, defined as $80^\circ + 20^\circ = 100^\circ$. Throughout the study, overlapping periods have been used; thus, from a given set of n hourly observations, $n - m + 1$ samples of direction range were derived for each interval of m hours duration.

TABLE 1. Example of Hourly Reports of Surface Wind Direction

Hour	Direction, degrees
1	90
2	120
3	150
4	170
5	110
6	70

Figure 1 shows a number of consistent features for the behavior of the wind direction range as defined here:

1. The mean direction range steadily increases with length of period, almost linearly in the case of winds greater than 6 mph;

2. The direction range varies inversely with wind speed, light winds exhibiting ranges three or four times larger than those recorded for strong winds over comparable interval lengths; note this places a quantitative estimate on the qualitative property of relative steadiness known to characterize stronger winds;

3. Direction ranges for periods beginning at midnight are 10° to 20° smaller than those calculated for periods beginning at noon (cf. Figure 1). This is true even for periods of 24 hours and stems from the superposition of the diurnal wind effect on the effects of large scale pressure systems exerting an influence over the Chicago area for periods exceeding 24 hours.

Values of the maximum clockwise direction changes observed, as well as the largest changes recorded in the counterclockwise sense, are listed in Table 2 for periods beginning at noon and at midnight; period lengths and classes of initial wind speed are the same as those used in Figure 1. Also tabulated are the differences between clockwise and counterclockwise changes recorded over similar time intervals.

In all but a few cases, clockwise changes exceed those observed in

FIG. 1.--Mean wind direction range vs. period for indicated wind speeds at Chicago Midway Airport (1966-1970). a) All periods begin at 0000 CST; b) all periods begin at 1200 CST.

TABLE 2. Cyclonic and Anticyclonic Wind Direction Mean Deviations and Their Differences in Tens of Degrees as a Function of Period of Observation and Wind Speed Interval at Onset Time 0000 CST and 1200 CST at Chicago Midway Airport, January 1, 1966-December 31, 1970

Speed interval, mph	Period, hr																	
	3			6			9			12			18			24		
	CW	CCW	Diff.	CW	CCW	Diff.	CW	CCW	Diff.	CW	CCW	Diff.	CW	CCW	Diff.	CW	CCW	Diff.
0000 CST																		
0 - 5	3	2	1	4	3	1	6	4	2	8	5	3	11	6	5	12	7	5
6 - 10	2	1	1	3	2	1	5	3	2	6	3	3	8	4	4	10	5	5
11 - 15	1	1	0	2	2	0	3	2	1	4	3	1	6	4	2	7	4	3
16 - 25	1	1	0	2	1	0	2	2	0	3	2	1	5	3	2	6	5	1
26 - 35	1	0	1	1	1	0	1	3	-2	2	3	-1	2	4	-2	3	5	-2
1200 CST																		
0 - 5	4	4	0	6	5	1	7	6	1	9	7	2	10	7	3	12	8	4
6 - 10	2	2	0	4	4	0	5	5	0	6	5	1	8	6	2	10	7	3
11 - 15	1	1	0	2	2	0	3	3	0	5	4	1	6	5	1	8	6	2
16 - 25	1	1	0	2	1	1	3	2	1	4	3	1	5	3	2	7	4	3
26 - 35	1	1	0	2	1	1	2	2	0	3	2	1	6	2	4	6	2	4

the counterclockwise direction; that is, the wind can evidently be expected to veer further than it backs. The few exceptions noted in the table are associated with larger wind speeds and intervals of 9 hours or more. Larger direction changes in the clockwise direction are probably due in part to the more frequent passage of low and high pressure centers to the north of Midway Airport. Thus, average storm tracks over a given site can be used to assess probable wind direction changes qualitatively. Then too, stronger veering is also due to the greater curvature, on the average, of cyclonic isobars; there are dynamic reasons which prevent high pressure systems from attaining the same circulation intensity as low pressure centers.

The information presented here outlines certain considerations, previously largely ignored, which should be taken into account in forecasting the long term, larger scale effects downwind of pollutants released from given point sources (e.g., the effects upon a large metropolitan complex of a distant nuclear power installation). These results suggest a statistical approach whereby the systematic departures from a wind direction line at the beginning of a period of concern might be included in models designed to analyze the longer term consequences of mesoscale dispersion from point sources.

The assistance provided by Frank C. Kulhanek and Gunther A. Zerbe in preparing these reports is hereby acknowledged.

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MICROMETEOROLOGICAL MEASUREMENTS OVER THE GREAT LAKES

Paul Frenzen

Introduction

During the past three years a series of experimental studies of the interaction between the lowest layers of the atmosphere and the open waters of the Great Lakes has been carried out by the Atmospheric Physics Section. Descriptions of these experiments and preliminary findings are reported in this and previous annual reports [ANL-7760-III (1970) and ANL-7860-III (1971)]; papers of wider interest are being prepared for journal publication.

These observations add to our understanding of the mass-energy balance of the Great Lakes. Very few direct measurements of the atmospheric turbulent transfers of heat and moisture which control this balance have been carried out. Rather, what is known has either been inferred from observations of long-term seasonal changes in water level and heat content, or has been based upon extrapolations from results of studies similar to the present experiments, but conducted over considerably smaller bodies of water, such as artificial reservoirs. Therefore, one goal of the present program has been to test the applicability of these widely-accepted methods of estimation to the special problems of large lakes.

The Argonne investigations have employed the techniques of micrometeorological measurements at selected sites where observations may be considered representative of conditions over open water. The experiments have been designed to examine not only the normal daily cycle of the several components of the air-water exchange (i.e., the net radiative transfer and the turbulent transfers of both sensible and latent heat) but also the occasional discontinuous changes in these mechanisms associated with abrupt changes in meteorological conditions. By providing additional insight into the interrelationships of the mechanisms involved, these observations of both cyclical and aperiodic short-term variations should improve our ability to assess the possibility of irreversible long-term changes which could

result from unrestricted thermal discharge to the lakes.

Field Sites and Tower Facilities

Measurements have been made from instrumented, offshore towers erected and maintained in Lake Michigan by the Lake Survey Center, National Oceanic and Atmospheric Administration, and in Lake Ontario by the Canadian Center for Inland Waters. In 1970 a relatively small tower was placed some 700 m off the eastern shore of Lake Michigan, about 5 miles south of Muskegon; in 1971 and 1972 a larger tower was erected 1600 m offshore in the same general location. Additional measurements were made in the spring of 1972 from a tower located in western Lake Ontario, about 3 km north of the mouth of the Niagara River.*

The Muskegon, Michigan site offers a number of advantages, the most important to the purposes of the experiment being an exposure to westerly winds with past trajectories over water in excess of 100 km. The particular shore location on the north side of the Mona Lake inlet has been used as a temporary field station in previous years by workers from both the Lake Survey and the University of Michigan. The beach at this point is not readily accessible to the general public, and therefore cables lying in shallow water or crossing the beach and nearshore dunes are not disturbed, while apparatus installed on the tower and recording equipment housed on-shore remain relatively secure. The shore site is immediately adjacent to a small, navigable inlet suitable for small boat operations.

The 1970 tower stood 5.5 m above the surface of Lake Michigan in water about 7 m deep, while in 1971 and 1972 a larger tower extending 8.5 m above mean water level was erected in water 15 m deep. The larger installation also included a vertical extension which supported sensors at

* This facility was operated for the International Field Year for the Great Lakes program by the Canadian Center for Inland Waters and the Atmospheric Environment Services of Canada; the kind cooperation of both these organizations in arranging our informal participation is greatly appreciated.

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the 10-m level. The working platform was 3 m above mean water level in 1970 and about twice that high in 1971 and 1972. A marine power cable laid along the lake bottom brought 115/230 V ac service to the Michigan towers. When this cable failed in the fall of 1972, turbulence measurements then under way were completed by installing storage cells on the tower to supply 30 V dc to the sensing and telemetry equipment.

In 1970 an instrumentation cable on the lake bottom was used to transmit data to recording equipment onshore. Although not designed for marine service, this standard, polyvinyl-insulated six-pair data cable served adequately for several months, but a gradual deterioration of the insulation eventually limited its use to only two circuits. However, by the end of the first season's trials, a two-channel radiotelemetry link had been placed in operation. Designed and constructed by the ANL Electronics Division, this equipment was subsequently modified to transmit four pulse-modulated AM signals along with two FM signals. The transmitter has a nominal output of 5 W at a licensed frequency in the 215 to 260 MHz telemetry band.

The Lake Ontario tower, one of three erected on the Niagara bar for the IFYGL program, extended 16 m above the water surface. Power for these installations was supplied by a diesel electric generator mounted on a barge anchored about 200 m away on the inshore side. This group's long period recording equipment (described below) was operated from the 7-m level for a two-week period beginning in late May in order to obtain data over waters colder than had been observed previously in this program.

Instrumentation and Experimental Procedure (1971-1972)

Average hourly exchange rates of sensible and latent heat between the water and air were computed by the bulk aerodynamic method, using the following standard meteorological observations recorded on the tower in the form of one-hour averages:

- a) air temperature, ventilated thermistor at 7 m;
- b) water temperature, sealed thermistor mounted on a few cm below mean water level;

- c) humidity, miniature dual-core dew cell probe (Yellow Springs Model YSI 9102) mounted with air temperature sensor at 7 m; and

- d) wind speed, 3-cup anemometer (Belfort airways type) at 7.5 m.

In addition, a Funk-type polyethylene shielded net-radiometer (Middleton and Co., Pty. Ltd.) was mounted on the tower to provide the additional information required to compute hourly average thermal energy budgets for the air-water interface. An extensive set of results of this kind are presented by Frenzen et al. (1971) and Hart and Kittel (this report).

During the preliminary trials of 1970, mean meteorological data were transmitted through the data cable to recorders onshore. Because of the difficulties encountered, the mean system was converted for the 1971 season to a compact signal-conditioning unit and a strip chart recorder mounted on the tower itself. The recorder, a two-channel time-sharing device equipped with two event marking pens in the chart margin (Rustrak Model 3146), could record four continuous analog signals (air and water temperature, humidity, and net radiation) and two integrating, pulsed signals (wind speed and net radiation). For 1972 a further modification was made in order to simplify data reduction; integrators (voltage to frequency converters) and accumulating digital registers were installed in every sensor circuit, and a clock-controlled digital printer was arranged to record the hourly averages directly.

Turbulence instrumentation used during the past two seasons included the following equipment:

A. Wind sensors:

1. low-inertia, 6-cup anemometers (ANL design);
2. fast-response vertical vane (ANL design);
3. fast-response direction vane (ANL design);
4. crossed-mills, an experimental device (similar in principal to crossed hot-wires used in wind tunnel work) designed to resolve the horizontal and vertical wind fluctuation components and to supply signals for direct computation of momentum flux by the portable, variance-reduction digital computer, VARCOMP (ANL design);
5. vertical propellers (Gill type, R. M. Young Co., used in

both Canadian and Australian fluxatrons during cooperative experiments).

- B. Temperature sensors: miniature bead thermistors (Victory Engineering Corp.).
- C. Humidity sensors:
 - 1. Lyman-alpha humidimeter (Electromagnetic Research Corp.);
 - 2. air-capacitor refractometer (Hay type; stabilized circuitry designed and constructed by ANL; operated in cooperative experiments by H. C. Martin, Atmospheric Environment Service of Canada); and
 - 3. Gjessing crystal hygrometer (operated in intercomparison tests by H. C. Martin).

In various experiments signals from the foregoing sensors were used individually or in combination to compute turbulent exchange rates as follows:

- A. Momentum flux (surface wind stress):
 - 1. by direct eddy correlation:
 - a) low-inertia cup anemometer plus vertical vane (ANL digital system);
 - b) low-inertia anemometer plus vertical propeller (stress Fluxatron, operated in cooperative experiments by B. Hicks, C.S.I.R.O., Australia); and
 - c) cross-mills plus VARCOMP (ANL design, experimental).
 - 2. by indirect, high-frequency spectral method: low-inertia cup anemometer plus ANL digital system or VARCOMP.
- B. Sensible heat flux:
 - 1. by direct eddy correlation method:
 - a) low-inertia cup anemometer, vertical vane, and miniature bead thermistor (ANL digital system);
 - b) vertical propeller and miniature bead thermistor (CSIRO thermal Fluxatron).
 - 2. by indirect, high-frequency spectral method: miniature bead thermistor (ANL digital system or VARCOMP, experimental).
- C. Latent heat flux (evaporation):
 - 1. by direct eddy correlation method:
 - a) low-inertia cup anemometer, vertical vane, and Lyman-

alpha Humidiometer (ANL digital system);

b) low-inertia cup anemometer, vertical propeller, and refractometer (Canadian evaporation Fluxatron);

c) vertical propeller, Gjessing crystal hygrometer (CSIRO evaporation Fluxatron).

2. by indirect, high-frequency spectral method: Lyman-alpha Humidiometer (ANL digital system or VARCOMP, experimental).

Intercomparison of exchange estimates obtained by different methods of observation and/or calculation in various experiments are now being prepared. These results will include estimates of the numerical constants appropriate to the several methods of computation.

Results

Recent results of these studies appear in the three papers which follow this introduction; related experimental investigations have been summarized in the following publications. Further reports are in preparation.

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SOME AVERAGE DIURNAL SURFACE ENERGY BUDGETS OVER THE SOUTHERN GREAT LAKES

Richard L. Hart and Alan W. Kittel^{*}

Introduction

In order to provide basic information on the natural exchange processes which form the background to thermal pollution investigations, micrometeorological measurements have been made over the Great Lakes during several seasons. Wind speed, water and air temperatures, humidity, and net radiation have been recorded for continuous periods ranging from a few days to several weeks over southern Lake Michigan and, during one two-week spring period, over western Lake Ontario. From these data diurnal cycles of the several components of the thermal energy budget of the air-water interface have been calculated by bulk aerodynamic methods.

Daily observations obtained over Lake Michigan have been reported in some detail in the previous volume of this series.⁽¹⁾ In the present note observations recorded over Lake Ontario (28 May through 10 June 1972) are added, and all the surface thermal budget data have been collected in the form of average diurnal cycles more or less typical of conditions prevailing over the southern Great Lakes at various times of the year. Although the total sample is limited, these measurements represent an objective view of the natural context within which considerations of thermal pollution in the Great Lakes should be placed.

Theoretical Background

The net vertical flux of thermal energy across an air-water interface may be written:

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$$F_n = H + LE + R_n, \quad (1)$$

where H is the sensible heat flux, LE is the flux of latent heat associated with evaporation, R_n is the net (outgoing) radiation, and F_n is the net (outgoing) flux of thermal energy from the lake. Note that Eq. (1) defines all upward transports as positive; thus the sign convention usually adopted for net radiation is reversed. For these studies, the exchange rates of sensible and latent heat have been calculated using the bulk aerodynamic formulae:

$$H = \rho c_p D_H \bar{U}_z (T_o - T_z) \quad (2)$$

$$LE = \rho L D_E \bar{U}_z (q_o - q_z). \quad (3)$$

Here ρ is the air density, c_p is the specific heat of air at constant pressure, L is the latent heat of vaporization of water, and D_H and D_E are bulk transfer coefficients for heat and water vapor; $\bar{U}_z$, T_z , and q_z represent mean wind speed, ambient temperature, and ambient specific humidity measured at height z , while T_o is the water surface temperature and q_o is the saturation specific humidity corresponding to T_o .

Experimental Procedure and Analysis

Meteorological data needed to evaluate Eqs. (2) and (3) were measured and recorded by the equipment previously described.^(1,2) Observations for the July, October, and November periods were made over Lake Michigan in 1971 from the Muskegon tower.⁽²⁾ The May-June data* were recorded in 1972 from one of the towers erected in western Lake Ontario for the program of the International Field Year for the Great Lakes. This tower was located on the Niagara River bar about two miles offshore on the Canadian side of the international boundary.

* For the opportunity to make these measurements over colder water we are indebted to H. C. Martin of the Atmospheric Environment Service of Canada and to M. Donnelan and F. Elder of the Canadian Center for Inland Waters.

Using $D_H = D_E = 1.35 \times 10^{-3}$ as transfer coefficients appropriate to the 7.6-meter level at which the ambient measurements were made, ⁽¹⁾ hourly average turbulent fluxes of H and LE were calculated. Combining these results with observed values of R_n , diurnal cycles of F_n were constructed according to Eq. (1). Individual 24-hour budgets were then averaged for periods from several days to two weeks to produce the "typical" diurnal cycles shown in Figures 1 and 2. This averaging procedure minimizes short period variations in thermal exchange associated with daily weather conditions and emphasizes those changes due to seasonal variations in lake temperature, net radiation, and the average climatological characteristics of the air moving over the lake. Keeping in mind that these average cycles will be modified, often severely, by changing weather conditions (see discussion below), the results summarized in Figures 1 and 2 may be considered to be more or less typical for the time of year in which the observations were taken. They may be taken to be more representative of the expected average daily exchange over a week than for the exchange which might occur on any single day.

Discussion

Figure 1(a)

Fairly low wind speeds prevailed up to the last day of this observation period. For one half of the period high humidity combined with the low wind speeds to restrict evaporation severely. Evidently the humidity was high enough to cause condensation on several occasions. It should be noted that the bulk aerodynamic relations become questionable in these circumstances and that the shields of the net radiometer become essentially opaque to longwave radiation when wet. The latent heat transfer averaged only 0.5 mW/cm^2 during this time, whereas, excluding the last day, LE averaged 4.0 mW/cm^2 during the rest of the period. Low wind speeds also restricted the rate of exchange of sensible heat, which averaged -1.3 mW/cm^2 up to the last day of the period. On the evening of 9 June the passage of a cold front brought higher winds and cooler, drier air, sharply increasing

FIG. 1.--Typical diurnal cycles of the terms entering the surface thermal energy budget over the southern Great Lakes for the dates indicated. R_N , net radiation, —; H , sensible heat flux, ---; LE , latent heat flux, -.-.-; F_N , net flux of thermal energy, Positive fluxes indicate upward transfer. U is the mean wind speed, T_S the surface temperature, T_A the ambient temperature, and T_D the dew point temperature.

sensible heat transfer and evaporation. Average rates for sensible and latent heat flux on 10 June were 8.3 and 18.2 mW/cm^2 , respectively.

Figure 1(b) and (c)

Low values of sensible and latent heat flux in Figure 1(b) are attributable to predominantly S-SW winds during the period. A frontal passage on the evening of 25 July brought in a mass of slightly cooler, much drier air with a shift in wind direction to W-NW. Note that the increase in thermal losses due to sensible heat exchange and evaporation after the frontal passage was almost sufficient to completely offset the net radiation in the mean [Fig. 1(c)]; F_n for the entire week is reduced to -0.7 mW/cm^2 .

Figure 2(a)

Relatively warm, moist southerly winds prevailing during this observation period account for the small evaporative losses and for the negative values of sensible heat flux at a time of year when one would ordinarily expect significant losses due to both mechanisms. Above normal cloudiness is reflected in the diurnal cycle of net radiation. On the average, there was very little exchange of thermal energy during this period.

Figure 2(b)

Although this sample is quite limited, the observations are included because they were made during the passage of an open wave cyclone, shortly after the first major polar outbreak of the season. As the net radiation averaged nearly zero, the net flux of thermal energy was dominated in the mean by the turbulent fluxes.

The dependence of the diurnal cycle of F_N upon that of R_N is obvious in each figure. The effect of seasonal variations of R_N on F_N may be ascertained by comparison between the figures. Figure 1(a) illustrates a case in late spring when, with light winds, radiation dominates the energy budget, while Figures 2(a) and (b) exemplify fall cases when incoming solar radiation is reduced to such a degree that the turbulent fluxes dominate the budget in the mean.

Long-term average rates of sensible heat transfer are rather small, usually only a few mW/cm^2 , with the largest outgoing values normally

FIG. 2.--Typical diurnal cycles of the terms entering the surface thermal energy budget over the southern Great Lakes for the dates indicated. R_N , net radiation, —; H , sensible heat flux, ---; LE , latent heat flux, -.-.-; F_N , net flux of thermal energy, Positive fluxes indicate upward transfer. U is the mean wind speed, T_S the surface temperature, T_A the ambient temperature, and T_D the dew point temperature.

occurring in the fall. Although this note contains no data to determine the behavior in early spring, it can be expected that the largest energy gains normally occur at this time. As the lake is warmest in mid- and late summer, long-term average rates of evaporation are normally greatest at this time of year. During the period of 19 July through 1 August 1971 the rate of latent heat exchange averaged 10.7 mW/cm^2 , a rate much larger than for any other period of comparable duration.

The "typical" days presented here are in reality only average days for relatively brief periods of observation. On any given day these diurnal cycles will be modified by conditions prevailing at the time. Modification of the exchange rates by a marked change in weather conditions is apparent in the contrast between Figures 1(b) and (c). Strong modification of the daily average exchange rates by varying weather conditions is exemplified by the minimum and maximum 24-hour mean rates of sensible and latent heat flux that have been observed. On 3 June 1972 the largest transfer of sensible heat into the lake and the smallest losses due to evaporation were observed, $\overline{H} = -4.5 \text{ mW/cm}^2$ (downward) and $\overline{LE} = 0.2 \text{ mW/cm}^2$. Only one week later, the largest turbulent thermal transfers out of the lake, $\overline{H} = 8.3 \text{ mW/cm}^2$ and $\overline{LE} = 18.2 \text{ mW/cm}^2$, were recorded.

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ON THE MEASUREMENT OF EVAPORATION BY THE INDIRECT HIGH-FREQUENCY METHOD

Paul Frenzen

Introduction

Since much of the energy transferred to the atmosphere from lakes or ponds is in the form of latent heat, a simple procedure for measuring evaporation from open water surfaces would be useful to many thermal pollution investigations. For the particular problem of assessing thermal transfer from artificial cooling ponds, measurement of this important component of the total energy exchange is complicated by the fact that the air is over the water for only a short time, and therefore the steady transfer regimes within which representative measurements can be made are quite shallow. As a rule, the lowest layers of the atmosphere are in approximate statistical equilibrium (90 percent adjustment) with conditions at their lower boundaries only to heights of about 0.5 percent of the distance the air has traveled over a uniform surface.⁽¹⁾ For cooling ponds of average size, this requires that measurements be limited to the lowest few meters above the water. Further, since turbulent exchange near the surface is accomplished by small scale eddies of relatively high frequency, fast-response instrumentation must be employed if the turbulent transfer rates are to be measured rather than estimated by bulk-average methods.

When turbulent fluxes are computed from high-frequency measurements, data reduction can be simplified by using the indirect or "spectrum-slice" method employed with some success for momentum flux (e.g., Ref. 2). What is more, since this kind of analysis can be applied to steady-state, vertical exchange of any quantity for which the lower boundary acts as a uniform source or sink, evaporation over open water is a natural candidate for application of the method. The principle simplification effected derives from the fact that only variances rather than covariances are calculated. Although somewhat higher frequency response is required of the equipment, small sensor misalignments and some low-frequency motion of the supports can be

tolerated. Therefore, semistable rafts or buoys rather than rigid masts can be used to mount the equipment.

The 1972 Muskegon tower program⁽³⁾ included several periods of trial evaporation measurements using the indirect method. The results are promising but somewhat inconclusive, in part because of the difficulty of obtaining reliable independent evaporation measures for comparison. Since this operationally simple procedure will be employed in studies of cooling ponds and thermal plumes in Lake Michigan during the coming season, the method is presented here in some detail.

Theoretical Analysis

In the inertial subrange of the turbulence spectrum the variance of a fluctuating scalar variable acting as a passive additive to the field of motion is neither generated nor destroyed. Rather, in a process analogous to the kinetic energy cascade over a similar frequency range, fluctuation variance is transferred from larger to smaller eddies until it reaches scale sizes sufficiently small for local variations in concentration of the scalar quantity to be smoothed out by molecular diffusion.* Somewhat surprisingly, experiments have shown that, for the higher frequency components of interest to this analysis, both temperature and humidity can be treated as passive quantities whose associated density anomalies do not appreciably distort the dynamics of the motion field. Conceivably, there is insufficient time for the buoyancy of small density differences to affect the rapid velocity fluctuations.

As in the familiar example of the kinetic energy spectrum, frequency distributions of scalar variance density in the inertial subrange can be written from dimensional considerations alone. For humidity fluctuations, we can write

$$F_q(n) = \beta \bar{u}^{-2/3} N \epsilon^{-1/2} n^{-5/3} \quad (1)$$

(e.g., see Ref. 4, p. 283). With few exceptions the notation used here and throughout the remainder of the discussion is standard; for convenience, all symbols are listed at the end of the article.

* The rate at which this destruction of variance occurs is defined as N , the scalar analogue of the turbulent energy dissipation ϵ .

In the atmospheric boundary layer the $-5/3$ frequency dependence predicted by Eq. (1) and similar relations for other fluctuating variables in the inertial subrange is observed to extend to frequencies as low as $\bar{u}/z$ Hz. Strictly speaking, the required properties of isotropy do not exist below $10 \bar{u}/z$ Hz; but, again appealing to experiment, field tests have shown that good estimates of higher frequency variance can be obtained by extrapolating according to Eq. (1) from measurements made near $\bar{u}/z$. The magnitude of the proportionality constant β depends upon the units of frequency employed and upon the normalization adopted for the spectral integral. Here we define

$$\int_0^{\infty} F_q(n) dn = \bar{q}^2 ; \quad (2)$$

consequently, the magnitude of β will be twice that found by those who arbitrarily include a factor of $1/2$ in the above in order to parallel the normalization of velocity variance spectra in terms of turbulent kinetic energy.

For the rate of dissipation of humidity fluctuations integration of Eq. (1) over an octave from n to $2n$ gives

$$N = \frac{\epsilon^{1/3}}{0.555\beta} \left(\frac{n}{\bar{u}}\right)^{2/3} \sum_n^{2n} F(q) . \quad (3)$$

A second relation for N derives from the turbulent balance equation:

$$N = \frac{E}{\rho} \left(- \frac{d\bar{q}}{dz} \right) \quad (4)$$

(e.g., see Ref. 4, p. 95). By retaining only the gradient production term in this fashion, the foregoing assumes that fluctuation variance is primarily dissipated locally.

Two expressions for N suffice to write a relation for E , but it is convenient to include the dimensionless specific humidity gradient

$$\phi_w = (\rho u_* k z / E) (d\bar{q}/dz) \simeq (1 - 15 z/L)^{-1/2} \quad (5)$$

which, in the unstable stratifications of principal interest to investigations of thermal transfer over heated ponds, can be approximated as shown in the last expression on the right. There are some differences of opinion on the exact value of the numerical factor included in the empirical expression for ϕ_w (cf. Refs. 5 and 6), but the consequences are small, and use of this relation in place of the experimentally difficult measurements of $d\bar{q}/dz$ is amply justified. Inclusion of ϕ_w also has the useful effect of increasing the degree of the final equation in E , thus reducing its sensitivity to the imperfectly known spectral constant, β .

Combining Eqs. (3), (4), and (5) results in the indirect evaporation relation:

$$\frac{E}{\rho} = \left(\frac{u_* k z}{\phi_w} \right)^{1/2} \left(\frac{\Sigma F_q}{0.555 \bar{u}} \right)^{1/2} \left(\frac{n}{\bar{u}} \right)^{1/3} \epsilon^{1/6} \quad (6)$$

Converting the flux of water mass E to an equivalent latent heat transfer H_w in mW cm^{-2} and inserting values for those terms essentially constant during the trials gives the test equation

$$\left(\frac{u_* \Sigma F_q}{\phi_w} \right) \left(\frac{n}{\bar{u}} \right)^{2/3} \frac{\epsilon^{1/3}}{H_w^2} = 4.91 \times 10^{-4} \beta, \quad (7)$$

which is to be examined by experimental determination of the constant β . The numerical factor on the right includes $\rho = 1.23 \times 10^{-3} \text{ g cm}^{-3}$, $z = 900 \text{ cm}$, and $k = 0.35$. In addition, u_* and $\bar{u}$ are in cm sec^{-1} , ϵ is in $\text{cm}^3 \text{ sec}^{-2}$, and ΣF_q is expressed in mb^2 , units of vapor pressure being appropriate to humidity measurements obtained with a Lyman-alpha Humidimeter during the trials.

Momentum and kinetic energy statistics (u_* and ϵ) required for Eq. (7) are also obtained by indirect methods. Briefly, ϵ is derived from an integral of the one-dimensional spectrum of kinetic energy density:

$$\int_n^{2n} F_u(n) dn = \alpha (\epsilon \bar{u})^{2/3} \int_n^{2n} n^{-5/3} dn = 0.555 \alpha (\epsilon \bar{u}/n)^{2/3} \quad (8)$$

in which $\alpha = 0.55$ (e.g., Ref. 7). Assuming local dissipation, the balance equation for turbulent kinetic energy reduces to

$$\epsilon = u_*^2 \frac{d\bar{u}}{dz} = \frac{(H_s + 0.07 H_w)}{\rho c_p} \frac{g}{\theta} \equiv \pi_\tau + \pi_B, \quad (9)$$

where the effects of both sensible and latent heat flux have been included in the buoyancy production π_B (e.g., see Ref. 8, p. 96). Upon substituting the dimensionless wind shear, ⁽⁵⁾

$$\phi_m \equiv (kz/u_*) (d\bar{u}/dz) = (1 - 15 z/L)^{-1/4}, \quad (10)$$

Equation (9) can be reordered to obtain

$$u_* = \left[\frac{(\epsilon - \pi_B) kz}{\phi_m} \right]^{1/3}. \quad (11)$$

The stability-dependent terms in Eqs. (6), (7), and (11) can be obtained from bulk aerodynamic estimates of sensible and latent heat flux through the numerical relations (with H in mW cm^{-2})

$$\pi_B \equiv \frac{(H_s + 0.07 H_w)}{\rho c_p} \frac{g}{\theta} \approx 2.75 H_t \quad (12)$$

and

$$-z/L \equiv \pi_B/\pi_\tau = \left(\frac{H_t}{\rho c_p} \frac{g}{\theta} \right) / \left(\frac{u_*^3}{kz} \right) \approx 0.96 \frac{H_t z}{u_*^3}, \quad (13)$$

in which the numerical factors correspond to the conditions of these trials. To obtain ϕ_m for Eq. (11) one can "bootstrap" by making an initial estimate of u_* to determine an initial z/L from Eq. (13); Eq. (10) then gives an initial ϕ_m , which through Eq. (11) determines an improved u_* . This result can be used to find an improved z/L and so on to an improved ϕ_m and a final u_* . The ultimate determination of evaporation rate is not overly sensitive to this stepwise approximation; u_* varies only as $\phi_m^{-1/3}$ and only $u_*^{1/2}$ enters Eq. (7) for E .

It is useful to note that the first factors in parentheses on the right of Eq. (6) can be rewritten in a form more easily evaluated in the field. Since $\phi_w^{1/2} \equiv \phi_m$ [cf. Eqs. (5) and (10)] these terms become

$$(u_* k z / \phi_w)^{1/2} = (u_* k z)^{1/2} / \phi_m = u_*^{3/2} (k z)^{-1/2} (d\bar{u}/dz)^{-1}. \quad (14)$$

This substitution results in a working relation for evaporation requiring only measurements of humidity fluctuations, horizontal wind fluctuations, and mean wind shear. When π_B can be neglected in comparison to π_T (i.e., when $z/L \simeq 0$ as in near neutral stratifications where π_B is small, or close to the surface where π_T becomes large), these same terms can be combined with ϵ through the neutral approximations

$$\phi_w^{1/2} \simeq 1 \quad \text{and} \quad \epsilon \simeq u_*^3 / k z \quad (15)$$

to obtain

$$(u_* k z / \phi_w)^{1/2} \epsilon^{1/6} \simeq (u_* k z)^{1/2} (u_*^3 / k z)^{1/6} = u_* (k z)^{1/3}. \quad (16)$$

With this simplification the indirect evaporation relation reduces to

$$\frac{E}{\rho} = u_* \left(\frac{n k z}{\bar{u}} \right)^{1/3} \left(\frac{\Sigma F_q}{0.555 \beta} \right)^{1/2}, \quad (17)$$

a form suitable for measurements at small elevations over unheated, natural lakes.

Experimental Results

Conditions for the three days in October when observations for the trial indirect evaporation calculations were made are summarized in Table 1. Winds were directed onshore in all cases, but the southerly flows of the 27th may have approached the curving, local shoreline (oriented NNW-SSE near the tower site) with over-water trajectories as short as 10 km. On the other two days, wind fetch over water exceeded 100 km.

Air temperatures and humidities were recorded at the 6-m level by a

TABLE 1. Average Conditions during Experimental Trials

Time (EST)	Wind, m sec ⁻¹	Temp., deg C	Flux, mW cm ⁻²			
			air	water	H _s	H _w
Oct. 9 (1418-1618)	NW	6.6	7.3	7.5	-0.3	- 6.0
25 (1304-1717)	WSW	10.0 ^(a)	8.9	10.5	-3.5	-10.7
27 (1043-1352)	S	7.0	10.8	10.9	-0.2	- 6.1

(a) Winds extrapolated from $z = 4$ m.

standard hygrothermograph located in a small shelter on the tower platform. These records were supplemented with periodic Assman psychrometric measurements taken from the tower before or after turbulence observation runs, or from a small boat downwind during observation periods. Water temperatures were obtained from bucket samples drawn near the tower base; these were reduced 0.2 to 0.3 deg to allow for surface cooling, the larger figure being applied to the conditions of more active evaporation on the 25th. Average wind speeds were extracted from the records of the turbulence instrumentation; wind directions were observed from shore.

Turbulent fluctuations of the horizontal wind were detected with low-inertia, cup anemometers of ANL design (distance constant ≈ 35 cm, see Ref. 9), while a Lyman-alpha Humidiometer (response time $\ll 0.1$ sec; Electromagnetic Corp. Model BLR) was used to record humidity fluctuations. These sensors were mounted 9 m above the water surface, clear of all other installations on the tower. As noted in the table, the anemometer at the 9-m level failed on the 25th, and it was necessary to extrapolate all wind data ($\bar{u}$, u_* , and ϵ) from measurements made at 4 m, this resulting in a noticeable deterioration in results as will be seen. Variance spectra of wind and humidity fluctuations were computed by the cosine transform autocorrelation method. In every case, both F_u and F_q were integrated over the same octave, from 1 to 2 Hz. This frequency band may have been less suitable for the 10-m sec⁻¹ winds of the 25th ($\bar{u}/z \approx 1.1$ Hz) than it was for the 7-m sec⁻¹ winds

of the 9th and 27th ($\bar{u}/z \simeq 0.8$ Hz). However, this factor is unlikely to have caused the differences observed between the cases.

Table 2 lists the observations obtained in the form required to evaluate the test relation, Eq. (7), the last column giving the result of each run as an estimate of β for frequencies in Hz and spectra normalized in terms of the complete variance. Considerable variation on the part of the supposed constant is evident, but most of it is eliminated if the dubious results of the 25th are ignored. It will be recalled that on this occasion, all wind data were extrapolated from 4 m. Observations on the other two days indicate $\beta = 0.45$ with a standard deviation of 0.08 over the ten estimates. This is equivalent to $\beta' = 0.77 \pm 0.14$ for n expressed as natural frequency, sec^{-1} , and a spectral normalization [Eq. (2)] in terms of the semivariance, $1/2 \overline{q^2}$; therefore, the present result compares quite favorably with magnitudes given by others using this convention. Paquin and Pond⁽⁷⁾ obtained $\beta' = 0.82 \pm 0.12$ using an approach based on the computation of higher order spectral statistics.

Additional trials during the coming season will use improved equipment to reduce the uncertainties on β . A dew point hygrometer will be employed to check both ambient humidity values and the local slope of the Lyman-alpha Humidimeter calibration curve, the latter being a significant source of uncertainty since this device steadily drifts as sensitive components age. In addition, improved covariance equipment now being prepared will make it possible to determine independent measures of evaporation rates for comparison experiments without depending upon bulk-aerodynamic approximations.

It can be noted that the application of indirect evaporation measurements to cooling pond and thermal plume studies need not await further refinements to the magnitude of β . In the demanding circumstances of short fetch, low elevations and relatively high frequency turbulence observations required by these investigations, indirect evaporation rate calculations based on the present determination of $\beta = 0.45$ should give results as good or better than estimates obtained by virtually any other method (save perhaps,

TABLE 2. Results of Experimental Indirect Evaporation Trials Expressed as Terms in the Test Relation, Eq. (7) (a)

Time	u_*	$\times$	ΣF_q	$\div \phi_w$	$\times (n/u)^{2/3}$	$\times \epsilon^{1/3}$	$\div H_w^2$	$\div 4.91 \times 10^{-4}$ $= \beta$
<u>9 Oct</u>								
1418	22.5		0.00921	0.70	0.0126	3.20	43.4	0.56
1448	21.6		0.00797	0.70	0.0135	3.05	36.4	0.57
1518	21.4		0.00672	0.73	0.0135	3.06	31.7	0.52
1548	22.2		0.00709	0.73	0.0131	3.17	34.7	0.52
<u>25 Oct</u>								
1304	34.7		0.0187	0.77	0.0104	4.96	132.2	0.67
1334	35.4		0.0177	0.79	0.0102	5.07	125.4	0.67
1404	37.0		0.0174	0.80	0.00985	5.31	125.4	0.69
1434	37.7		0.0187	0.82	0.00965	5.43	116.6	0.79
1517	39.9		0.0184	0.86	0.00947	5.76	110.2	0.86
1547	40.6		0.0231	0.88	0.00943	5.88	112.4	1.07
1617	41.3		0.0180	0.89	0.00927	5.99	112.4	0.84
1647	40.4		0.0174	0.89	0.00936	5.86	99.6	0.89
<u>27 Oct</u>								
1043	23.4		0.00629	0.63	0.0120	3.27	40.2	0.46
1113	21.5		0.00420	0.61	0.0133	2.94	31.9	0.37
1152	21.0		0.00542	0.63	0.0132	2.88	39.8	0.35
1222	22.2		0.00679	0.74	0.0124	3.12	45.6	0.35
1252	21.8		0.00780	0.84	0.0130	3.10	39.1	0.42
1322	23.5		0.0100	1.00	0.0124	3.45	47.9	0.42

(a) Units as described in the text.

high-frequency covariance observations). Furthermore, since H_w (or E) varies only as $\beta^{-1/2}$, uncertainties of 10 or more percent in the constant contribute errors of no more than 5 or 10 percent to the final measurement.

LIST OF SYMBOLS

α	the Kolmogorov constant for the one-dimensional kinetic energy spectrum ($= 0.55$)
β	the Kolmogorov constant for scalar, passive-additive spectra (humidity and temperature, e.g.)
c_p	the specific heat of air at constant pressure
E	the rate of evaporation; vertical flux of specific humidity
ϵ	the rate of dissipation of kinetic energy
$F_q(n)$	the variance density of spectra of scalar, passive additives
$F_u(n)$	the variance density of the one-dimensional kinetic energy spectrum
g	the acceleration of gravity
H_s	the vertical flux of sensible heat
H_w	the vertical flux of latent heat
H_t	$\equiv (H_s - 0.07 H_w)$; the total, buoyancy-effective heat flux
θ	the temperature of the air (superbar = mean value)
k	the von Karman constant ($= 0.35$)
L	the Obukhov length ($\equiv (\pi_\tau z) / \pi_B$)
n	fluctuation frequency
N	the rate of dissipation of the variance of scalar, passive additives
π_B	the rate of production of turbulent kinetic energy by buoyancy
π_τ	the rate of production of turbulent kinetic energy by wind shear
q	the specific humidity (superbar = mean value)
ρ	the air density
τ	the surface stress; the vertical flux of momentum
u	the horizontal wind (superbar = mean value)
u_*	the friction velocity ($= (\tau/\rho)^{1/2}$)
ϕ_m	the dimensionless gradient function for horizontal wind velocity
ϕ_w	the dimensionless gradient function for humidity
z	the height above the surface

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ATMOSPHERIC DISPERSION OVER WATER INFERRED FROM TURBULENCE STATISTICS

Paul Frenzen and Richard L. Hart

Introduction

Reduced dispersion of atmospheric pollutants in flows over water is a well-known characteristic attributable to the small aerodynamic roughness of the surface and a consequent decrease of turbulence production by shearing stresses. When this inherently weak potential for mechanical mixing is further damped by the negative buoyancy forces caused by stabilizing air-water temperature differences, plumes emitted by point sources can stream over water for considerable distances with negligible dilution. These conditions of restricted atmospheric dispersion are of special concern to the Atomic Energy Commission since many nuclear power stations operating at present are in coastal locations and sites offshore have been proposed for future installations.

Diffusion experiments over water have recently been initiated by another AEC/DBER group.⁽¹⁾ Although such direct measurements will no doubt prove useful, acceptable estimates of the reduced rates of plume dispersal to be expected in these circumstances can be calculated from fixed-point turbulence measurements that can be recorded with less difficulty. The relative simplicity of the method is particularly useful since dispersion estimates over water will be required over a wide range of wind speed and stability conditions. This note reports the results of a short series of test measurements of the data required, the trials having been included in the 1972 Muskegon tower experiments conducted over Lake Michigan.

Theoretical and Experimental Background

Beginning with Cramer⁽²⁾ in 1957 and Hay and Pasquill⁽³⁾ in 1959, many experimental studies over land have shown that standard deviations of the lateral and vertical pollutant concentrations in the Gaussian plume model, σ_y and σ_z , can be related to standard deviations of the lateral and

vertical fluctuations of wind direction, σ_θ and σ_ϕ , recorded near the source [see summary discussion in Meteorology and Atomic Energy⁽⁴⁾]. Formally the relation is not a simple one in that observations in fixed, Eulerian coordinates are used to describe a phenomenon, turbulent dispersion, more appropriately referred to the Lagrangian coordinates of moving fluid elements. Nevertheless, by introducing certain simplifications involving data averaging and implying an Eulerian-Lagrangian scale factor,⁽³⁾ it has been shown that useful estimates of dispersion distributions downwind of a point source can be obtained by this method. In particular, Islitzer⁽⁵⁾ has shown that acceptable σ_y estimates up to 2 or 3 km downwind of an elevated source can be made in unstable conditions using the simple relation

$$\sigma_{\theta_5} x = 1.23 \sigma_y, \quad (1)$$

in which σ_{θ_5} represents the standard deviation of the horizontal wind direction fluctuations derived from 5-sec, end-to-end averages read from a graphical record.

An approximation to the operational requirement for the measurement of wind fluctuations over open water was carried out by Slade,⁽⁶⁾ who compared the statistics of wind fluctuations recorded by instruments located on the upwind and downwind edges of a broad estuary. These observations indicated that dispersion rates over water are between 3 and 20 percent of those prevailing in similar winds over the nearby, fairly rough land, depending upon whether the air was warmer (stable) or colder (unstable) than the water surface. These findings were subsequently used by Van der Hoven⁽⁷⁾ to assess the consequences to reactor safety analyses of the changing dispersion regimes which characterize coastal locations. More recently, Wierenga⁽⁸⁾ has reported measurements of vertical wind fluctuations recorded over a moderately large body of water (10-km fetch) in a wide range of atmospheric stabilities. Since lateral wind fluctuations did not figure in this study of vertical flux behavior, σ_θ data of particular interest to dispersion analyses were not included.

Experimental Arrangements

Observations described here were obtained from a tower located in 15 m of water, 1.6 km off the eastern shore of Lake Michigan [see Frenzen⁽⁹⁾]. Data were recorded on 7, 8, and 9 October 1972 in wind speeds of 5 to 12 m sec⁻¹ with wind directions ranging between WSW and NNW. All winds were directed onshore, the upwind fetch over open water exceeding 100 km in every case. In the early morning hours of the 7th the air was as much as 1.5 deg warmer than the water, while at dawn on the 9th it was 1.8 deg colder. More to the point of specifying the stability damping or enhancement effects upon turbulence in the lower atmosphere, the ratio z/L (height of observation relative to the Obukhov stability length) varied from +0.02 to -0.24; that is, the stratification conditions ranged from slightly stable to significantly, if not moderately, unstable.

Vertical components of wind direction fluctuations were recorded 9.6 m above the water surface using a bivane of ANL design. Adjusted for slightly underdamped response in the usual fashion, this device has a distance constant of 1.3 m. Horizontal direction fluctuations were measured at the 8.0-m level using a small vane assembled especially for these trials. Unfortunately, the dynamic characteristics of this device are not available since the exact configuration in which it was used was not recorded. The horizontal sensor appeared to respond significantly faster than the bivane, to a first approximation as much as twice as fast. Wind speeds were measured with an ANL low-inertia cup anemometer which has a distance constant of about 35 cm [Frenzen⁽¹⁰⁾]. This sensor was mounted at an elevation of 9.4 m, just below the bivane.

Results

Fluctuations in wind speed and direction detected on the tower were transmitted to shore by radiotelemetry, where they were recorded as audio frequencies on magnetic tape. The taped data were subsequently integrated over intervals of 0.5 or 0.6 sec, and standard deviations of these slightly

smoothed data sets were computed for sampling periods of 30 min. The latter were then combined in overlapping longer samples of one hour duration.

Wind direction fluctuations were converted to equivalent component turbulent velocity fluctuations according to the relations

$$u' \simeq s', \quad v' = s' (\sin \theta') \quad \text{and} \quad w' = s' (\tan \phi'), \quad (2)$$

where primed values denote "instantaneous" departure (0.5- to 0.6 sec averages) referred to 30-min mean values. Here s' is the horizontal wind speed fluctuation recorded by a cup anemometer, u' , v' , and w' designate orthogonal turbulent velocity components in the usual sense, while θ' and ϕ' represent fluctuation angles of horizontal and vertical wind direction, respectively.

These observations confirm the useful fact that lateral wind fluctuations over extensive water surfaces tend to be symmetrically distributed about their longer term mean values; evidently no significant correlation exists between left or right deflections and plus or minus speed fluctuations since the uniform lower surface virtually eliminates the possibility of transverse momentum fluxes in the mean. Consequently, one can simply compute

$$\sigma_v \simeq \bar{s} (\sin \sigma_\theta), \quad \text{rather than} \quad = \left[\overline{(s \cdot \sin \theta)^2} \right]^{1/2}. \quad (3)$$

A linear correlation of some 50 pairs of σ_v estimates calculated from the present observations by the two methods gave $R = 0.99$. A similar comparison of σ_w calculations gave a somewhat smaller correlation of $R = 0.95$, the difference attributable to the fact that significant mean vertical fluxes of momentum do occur, and therefore non-zero correlations between horizontal speed and vertical direction fluctuations exist. However, it is useful to note that since momentum fluxes over water are small, the approximate calculation can evidently be used for σ_w as well as σ_v with little error.

Figure 1 shows the results of the Lake Michigan trials in the form of component, root-mean-square velocity fluctuations normalized by the scale velocity, u_* . The velocity fluctuation ratios obtained and indicated here for

FIG. 1.--Normalized root mean square turbulent velocity fluctuation σ_u/u_* , σ_v/u_* , and σ_w/u_* , recorded in various stabilities at heights of 9.5 (σ_u and σ_w) and 8 m (σ_v) over Lake Michigan. Vertical lines indicate standard error of estimate ranges for numbers of samples shown; dashed lines indicate stability trends drawn by eye.

the several components in neutral stratification (i.e., at $z/L = 0$) compare quite favorably with values previously observed over land [e.g., see Lumley and Panofsky⁽¹¹⁾]. Of particular interest is the observation $\sigma_w/u_* \simeq 1.25$ which closely corresponds to the value now widely accepted for the neutral case over land. In the present measurements over water, this result obtains after the data are corrected for attenuation by the limited response of the bivane, the correction being based upon the second order response characteristics of this instrument. In contrast Wierenga⁽⁸⁾ reports $\sigma_w/u_* \simeq 1.0$ for neutral conditions over water, but it seems likely that a correction for sensor response has been omitted. The uncorrected vertical velocity fluctuation ratio observed in the present trials was, in fact, about 1.05. Since the distance constants of the bivanes used by Wierenga and ourselves were similar (1.15 and 1.3 m, respectively), the two sets of σ_w ratio observations in neutral conditions over water are probably closely equivalent.

For longitudinal velocity fluctuation measurements, comparatively fast sensors of nearly identical response characteristics (distance constants of 0.30 and 0.35 m) were used in both studies. Since there is little fluctuation variance in the higher frequencies attenuated by these sensors, no correction was applied in either case. The two studies essentially agree on $\sigma_u/u_* \simeq 2.5$ at $z/L = 0$, this in spite of the fact that the present data are obtained by combining 1/2-hr means, whereas Wierenga's results are based on 10-min sampling periods.

The trends with stability indicated by the dashed lines in Figure 1 have been drawn by eye and should be considered preliminary estimates. Additional measurements are needed in order to reduce scatter and extend the observations over a wider stability range. It should be noted that Wierenga's observations of σ_u and σ_w suggest somewhat less variation with stability in the relatively narrow range covered by the present data. He finds little change with stability in a widely scattered set of σ_u/u_* evaluations, while his report of a 10 percent increase in σ_w/u_* is only about one third that shown here.

Estimates of plume dispersion over open water were obtained from the present measurements through the empirical expression developed by Islitzer⁽⁵⁾ for an elevated source. Although the analysis by Hay and Pasquill⁽³⁾ indicates that fluctuation averaging periods should be adjusted for time of travel (i.e., for the distance downstream to the point of σ_y estimation), the simplified Islitzer relation [Eq. (1), above] has been found to give adequate estimates of lateral plume dispersion up to distances of several kilometers using a constant 5-sec direction fluctuation averaging interval and introducing a numerical proportionality factor. For the present application, the taped records of wind direction were integrated over 6 sec, the slightly longer interval being more convenient to the reduction of these data. Standard deviations of the resulting smoothed series (σ_{θ_6}) were then computed for consecutive one-hour sampling periods.

Figure 2 shows the variation of σ_{θ_6} with stability observed, a dashed line again drawn by eye to fit grouped averages. Characteristic values thus indicated for several stabilities were substituted in Eq. (1) to develop the listing in Table 1 of lateral plume dimensions to be expected at a downstream distance of 1 km. For comparison (if only qualitatively),

FIG. 2.--Root mean square horizontal wind direction fluctuations σ_{θ_6} in degrees of azimuth. Data smoothed over a 6-sec averaging interval for dispersion estimates 1 km downstream (see text), dashed line indicating stability trend drawn by eye.

TABLE 1. Comparative plume dimensions (σ_y) predicted for various stability conditions, 1 km downwind of a point source, by the observations of Figure 2 and the Islitser relation (column a), by the Pasquill nomograms for smooth land (column b), and by the adjusted Pasquill nomograms for rough land (column c).

Stability conditions	z/L	(a)		Stability class	(b)		(c)
		σ_{θ_6} , deg	σ_y over water, m		σ_y over land Smooth		Rough
Slightly stable	+0.03	3.5	50	E	50		80
Neutral	0.0	3.8	54	D	70		112
Slightly unstable	-0.03	4.2	60	C	105		168
Significantly unstable	-0.10	5.0	71	B-C	130		210

the table also lists the dispersion predicted by the well-known Pasquill nomograms^(4,12) for what are estimated to be similar conditions of stratification. Since the Pasquill curves correspond to quite smooth terrain (roughness length $z_0 \simeq 1$ or 2 cm, characterizing "open downland"), the last column adds plume statistics to be expected over a more realistic site ($z_0 = 15$ cm), the extrapolation carried out along the lines outlined in the last section of this paper.

Although there may be some reservations attached to the comparative plume estimates given here for other than neutral conditions (due in part to the somewhat arbitrary association of particular z/L magnitudes with subjectively defined Pasquill stability classes), Table 1 suggests that in stable conditions, dispersion over open water can be considered to be similar to that found over smooth grasslands in comparably steady mean winds. For z/L = +0.03 Table 1 shows identical σ_y values for smooth land and water surfaces, a similarity which can be expected to obtain in stronger stabilities as well since in these circumstances, turbulent structure becomes decoupled from the characteristics of the lower boundary. Once stratification of the

lower atmosphere crosses neutrality and becomes unstable, lateral dispersion rates over both smooth and rough land increase rapidly, while that over open water lags behind. In the strongest instability observed during the present trials ($z/L \simeq -0.15$ for a one-hour average, occurring with air 1.5° colder than the water and a wind speed of 6.4 m sec^{-1}), the recorded value of $\sigma_{\theta_6} = 5^\circ$ indicates $\sigma_y = 70 \text{ m}$ at the 1-km range, this amounting to only a 40 percent increase over the plume dimension in neutral conditions. In contrast, dispersion in similar instabilities over land more than doubles the plume widths observed in neutral conditions.

A Theoretical Generalization and Design for Experiment

With $\bar{u} \simeq \bar{s}$, both sides of Eq. (3) can be divided by the scale velocity u_* to obtain an expression in terms of the normalized lateral velocity fluctuation ratio and a surface drag coefficient:

$$\sigma_v/u_* = (\bar{u}/u_*) \sin \sigma_\theta \simeq (C_d)^{-1/2} \sigma_\theta, \quad (4)$$

where the coefficient $C_d \equiv (u_*/\bar{u})^2$ and the small angle approximation has been used to write σ_θ for $\sin \sigma_\theta$. Upon solving for σ_θ and substituting the result in a form of the Isiltzer dispersion relation, we obtain

$$\sigma_y/x \simeq A (\sigma_v/u_*) (C_d)^{1/2}, \quad (5)$$

for which Isiltzer's observations over land would indicate $A = 1/1.23 \simeq 0.8$ when the lateral fluctuation measurements are averaged over 5 sec.

It remains to be seen whether or not a constant fluctuation averaging time will be appropriate for dispersion estimates over a useful range of downstream distances over water. Alternatively, Eq. (5) can be written in terms of pollutant travel time $T = x/\bar{u}$ and a data averaging interval $t = T/\beta$, where $\beta \simeq 4$ is the Lagrangian-Eulerian scale factor [Hay and Pasquill⁽³⁾]; the expression reduces to

$$\sigma_y(T) = 4 t \sigma_{vt}, \quad (6)$$

which is appropriate to the plume width attained at time T after release.

But over open water the non-dimensional parameters of Eq. (5), σ_v/u_* and C_d , are functions of stability alone (though C_d may exhibit a weak dependence upon wind speed). This form therefore lends itself to experimental testing through a series of simultaneous dispersion and turbulent wind fluctuation measurements over water in order to determine values of these parameters appropriate to a useful range of data-averaging intervals, dispersion distances downstream, and atmospheric stability conditions. The outwardly simpler form in Eq. (6) could prove useful for routine monitoring of potential plume dispersal conditions downwind.

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INSTRUMENTATION FOR ATMOSPHERIC BOUNDARY LAYER STUDIES: THE WHAT SYSTEM

Paul Frenzen

A semiautomatic system for measuring vertical distributions of wind and temperature in the lower atmosphere with improved resolution and accuracy is being developed at this Laboratory. Figure 1 shows the principal components, one of a pair of digitized theodolites plus associated equipment to record the data in computer compatible form. Balloons being tracked for wind measurements will also carry transistorized radiosondes to measure the air temperature continually during ascents. This equipment for making successive measurements of wind, height, and temperature has been termed "the WHAT system." At this writing, the component hardware is undergoing bench trials; flight tests will begin shortly.

Theodolites used are of the standard U.S. Weather Bureau pattern (Warren-Knight Company, Model 85), except for installation of optical shaft encoders (Optisyn, DRC Model 29), one of which is visible in the figure. These transducers are directly connected to the principal elevation and azimuth axes of each instrument through non-backlash, metal bellows couplings. Note that, with this arrangement, the accuracy of the digitized record is not disturbed by disengaging the vernier controls during a run. Further, the shaft encoders do not interfere with the use of the theodolites in the usual, nonrecording mode. A cable between the two instruments permits them to be operated up to 1/2 km apart on a baseline for double-theodolite pilot balloon observations.

Each of the four shaft encoders produces one pulse for each 0.036° change in theodolite orientation (10,000 pulses per complete revolution). In the control unit, individual backwards-forwards registers maintain continuous digital records of the four angles. At preset intervals (typically one second, but adjustable over a wide range of times), values of the registers are transferred to the incremental tape recorder shown. A fifth register, periodically interrogated in similar fashion, records the radiosonde temperature signal detected and demodulated by the small receiver seen on the right in the

FIG. 1.--The WHAT system, a semiautomatic pilot balloon tracking system for obtaining measurements of vertical distributions of wind and temperature. The second theodolite (not shown) is located 600 m away.

photograph.

In one mode of operation planned, the computer program will generate locally averaged balloon locations from the continuous tape record by periodically summing the angles observed during 5 sec, calculating the mean x , y , z coordinates implied, and assigning these positions to the central observation times. With the balloon rising 3 m/sec, observations collected in this way at intervals of 20 sec between central times (discarding the intervening 15 sec of data) will produce winds averaged over layers 15 m thick, the central heights of the layers being separated by 60 m. Of considerable importance is the fact that this averaging will smooth the effects of operator tracking errors. Even without being required to interrupt his view of the balloon periodically in order to read elevation and azimuth angles visually, a theodolite operator will be hard-pressed to keep his instrument's cross-hairs as close to the center of the target as the ± 1 count digital accuracy of the WHAT system might be considered to justify; the equivalent resolution angle of $\pm 0.036^\circ$ intercepts 1 m at a range of 1600 m. Temperature data can either be collected over similar intervals to represent averages of the 15-m layer, or single values observed at the central times can be extracted to approximate (within the limitations of the temperature sensor response) "instantaneous" measurements representative of conditions at the central elevation.

The principal advantage of the WHAT system will be the greatly increased facility with which double-theodolite pilot balloon data can be reduced. Closely spaced balloon releases designed to investigate developing boundary layer phenomena will be feasible; moreover, the addition of simultaneous temperature measurements can be expected to contribute significantly to quantitative interpretation of the observations. In an alternative field experimental arrangement, the WHAT system can be used to track constant-level balloons to obtain Lagrangian dispersion statistics aloft. The system is fully portable and can be operated from storage batteries in the field.

The control unit and incremental tape recorder circuitry have been designed and constructed by L. L. Prucha of the Electronics Division. Paul Hess of the Atmospheric Physics Section contributed the computer program.

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SOME IMPROVEMENTS IN DESIGN OF A 403-MEGAHERTZ RADIOSONDE

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Introduction

An inexpensive, solid state 403-MHz radiosonde transmitter for temperature measurements originally designed at the Atmospheric Environmental Service of Canada (AESC)^{**} has been modified to increase useful range and facilitate mass production.

The original unit was constructed on a small fiberglass board using point-to-point wiring; the modified version uses a printed circuit board of similar dimensions designed to allow fast construction by nontechnical personnel. Parts cost for the modified unit is under \$5.00 in quantities of 100 (see parts list, Appendix A). It weighs approximately 25 g without a battery and about 65 g with a 9-V battery of the 216 format. The transmitter has a range of at least 10 km using an omnidirectional receiving antenna.

Modifications

Tests performed with the sonde provided by the AESC led to the following modifications (see circuit diagram, Figure 1):

- A. The trim inductor, L_2 , was removed along with the 250-pF antenna coupling capacitor, C_2 . These changes allowed the tank coil to be reduced from two turns hand wired on the original board to a one-turn coil of approximately 1-cm diameter printed on the new circuit board, while also realizing an improved impedance match to the end-fed, half-wave antenna.

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FIG. 1.--Circuit diagrams of original and modified 403-MHz radiosondes.

B. To simplify construction, an air variable capacitor (1.7 to 11 pF) was chosen in preference to the special pressure type used by AESC. Although the use of only a single-turn tank coil makes the air variable capacitor more difficult to tune, it is an inexpensive, commercially available component which performs satisfactorily

in the present application.

- C. To reduce errors in the temperature sensing circuit, a low temperature coefficient, polycarbonate capacitor ($0.01 \mu\text{F}$) was used in place of the original ceramic capacitor. The unijunction transistor remains somewhat temperature sensitive, and to minimize errors resulting from its temperature drift, the circuit will be packaged in a light-weight polystyrene insulating container.
- D. The current drawn by the circuit (about 25 mA) is sufficiently high that choice of a suitable battery is relatively critical. A significant improvement in performance results when alkaline cells are substituted for carbon-zinc batteries: battery voltage drop is lower in magnitude and largely confined to the first few minutes of operation. After ten minutes of operation, errors introduced as a result of battery performance are of the order 0.2°C using alkaline cells, in comparison to about 0.6°C in the case of carbon-zinc. Units will therefore be preconditioned before being used for radiosonde flights.

The Choice of Thermistor

The requirement for rapid response to changes in atmospheric temperature necessitates the use of a physically small sensor. One of the thermistors specified in the AESC design is a glass-coated bead 1 mm in diameter, compatible with the present response requirement. However, the dissipation constant of this thermistor is $0.7 \text{ mW}/^\circ\text{C}$ in still air, so that in the present circuit a temperature error of up to 0.5°C may occur when the thermistor resistance is $50 \text{ k}\Omega$. To minimize resistive heating errors, specific thermistors should be chosen (of the same physical dimensions) so that the resistance seen by the circuit is of the order $100 \text{ k}\Omega$. For surface temperatures of about 30°C , thermistor type GA51J1 is suggested, while type GA45J1 is better suited to colder operations (ground level temperature about 15°C).

Thermistors are normally coated with white epoxy paint to reduce the effect of solar heating. Depending upon the magnitude of the solar radiation, errors of up to 0.2°C might still occur.

Performance Tests

Laboratory tests of various modifications to the radiosonde transmitter were performed using an external field strength meter and a Metrodata MR17B receiver. In addition to inserting an attenuator ($\times 10$) in the receiving antenna feed line, the rf gain of the receiver was reduced to prevent overloading. The receiver and field strength meter were calibrated using a Marconi calibrated signal generator to simulate the signal at the receiving antenna. Table 1 summarizes the relative performances of various stages of modification, with the final version producing about sixteen times more effective radiated power than the original. Note that all tests used the same 31.1 cm (12.25 in), end-fed half-wave antenna.

TABLE 1. Laboratory Tests of the Radiosonde Transmitter.

Circuit	<u>Approximate relative field strength</u>	
	As metered, arbitrary units	Equivalent output of calibrator, mV
Original circuit	20	4.5
Original w/o 250 pF antenna coupling cap	30	6.8
Original with one-turn tank, w/o 250 pF and w/o trim coil; antenna center tapped on tank	35	8.8
Same as above, but antenna tapped at top of tank	50	16
New layout, one-turn tank, center tapped	38	9.4
As above, with antenna on top of tank	54	22

Range tests were conducted with the sonde suspended from a tethered balloon flown at various locations. The receiver and a quarter wavelength ground plane vertical antenna were mounted on a tower approximately 45 m high to allow "line-of-sight" propagation measurements without requiring long lines on the tethered balloon. The results presented in Table 2 indicate that the sonde is useful to distances of at least 10 km when used in conjunction with an omnidirectional receiving antenna and a Metrodata receiver. This useful range could be extended by employing a directional antenna at the receiver.

TABLE 2.

Approximate range, km	Signal strengths, S units	Remarks
3.	80	Sonde handheld
7.5	68	Sonde 30 m high on balloon
10.	55	Sonde handheld, noise ≈ 40 S units
13.	55	Sonde 100 m on balloon, much fading, noise ≈ 50 S units

Field trials of the present radiosonde have indicated that usable signals are received for at least 20 min after launch, even though temperature errors may be quite large at this stage. With balloon ascent rates of about 3 m sec^{-1} , it appears that the system is suitable for sounding of the lowest 3 km of the atmosphere.

APPENDIX A:

Parts List for 403-MHz Radiosonde

Number required	Cost	
	Small quantity	100 or more
<u>Resistors, all 1/4 W, carbon</u>		
1 75 Ω	\$0.10	\$0.07
1 220 Ω	0.10	0.07
1 2.2 k Ω	0.10	0.07
1 5.6 k Ω	0.10	0.07
1 8.2 k Ω	0.10	0.07
<u>Capacitors</u>		
1 2 pF Silvered mica: Elmenco	0.11	0.10
2 56 pF Silvered mica: Elmenco	0.24	0.10
1 0.01 μ F Polycarbonate: CDE, 50 Vdc	0.51	0.25
1 0.1 μ F Disc ceramic: Hypercon, 16 Vdc	0.15	0.11
1 1.7-11 pF Air variable: Johnson type T187-0106-005	0.70	0.58
<u>Other components</u>		
1 Thermistor, Fenwell type GA45J1 or GA51J1	2.67	1.74
1 2N3563	0.29	0.18
1 2N4582	0.97	0.65
1 Battery clip	0.30	0.10
1 Printed circuit board, approximately	1.00	0.50
1 Antenna, 31.1 cm	--	--
TOTALS	\$7.44	\$4.66

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