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PREPARING METEOROLOGICAL DATA FOR USE IN ROUTINE DISPERSION CALCULATIONS - WORKGROUP SUMMARY REPORT

Air Resources Laboratory
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PREFACE

During June and August of 1982, a discussion workgroup of scientists of the Meteorology Laboratory was convened for the purpose of outlining the structure and design of a meteorological preprocessor for routine use is air quality simulation of dispersion. The participants in the workgroup were: Dr. Francis Binkowski, Dr. Gary Briggs, Dr. John Clarke, Dr. Kenneth Demerjian, Dr. Robert Lamb, Mr. William Petersen, Dr. Jack Shreffler, and Mr. D. Bruce Turner. During the following months, as the workgroup chairman, I drafted a manuscript outlining the suggestions and recommendations. I am indebted to these colleagues who have patiently offered suggestions or criticism on various parts of the manuscript. Despite everyone's assistance, however, some errors and inadequacies no doubt exist, which of course are solely my responsibility.

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Abstract. The conclusions of a discussion workgroup are presented outlining methods for preparing meteorological data for routine use in air quality simulation of dispersion. The goal of the workgroup was to initially accommodate Gaussian plume modeling techniques, and to expand the meteorological variable list, as needed in the future, to accommodate other dispersion estimation techniques. Methods are suggested for estimating the vertical profiles of wind velocity, temperature, and the variances of the vertical and lateral wind speed fluctuations. Procedures are suggested for estimating the mixing height and the surface layer scaling parameters, including the Monin-Obukhov stability length. Coupled with near-surface measurements from a fully instrumented low-level meteorological tower, the winds, turbulence intensities, and temperatures are estimated using empirical formulations of the vertical profiles of these variables, defined in terms of mixing height and stability. The simplistic set of methods outlined offers specific ideas and suggestions to focus future deliberations on the critical concepts which need to be examined and evaluated.

1. INTRODUCTION

Most of the early attempts to estimate routinely the dispersion of air pollutants were based on the Gaussian-plume model, for example see Sutton (1953), Slade (1968), and Stern (1970). For routine studies, it was recognized that the only relevant measurements of wind fluctuations likely to be available would be those contained in conventional traces of the horizontal wind direction. Pasquill (1961) gave simple rules for obtaining the lateral spread based on wind-direction trace data. As measurements of the vertical wind-direction fluctuations would not be available routinely, Pasquill suggested the important effects (on turbulence) of thermal stratification in the lower atmosphere be represented in broad categories of stability, defined in terms of routine meteorological data. Hence, by judicious and clever manipulation of available meteorological data, the early dispersion models could be implemented using the hourly surface weather observations and the twice-daily upper-air observations collected by the National Weather Service.

These models estimated dispersion for each hour using hourly values of,

- o the surface wind speed and direction,
- o the ambient air temperature,
- o the Pasquill stability category, and
- o the mixing height.

Whether stated explicitly or not, most of the models assumed the wind direction was constant with height. Some of the models made provision for the wind speed to vary with height using a power-law approximation. The power-law exponent was specified as a function of Pasquill stability category. When temperature gradients were needed, as in the case of estimating plume rise during stable stratification, values were assumed for each Pasquill category. It was further assumed that the dispersive characteristics of the atmosphere were vertically and horizontally homogeneous within the mixed layer.

Much of the work on the characteristics of dispersion has been concentrated on the lower portions of the planetary boundary layer (PBL). However, to an increasing degree, more studies are being conducted of the upper regions of the PBL. These observations indicate that the structure of PBL turbulence (and thereby the mean distributions of velocity, temperature, and humidity) is strongly influenced by the surface conditions (McBean et al., 1979; Nieuwstadt and Van Dop, 1982). Practical methods are beginning to be developed for characterizing the profiles of wind, temperature, and turbulence within the planetary boundary layer. Obviously, these methods will improve as new data are acquired and analyzed. In the next decade, measurement methods may develop to the point that we can routinely measure the structure of the PBL through the use of Doppler-acoustic radars. Already, the capabilities of Doppler-acoustic radars to measure the vertical profiles of wind velocity have been reported by Hall et al. (1975), Balser et al. (1976), and Kaimal and Haugen (1977). More recently, Balser and Netterville (1981) discussed methods for further processing the received acoustic signals to produce measurements of the standard deviations of the wind direction fluctuations.

The purpose of the following discussion is to outline a set of methods for processing meteorological data to specify routinely,

- o the profile of the variance of the vertical wind-speed fluctuations,
- o the profile of the variance of the lateral wind-speed fluctuations,
- o the profile of wind velocity,
- o the profile of temperature,
- o the mixing height, and
- o the atmospheric stability.

Availability of data as listed above on an hourly basis has important implications with respect to dispersion modeling. The variances of the wind-speed fluctuations can be used to characterize the sigmas in a Gaussian-plume model (Draxler, 1976 and Irwin, 1979; 1983). The data listed would allow the improved algorithms to be used for estimating plume rise (Briggs, 1975) and wind direction shear effects (Pasquill, 1976), which to date have largely been ignored in routine modeling studies of dispersion. Also, Monte-Carlo particle trajectory models of dispersion or grid models using eddy diffusivities could be driven using these data (see Hanna, 1982).

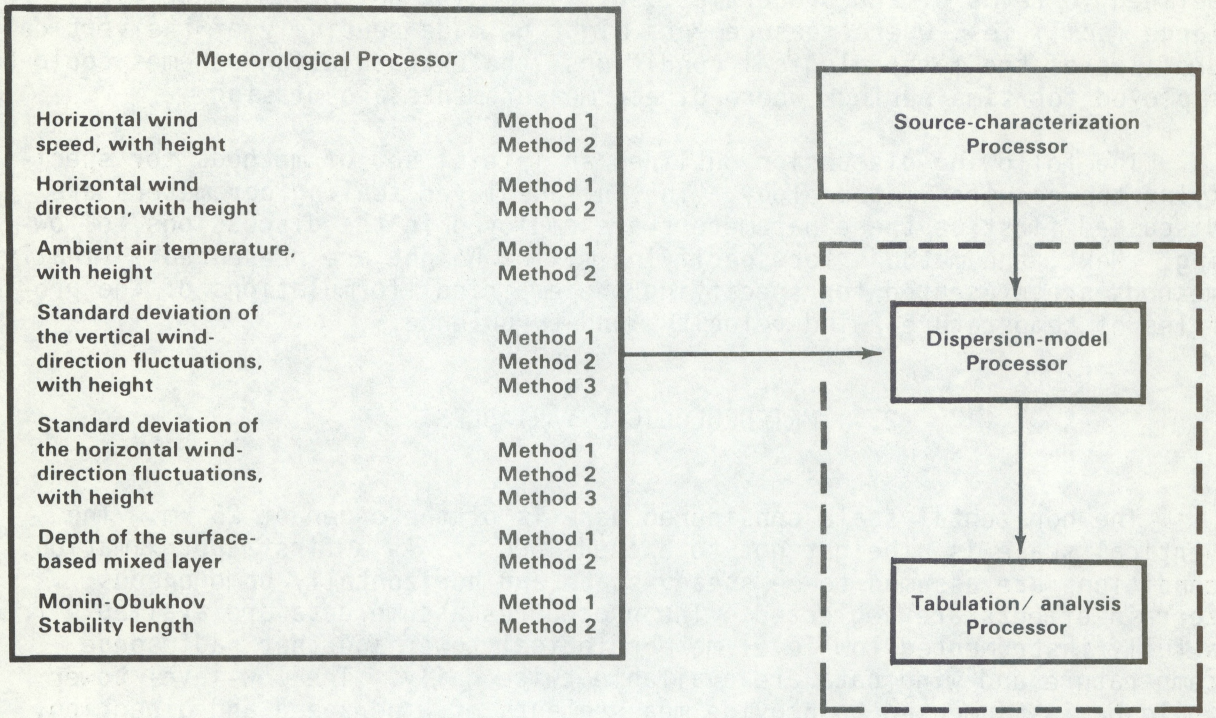

Figure 1. A depiction of the processors employed in formulating concentration estimates. Optimally, the meteorological processor has several methods to derive the required variables, the number depicted is for illustration purposes only. Many of the routine air-quality dispersion models combine the functions of the estimation of the dispersion and the tabulation and analysis of the concentration estimates.

In the following discussion, the model formulation of concentration estimates is viewed as a combination of several separate processors (Fig. 1). The dispersion model characterizes the transport and dispersion of the pollutants based on the information it receives about the meteorological conditions and emission characteristics, which are typically specified hourly. The meteorological and emissions processors are referred to as preprocessors, as their tasks need to be accomplished before the dispersion model processor can accomplish its task. If the meteorological and emissions data are measured directly, these processors do little more than format the data for input to the dispersion model. If the required input data are not measured, then procedures are employed to derive or estimate the required input from other available information. As the quality and completeness of the meteorological data will vary from one application to the next, it is desirable to provide alternate methods for specifying each of the input variables to the dispersion model. For analyses involving an existing source, having only a limited meteorological measurement program onsite, the winds and turbulence intensities aloft might be estimated using empirical formulations of the vertical profiles of these variables coupled with near-surface measurements of wind velocity and turbulence intensity. The vertical profiles, derived either from field data or from numerical simulations of the PBL, might be defined in terms of PBL properties, such as mixing height and stability. For large facilities, where measurements might be made routinely of the vertical profiles of the meteorological conditions, the extrapolation schemes could be employed for time periods where direct measurements are missing.

The following discussion outlines an initial set of methods for specifying the required meteorology. The surface layer scaling parameters are discussed first as these parameters are employed in the discussions following. Next, the methods for specifying mixing height are presented. Finally, methods are presented for specifying the empirical formulations of the profiles of temperature, wind velocity, and turbulence.

2. METEOROLOGICAL VARIABLES

The horizontal scale considered here is of the order of 25 km. The vertical scale is a height not to exceed 4000 m. As a first approximation, conditions are assumed to be steady-state and horizontally homogeneous; terrain effects are neglected. The procedures assume data are available from a fully instrumented low-level meteorological tower and that radiosonde temperature and wind data are available twice daily. The low-level tower would be instrumented to provide measurements of wind speed and direction, temperature and vertical temperature difference, and the vertical and horizontal standard deviations of the wind direction (Hoffnagle et al., 1981). The diurnal variations in the vertical profiles of wind velocity and temperature are strongest in the first few thousand meters above the ground in the atmosphere. These conclusions are based on intensive field studies where observations of the vertical profiles were available as frequently as every 1 to 3 hours. Such data are not routine. At best, one might hope to have data available routinely every 12 hours. From these, hourly profiles could be constructed in a crude manner using linear interpolation. The procedures outlined below attempt to adjust these interpolated profiles to be in better accord with the observed surface conditions.

2.1 Surface Layer Scaling Parameters (z_0, $u*$, and L)

2.1.1 Method for estimating z_0

The problems associated with determining the roughness length, z_0, for a particular location are discussed by Wieringa (1973, 1980). Even though the mean wind profile in near-neutral conditions can be represented by the well known logarithmic wind profile, practical problems associated with variations in land-use and ground-cover argue against wind-profile analysis in order to determine z_0.

TABLE 1. Terrain classification adapted by Wieringa from Davenport (1960) in terms of aerodynamic roughness length z_0 (m).

Class	Short terrain description	z_0(m)
1	Open sea, fetch at least 5 km	0.0002
2	Mud flats, snow; no vegetation, no obstacles	0.0050
3	Open flat terrain; grass, few isolated obstacles	0.030
4	Low crops; occasional large obstacles, $x/h > 20$	0.10
5	High crops; scattered obstacles, $15 < x/h < 20$	0.25
6	Parkland, bushes; numerous obstacles, $x/h \sim 10$	0.50
7	Regular large obstacle coverage (suburb, forest)	(1.0)
8	City center with high- and low-rise buildings	(uncertain)

Notes: Here x is a typical upwind obstacle distance and h the height of the corresponding major obstacles. Class 8 is theoretically intractable within the framework of boundary layer meteorology and can better be modeled in a wind tunnel. For simple modeling applications it may be sufficient to use only classes 1, 3, 5, 7, and perhaps 8.

In areas with a homogeneous surface, the roughness length can be estimated using the terrain descriptions adapted by Wieringa from Davenport (1960), Table 1. If the surface is not homogeneous, Van Dop et al. (1982) recommend the roughness length be estimated by computing an average drag coefficient, C_d, for the area as,

$$C_d = f_1 C_{d1} + f_2 C_{d2} + f_3 C_{d3}, \qquad (1)$$

where the weighting factors f_1, f_2, and f_3 are 0.85, 0.125, and 0.025 respectively. C_{d1} denotes the drag coefficient of the dominant terrain type, C_{d2} is the next important, and C_{d3} is the least extensive terrain type. The values of C_{d1}, C_{d2}, and C_{d3} are computed using the z_o values listed in Table 1 and the relationship that the 10-m drag coefficient, $C_d(10) = (k/\ln(10/z_o))^2$, where $k = 0.41$ is the von Karmen constant. The value of C_d determined using (1) is then converted to z_o from, $z_o = 10/\exp(k/C_d^{0.5})$.

2.1.2 Method 1 for estimating u* and L

Using surface-layer similarity relationships, the surface friction velocity, u*, and the Monin-Obukhov (M-O) length, L, can be determined using temperature difference and wind speed measurements from a low-level meteorological tower (Irwin and Binkowski, 1981; Wang, 1981) as,

$$z/L = (\phi_m^2/\phi_h)\, Ri,$$

$$z/L = k\, Ri_B F^2/G,$$

where ϕ_m and ϕ_h are the empirical nondimensional functions for wind shear and temperature gradient, and F and G represent the integral forms of the empirical flux-profile relationships for wind shear and temperature gradient, respectively. Ri and Ri_B are the gradient and bulk Richardson numbers defined using the low-level tower measurements of wind speed and temperature. Estimates of L from such measurements can be expected to be within a factor of two, which is sufficient accuracy for most purposes.

2.1.3 Method 2 for estimating u_* and L (daytime)

The flux-gradient relationships employed in the above procedure to estimate u_* and L are applicable in steady-state, horizontally homogeneous conditions. The low-level tower data should satisfy the joint constraints of being high enough to be above the influence of the individual surface roughness elements and yet low enough to be within the surface constant flux layer. Routinely satisfying these requirements is difficult. Hicks (1983) suggests less extensive use of the flux-gradient relationships can be accomplished through the use of Dyer's (1965) results.

The sensible heat flux, H, is estimated using the free-convection approximation,

$$H/(\rho c_p) = C\ (\ \theta_1 - \theta_2\ )^{3/2}, \qquad (2)$$

where c_p is the specific heat capacity of dry air at constant pressure, $C = H_*(g/\theta_1)^{1/2} z_1 z_2/(z_2-z_1)^{3/2}$, H_* is a dimensionless empirical constant (1.32 ± 0.06), g is gravity, θ is potential temperature, and z is measurement height. The subscripts 1 and 2, refer to the lower and upper measurement levels respectively. With z_1 = 10 m and z_2 = 30 m, C is approximately 0.80. Dyer (1965) tested (2) using data collected at 1 and 4 m and found excellent determination of the sensible heat flux provided z/L was less than -0.2, where $\bar{z}=(z_1 z_2)^{1/2}$. Hicks suggests that for cases when $-0.2<z/L<0$, the errors arising from inappropriate application of (2) are likely to have little practical consequence. The values for u_* and L are determined by iteration using,

$$u_*/u = k/\ln\langle(\bar{z}-d)/\bar{z}\rangle - 0.085/L,$$

$$L = -u_*^3\ \theta_1/\langle kg(H/\rho c_p)\rangle,$$

where d is the displacement height and u is the horizontal wind speed near the surface.

2.1.4 Method 3 for estimating u_* and L

Where the primary meterological data available are the standard weather observations, the values of L and u_* can be estimated following procedures discussed by Holtslag and Van Ulden (1982a,b). The procedures parameterize the surface energy fluxes in terms of cloud cover, time of day, near-surface wind speed, air temperature, and surface roughness. M-O similarity relationships are then used to estimate L and u_* as continuous variables of the parameterized surface energy fluxes. Over extensive areas of water, L and u_* can be estimated using procedures outlined by Nieuwstadt (1977) and Van Dop et al. (1982). Over water, the heat flux and friction velocity are parameterized in terms of 10-m wind speed, 10-m air temperature, specific humidity at 10-m, and the water temperature. It is assumed that the air just above the water is saturated.

2.1.5 Method 4 for estimating L

Based on an analysis of the meteorological data collected during the Project Prairie Grass field experiment, Briggs (1982) provides empirical relationships relating L to the more easily measured parameters $L_n = -u^3/R_n^*$ and $L_s = -u^3/R_s^*$, where u is the horizontal wind speed near the surface. R_n^* and R_s^* are the net radiation and solar radiation times $g/c_p\rho T$, hence L_n and L_s have units of length. During nighttime L_n is used to estimate L and during daytime either L_s or L_n can be used. As a rule of thumb, Briggs suggests u be measured at a height near or less than the minimum value of L.

2.2 Mixing Height.

For dispersion modeling purposes, the mixing height defines the layer above the surface through which pollutants are routinely mixed. In a study of the determination of mixing height during the daytime by means of an instrumented aircraft, McCaldin and Sholtes (1970) compared three definitions of mixing height -- temperature (height above ground at which the temperature gradient first became isothermal), turbulence (height at which accelerometer amplitude decreased to 50 % of the mean amplitude in the mixing zone), and concentration of suspended particulates of size 0.3 μm or greater (height at which concentration dropped to 5 % of full scale, 2 particles per cubic cm.). Based on an analysis of 145 soundings taken near Waco Texas on 33 days during the period 8 December 1969 to 9 March 1970, the mean mixing heights as determined by the three methods were 731 m (temperature), 701 m (turbulence), and 762 m (aerosols). These and other similar studies demonstrating the strong correlation during the daytime between the temperature profile and the depth of the aerosol layer resulted in the practice of estimating the daytime mixing height using radiosonde data and the surface shelter temperatures, (Holzworth, 1964 and Hanna, 1969). During nighttime, the criteria are uncertain for defining the layer through which pollutants would be routinely mixed. Caughey et al. (1979) defined the top of the stable boundary layer as the height at which the turbulent heat flux fell to 5 % of its surface value. Typically this corresponds to the height of the low-level nocturnal wind-speed jet maximum (Thorpe and Guymer, 1977). Mahrt et al. (1982) argue that the height of the low-level wind maximum may reflect the influences of baroclinity and the history of the wind more than the distribution of turbulence. They suggest the mixed layer is the layer below which the local gradient Richardson number is less than a critical value (0.5 was chosen as their cutoff value). Analysis of results from numerical simulations of the nighttime stable boundary layer by Garrett and Brost (1981) and Zeman (1979) suggest a shallow mixed layer depth of about 3L. Based on observations from field studies, Mahrt et al. (1982) found the mixed layer depth to be 6L. For his numerical simulations of the stable boundary layer, Yamada (1979) assumes that the top of the mixed layer is the height where the turbulent heat flux first vanishes. Most of the field studies of the stable boundary layer involve near-cloudless nights. The determinations of the mixing height are likely complicated by the presence of clouds, weather fronts, and the influences of terrain and large bodies of water.

For routine determinations of the mixing height, there appears to be three nearly equivalent choices.

2.2.1 Method 1 for estimating mixing height

A procedure was outlined by Benkley and Schulman (1979), which involved the determination of two heights during the daytime. A convective height is determined allowing adiabatic temperature modification to the radiosonde temperature sounding. The procedure adjusts for temperature advection effects during the day between the morning and evening radiosondes, by a linear interpolation in time between soundings. A mechanical depth is determined as 90 u_m, where u_m is the wind speed measured at 10 m, centered-averaged for 3 hours around the time when the depth is computed. The greater of the two heights is selected as the mixing height. During nighttime, the depth of the mixed layer is set equal to the mechanical mixed depth. The mechanical mixed depth defined seemed to perform adequately in the comparisons reported by Benkley and Schulman. However, as defined, it is unlikely that the depth would ever be below 90 m, which appears to contradict the shallow depths of 3L and 6L reported elsewhere.

2.2.2 Method 2 for estimating mixing height

Garrett (1981) proposed a similar procedure to that suggested by Benkley and Schulman except he determined the daytime convective depth using the rate equations derived by Tennekes (1973) for a zero-order jump model coupled with a free-entrainment formula based on the work of Deardorff (1974).

2.2.3 Method 3 for estimating mixing height

Finally, Van Dop et al. (1981) employ the model suggested by Tennekes (1973) during the daytime and use the model suggested by Nieuwstadt (1981) during the nighttime. Nieuwstadt suggested the stable mixed layer depth be estimated as $0.3u_*/f\ (1 + 1.9Zi/L)^{-1}$, where f is the Coriolis parameter and Zi is the mixing height.

2.2.4 Method 4 for estimating mixing height (nighttime)

An alternative to the above schemes would be to estimate Zi during the nighttime as 3L for moderately stable conditions and 6L for very stable conditions. An empirical expression that accomplishes this and fits the 1973 Minnesota data reasonably well is $Zi/L = 6/\log_{10}L$(meters).

2.3 Temperature Profile

During nighttime, with clear skies and strong radiative heat loss at the surface, the surface-based temperature inversion, formed soon after sunset, deepens throughout the night. It is important to make a distinction between the mixed layer height, Zi, and the temperature inversion height, Hi. During early evening Zi may be higher than Hi, while in the early morning hours, the turbulence may be suppressed by the temperature structure such that Zi may be less than Hi. From analyses of monostatic Lidar data and tethered balloon temperature soundings carried out in the Po Valley in Northern Italy, Anfossi et al. (1974, 1976) found $Hi = A_i t^b$, where Hi is in meters, $A_i = 70$, $b = 0.5$, and t was time in hours after the surface shelter temperature attained its maximum value. A similar type of analysis was conducted by Godowitch and Ching (1980), using data collected near St. Louis, Missouri during 20 fairweather July-August evening experiments. Hi was determined from the helicopter temperature soundings as the height at which the temperature gradient first became zero or negative. Their results suggest that $A_i = 95$, which is in good agreement with Anfossi et al. (1974), especially considering the geographical and terrain differences.

Yamada (1979) suggests a simple empirical expression for the potential temperature structure within the inversion layer,

$$(\theta - \theta_i)/(\theta_i - \theta_s) = -(1 - Z'/Hi)^c, \qquad (3)$$

where the subscripts s and i refer to values at screen height (1.2 m) and inversion top, θ is the potential temperature, $Z' = z - z_s$, and a value of $c = 3$ seemed supportable by the Wangara data. Yamada cautions that the value of c might be site specific. If we assume that radiosonde data are available, we can construct a crude procedure for determining the temperature structure as a function of time.

2.3.1 Method for estimating temperature profile

We recommend Hi be determined by the method of Anfossi et al. (1974, 1976). For heights above Hi, use the radiosonde temperatures. For heights below 10 m, use the 10-m tower temperature data. For heights between 10 m and Hi, use (3).

During the daytime, when convective processes dominate, the potential temperature is typically assumed to be constant within the convectively mixed layer (Benkley and Schulman, 1979; Tennekes, 1973). The following procedure is suggested. For heights greater than Zi, the radiosonde temperatures would be used. For heights below Zi, we would assume conditions are well mixed and everywhere the potential temperature is that recorded on the 10-m tower. This approximation neglects the condition that occurs during strong heating, where the near-surface potential temperatures are warmer than values aloft at 100 to 200 m.

In the absence of strong heating or cooling at the surface or when the wind speeds are strong (6 m/s or more), the nighttime-stable model and the daytime-convective model of temperature structure will likely not be appropriate. As a first approximation, we suggest that the radiosonde temperatures be used with no modifications.

The temperature structures outlined assume conditions are near steady-state. In the review presented by McBean (1979), it was mentioned that very small $-Zi/L$ values (perhaps near 1) are sufficient to drive the boundary layer into a convective state. It would also appear, as discussed earlier, that within a stable boundary layer Zi/L values are generally between 2 and 6. We propose the neutral model be employed only for situations when $|Zi/L|$ is less than or equal to one.

2.4 Wind Velocity Profile

During the nighttime, with strong radiative heat losses ($Zi/L > 1$), the wind speed increases with height, for heights within the stable boundary layer. At some height, typically just above the stable mixed layer height, the winds may become supergeostrophic (Thorpe and Guymer, 1977). The degree to which the winds become supergeostrophic is dependent upon terrain-slope effects, baroclinic effects (thermal wind), and isallobaric wind effects. It is conceivable that the balance of forces might reverse the normal tendency for wind speeds to increase with height through the stable boundary layer.

During the daytime, when convective processes dominate ($-Zi/L > 1$), the wind speed and direction remain nearly uniform within the well-mixed convective layer. At the top of the convective layer, the wind rapidly adjusts to the upper flow conditions. This can result in a rapid shift in speed and direction at the top of the convective layer. As with the nighttime situation, terrain effects (differential heating resulting in upslope winds), baroclinic, and isallobaric wind effects can all combine to alter the typical wind profile described above.

Of all the states of the boundary layer, the neutral (barotropic or baroclinic) atmosphere is one of the most frequently discussed. Since conditions are assumed to be steady-state, the observations of the neutral boundary layer most often occur during periods of moderate winds, or periods with overcast skies. In theory, neutral conditions occur whenever transition is made between stable and unstable conditions, as at sunrise or sunset, but the neutral condition may be quite brief and would not be considered a steady-state condition.

2.4.1 Method for estimating wind velocity profile

As a first step towards the development of a more refined technique, we suggest the wind profile in the vertical be determined using the procedures outlined by Van Dop et al. (1981). In the surface layer the variation of wind speed is assumed to obey M-O similarity theory. The "free atmosphere" is assumed to be in near-geostrophic balance. At the intermediate heights, the wind profile is obtained by interpolation between the surface layer wind and the free atmosphere wind. It is assumed that the geostrophic wind varies linearly with height. The surface layer wind is determined using the 10-m tower data as,

$$u_L = u_{10}\, F(z/z_o,\ z/L)/F(10m/z_o,\ 10m/L),$$

where the subscripts refer to the 10-m wind speed data, and F is a stability function given by various authors, e.g. Businger et al. (1971), Paulson (1970), Nickerson and Smiley (1975), Benoit (1977). Define the x-axis to be aligned along the surface geostrophic wind direction. With ϕ as the angle between the 10-m wind direction and the 1000 mb height contours, the surface layer wind components are determined as, $u_s = u_L \cos \phi$ and $v_s = u_L \sin \phi$. The components of the wind are then determined as,

$$u = f_s\, u_s + f_g\, u_g,$$

$$v = f_s\, v_s + f_g\, v_g,$$

where $f_s + f_g = 1$ and f_s and f_g are weighting functions. The function f_s varies smoothly from a value of one, for heights within the surface layer, to zero, for heights above the mixed layer depth. Whether the surface layer is restricted to heights such the $z/Zi < 0.2$ and $z/L < 2$, as suggested by Van Dop et al. (1981), or to heights such that $z/Zi < 0.1$ and $z/L < 1$, as suggested by Nicholls and Readings (1979), will likely make little difference as f_s as defined by Van Dop et al. (1981) yields similar values for either specification of the surface layer.

2.5 Turbulence Profiles

During nighttime, when the atmosphere is stably-stratified, the vertical wind-speed fluctuations decrease in amplitude as a function of height, to near-zero values at Zi by definition (Caughey, 1982 and Yamada, 1979). We found that a reasonable fit to the 1973 Minnesota field data reported by Caughey et al. (1979), is of the form,

$$\sigma_w^2/u_*^2 = A_w\,(1 - z/Zi)^d,$$

where σ_w is the standard deviation of the vertical wind-speed fluctuations, with $A_w = 2.2$, and $d = 3/2$. Nieuwstadt's (1982) results for the Cabauw data collected using a 213 m meteorological tower are quite similar with $A_w = 2.4$ and $d = 3/2$.

For daytime convective conditions, Caughey (1982) notes that the observations of σ_w suggest a broad maximum centered at Zi/2 where σ_w^2 is $0.4w_*^2$ (w_* is the convective velocity scale). The numerical results of Deardorff (1974) suggest a lower height for the level of maximum variance at Zi/3.

Caughey (1982) reports that near Zi, water tank and numerical results indicate σ_w^2 is $0.1w_*^2$, in agreement with the atmospheric data. Above Zi, Caughey reports that σ_w^2 decreases to $0.01w_*^2$ near 1.5Zi.

The general consensus in the literature is that the standard deviation of the cross wind-speed fluctuations, σ_v, is poorly characterized using M-O similarity scaling, for example see Panofsky et al. (1977) and Binkowski (1979). The numerical results and atmospheric data suggest a slight maximum near 0.8H in σ_v^2; the average value over the entire profile for σ_v^2 is $0.4w_*^2$.

2.5.1 Method 1 for estimating σ_w profile

We suggest an empirical model for describing the stability dependence of the vertical profile of σ_w as, for $L < 0$ and $-Zi/L < 1$,

$$\sigma_w = \sigma_w^o,$$

for $L < 0$ and $-Zi/L > 1$,

$$\sigma_w = \sigma_w^o \left[1 + \langle (Zi/L + 1)/(Zi/L) \rangle \sin(\pi z/Zi)\right],$$

and for $L > 0$,

$$\sigma_w = \sigma_w^o (1 - z/Z^*)^{3/4},$$

where σ_w^o is the standard deviation of the vertical wind-speed fluctuations measured at 10 m and $Z^* = Zi + L$. When $L < 0$, σ_w is assumed to be no greater the σ_w^o for heights greater than Zi, and when $L > 0$, σ_w is assumed to be zero for heights greater than Z^*. During unstable conditions ($-Zi/L > 1$), the profile rapidly approaches the limit $2\,\sigma_w^o \sin(\pi z/Zi)$, which fits reasonably well the observations summarized by Caughey (1982). The Z^* scaling, for cases when $L > 0$, provides a reasonable fit to the 1973 Minnesota nighttime data and satisfies the neutral limit (when L approaches ∞) that σ_w is nearly independent of height.

2.5.2 Method 2 for estimating σ_w profile (daytime)

Based on analyses of data from several field studies, with careful avoidance of possible complex-variable contamination, Hicks (1983) suggests σ_w be estimated during the daytime, $L<0$, as,

$$\sigma_w = 1.1u_*(1 - 2z/L)^{1/3}, \quad \text{for } z/Zi < 0.1$$

$$\sigma_w = 0.65\, w_*, \quad \text{for } 0.1 < z/Zi < 1.0.$$

2.5.3 Method 1 for estimating σ_v profile

For the lateral turbulence component, σ_v, we assume that σ_v is equal to the measured value at 10 m at all heights for all values of L. If measurements of the standard deviation of the lateral wind direction fluctuations, σ_a, are available at 10 m, then σ_v is estimated as $\sigma_a U_z$, where U_z is the wind speed at 10 m.

2.5.4 Method 2 for estimating σ_v profile (daytime)

For estimating σ_v during the daytime, L<0, Hicks (1983) suggests,

$$\sigma_v = u_*(1.9 - 3.5z/L), \quad \text{for } -z/L < 0.3$$

$$\sigma_v = 3.0u_*, \quad \text{for } -z/L > 0.3.$$

The above characterizations of σ_w and σ_v are very crude and should be considered only as a first step from which to measure future development.

3.0 CONCLUSIONS

Several methods have been outlined providing estimates of the vertical profiles of wind velocity, temperature, and the variances of the vertical and horizontal wind-speed fluctuations. Procedures are also suggested for estimating the mixing height and the Monin-Obukhov stability length. Although initially the goal is to accommodate Gaussian-plume modeling techniques, the goal is to expand the meteorological variable list, as needed, in order to accommodate a wide range of dispersion estimation techniques. Initially, it appears feasible to encode all of the procedures into one computer program. As the system expands, with the addition of new procedures and variables considered, the inclusion of all procedures in one program might become impracticable. However, the overall system design would still provide a useful means for tracking the procedures and cataloging the relative performance and costs of the procedures. Organizing the information as described will facilitate decision-making on the selection of procedures for particular air quality analyses, and on the selection of research priorities.

To develop a meteorological processor, the procedures should be tested using readily available data. Tables 2, 3, and 4 summarize some of the available comparison results. Where results were extracted from available articles and reports, it was not always possible to complete the entries to the tables. Other statistical measures might have been selected, but those presented serve the purpose of summarizing the results known. In reviewing these results, it becomes clear that there are only a few comprehensive data sets available. Those field studies often cited are O'Neill, Prairie Grass, Kansas, Wangara, Minnesota, and Cabauw. Caution should be exercised in interpreting the comparison results listed in Tables 2, 3, and 4 since these data were often employed in the development of the procedures. Moreover, few of the methods for estimating the same variable have been compared using a common data set. Hence, the few comparison results presented allow some subjective assessment of the performance of the methods but they are insufficient to provide a basis for choosing between alternate methods. We recommend that a more complete set of comparison statistics be developed for each method using common data sets. It is anticipated that such comparison results will highlight the estimation methods having the least skill. Research can then be focused on these meteorological variables, and on developing methods to characterize the spatial variations in the variables.

TABLE 2. Summary of comparison results for estimates of the Monin-Obukhov length and the surface friction velocity. N is the number of values, N2 the number of comparisons within a factor of two, MFE is the mean fractional error computed as 2(P-O)/(P+O) where P is the estimate and O is the observation, RMSE is the root mean squared error, and r is the correlation coefficient.

Variable	Data set	Method	N	N2	MFE(%)	RMSE	r	Remarks
Monin-Obukhov length	Minnesota and Kansas Wang (1981)	1	43	43	21	63 m	0.96	Both daytime and nighttime cases. (Dyer-Hicks constants).
	Cabauw, Holtslag and Van Ulden (1982a)	3	999			0.67*	0.85*	Daytime cases. *Results cited for 1000 z_0/L, where z_0 = 0.2m.
	Cabauw, Holtslag and Van Ulden (1982b)	3	1643			50 m	0.85	Nighttime cases.
	Kansas, Briggs (1982)	4	65	52	19	22m	0.97	Daytime and nighttime cases. Results for 8 m wind speeds using R_s^* during daytime and R_n^* during nighttime.
Surface friction velocity	Minnesota and Kansas, Wang (1981)	1	31	31	11	0.05 cm/s	0.98	Daytime cases.
	Cabauw, Holtslag and Van Ulden (1982a)	3	999				0.99	Daytime cases.
	Cabauw, Holtslag and Van Ulden (1982b)	3	1643			0.90 cm/s	0.99	Nighttime cases.

TABLE 3. Summary of comparison results for estimates of the mixing height. N is the number of values, N2 the number of comparisons within a factor of two, MFE is the mean fractional error computed as 2(P-O)/(P+O) where P is the estimate and O is the observation, RMSE is the root mean squared error, and r is the correlation coefficient.

Variable	Data set	Method	N	N2	MFE(%)	RMSE	r	Remarks
Mixing Height	Wangara and O'Neill, Garrett (1981)	1				430 m	0.86	Daytime cases.
		2				250 m	0.98	
	Four southeastern U.S. stations, Garrett (1981)	1				552 m		Daytime cases.
		2				516 m		
	Kincaid, Benkley and Schulman (1979)	1	48		12		0.88	Daytime cases.
	Minnesota, Caughey et al. (1979)	1,2	7	6	22	46 m	0.97	Nighttime cases.
		3	7	3	-69	123 m	0.98	
		4	7	7	-14	54 m	0.97	
	Wangara*	1,2	60	19	85	113 m	0.71	Nighttime cases. *Values of L and u* from Melgarejo and Deardorff (1975) and Zi values (defined using Richardson number) from Mahrt et al. (1982).
		3	60	33	-39	61 m	0.76	
		4	60	46	15	54 m	0.77	

TABLE 4. Summary of comparison results for estimates of the standard deviations of the vertical and lateral wind speed fluctuations. N is the number of values, N2 the number of comparisons within a factor of two, MFE is the mean fractional error computed as 2(P-O)/(P+O) where P is the estimate and O is the observation, RMSE is the root mean squared error, and r is the correlation coefficient.

Variable	Data set	Method	N	N2	MFE(%)	RMSE	r	Remarks
Vertical wind-speed fluctations	Minnesota, Izumi and Caughey (1976)	1	66*	66	6.5	20 cm/s	0.81	Daytime cases. *Used observations taken at 32 m in forming estimates. Comparisons for 32 m data excluded from statistical analyses.
		2	77	77	-4.2	17 cm/s	0.91	
	Minnesota, Caughey et al. (1979)	1	8*	8	-3.3	2 cm/s	0.97	Nighttime cases. *Used observations taken at 32 m in estimates. Comparisons for 32 m data excluded from statistical analyses.
Lateral wind-speed fluctuations	Minnesota, Izumi and Caughey (1976)	1	66*	66	-4.5	33 cm/s	0.67	Daytime cases. *Used observations taken at 4 m in forming estimates. Comparisons for 4 m data excluded from statistical analyses.
		2	77	71	-38.4	45 cm/s	0.63	

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