COST Action 710
Preprocessing of Meteorological Data for Dispersion Modelling

Report of Working Group 3

Vertical profiles of wind, temperature and turbulence

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1 INTRODUCTION AND MOTIVATION

1.1 Background and aim of the work

COST 710 was set up in spring 1994 to act as a catalyst for the harmonisation process in Europe on meteorological preprocessing for atmospheric (dispersion) modelling. In COST-710 many organisations cooperate while carrying out their own research programme. COST stimulates the exchange of methods and data. The work presented here was not meant to be exhaustive since the resources available were very limited. The activities were split up into 4 working groups on 1) surface energy balance 2) boundary layer height, 3) vertical profiles and 4) complex terrain items. In this report we describe the results of the studies of working group 3.

For air pollution studies the basic parameters that should be known concern **the wind profile (wind speed and direction**, determining the transport process), **the degree of turbulence** near the surface (determining the mixing and dilution) and **the height of the boundary layer** or mixing layer (determining to what height pollutants may be carried up into the atmosphere). In addition, **the temperature profile** can be important for calculating the rise of hot plumes, in particular in estimating when plumes penetrate (fully or partially) the stable layer above the mixed layer and in estimating mixing layer depth. Most of the time the mixing layer is capped by a stable layer. The growth of the boundary layer height during daytime conditions is strongly determined by the temperature lapse rate in this inversion layer.
Profiles of wind, temperature and turbulence and the height of the boundary layer are not measured on a routine basis. Therefore, indirect methods are introduced to calculate these parameters. Such methods are generally based on concepts in which the heat, momentum and moisture fluxes at the surface play a central role. The profiles of temperature, wind and turbulence are all interrelated and dependent on atmospheric stability. This is shown qualitatively in Figure 1.
As well as being of interest in its own right, the mixing height also plays a role in determining the turbulence profiles. In this report we study the problem of determining the profiles of wind, tempertaure and turbulence under the assumption that the surface fluxes and the mixing height are known. The problem of estimating the latter quantities is considered by working group 1 (surfaces fluxes) and 2 (mixing height).

1.2 The scope of working group 3

We decided to focus on the profiles of wind, temperature and turbulence because these are the main profiles of interest in atmospheric dispersion.
More specifically we consider profiles of:
- wind speed and wind direction
- the turbulence parameters σ_v, σ_w and the Lagrangian time scale T_l.
- temperature

Figure 1 Qualitative presentation of atmospheric turbulence intensity, temperature profile, and the radiation during unstable, neutral and stable atmospheres.

Other parameters may be of interest (K_z is briefly considered), but will not be addressed in this report. For example, moisture profiles could be important in predicting the visibility of condensing plumes, but we have not considered them here. Also, for particle models the third moment in the turbulent fluctuations (the skewness) is often of importance, but will not be addressed in this paper. These higher moments (such as skewness and kurtosis) are reported by Sorbjan (1991), considering field experiments and by Nieuwstadt (1990) for Large Eddy Simulations. A recent extended discussion on higher moments has made by Du (1996). For most practical applications in short range dispersion problems these items are of minor importance.

1.3 Motivation of the work

Governments have taken many measures to reduce air pollution and to monitor air quality. In doing so, air quality standards are introduced, such as the maximum admissible hourly or daily mean concentrations during a year and the so-called high percentiles. These standards are evaluated by means of measurements or (e.g. for future emissions) by applying dispersion models. This can be done on several spatial scales: local, regional and continental scales. In the Netherlands all parties that intend to start (or significantly change) activities with an environmental impact are obliged to present an Environmental Impact Statement in which all environmental effects are described. In such reports the air pollution concentrations around stacks should also be given and compared to the standards.

In this respect there is an obvious need for adequate mathematical models and calculational tools that allow reliable estimates of concentrations. The use of modelling techniques is a strong tool to calculate the effects of different kinds of air pollutants. Local authorities also use models to set up most effective strategies for controlling air pollution problems on a local or regional scale. The effect of environmental measures should also be evaluated by mathematical dispersion models.

On the local scale (a few kilometers), individual sources have occasionally proven to cause large problems. The pathway of pollutants in the air on a local scale is presented in Figure 2. For estimating concentrations from such sources, dispersion models for stack emissions are necessary. Such models are being implemented on computers in which the transport and dispersion of air pollutants is described. This type of model describes physical processes influencing dispersion in the atmosphere and depends on a good understanding of these processes.

1.4 Important features of dispersion models

Dispersion theory started with G.I. Taylor’s analysis (1921), who described the behaviour of particles in homogeneous turbulence. This analysis proved to be very worthwhile and was taken as the basis for many recommendations. Cramer (1976), Draxler (1976) and Pasquill (1976) proposed pragmatic formulations based on this concept and fitted to measurements. These formulations appeared to be more reliable than others especially for the value of the lateral dispersion parameter σ_y (Irwin, 1983). However, more empirical formulations, which express σ_y and σ_z as functions of distance for each of a number of "stability categories", proposed by Pasquill (1961), Briggs (1973) or Singer and Smith (1966) became more popular, partly because they do not require turbulence data as input.

Figure 2 The pathway of pollutants in the air on a local scale.

On the other hand, formulations based on **the similarity theory**, giving direct relations between relevant meteorological parameters and values of the two determining dispersion parameters (the horizontal standard deviation σ_y and the vertical standard deviation σ_z of the plume shape - see Fig. 3), were developed somewhat later (see e.g. Weil and Brower, 1984; Berkowicz et al., 1985). However, their dependence on such relevant parameters as heat and momentum fluxes made them, in spite of their more sound theoretical basis, more difficult to be accepted and implemented so that most consultants were used to providing dispersion models with stability classes.

To verify the dispersion schemes, many field measurements have been used. We mention here the CONDORS field experiments at the 380 m meteo-mast near Boulder (Moninger et al., 1983; Eberhard et al., 1985 and Kaimal et al., 1986), the smelter experiments of Carras and Williams (1981) over large distances in Australia, and the Danish SF_6 measurements in Copenhagen (Gryning, 1981 and Gryning and Lyck, 1984). Measurements at large power plants were reported by Weil (1979) and by EPRI at the Kincaid and Bull Run power stations (EPRI, 1983a and 1983b). Also, in England some plume measurements around power plants were reported (Moore and Lee, 1981).

In the eighties a more physical approach led to models that did not use stability classification and simple schemes, but coupled dispersion directly to physically meaningful parameters. This understanding led to the **next generation of dispersion models**, resulting in well-known models such as the Danish OML model (Berkowicz et al., 1985), the British UK-ADMS (Carruthers et al., 1992) and the American HPDM model (Hanna and Chang, 1993). Other validated examples of models where σ_y and σ_z are continuous functions of atmospheric parameters are given by Briggs (1993a and 1993b).

Improved dispersion algorithms in advanced gaussian models calculate σ_y and σ_z mainly in two ways. First, parameters such as friction velocity (u_*), Monin-Obukhov length scale (L), convective velocity scale w_* and boundary layer height z_i are frequently used in continous functions to calculate σ_y and σ_z with fit parameters on the basis of dispersion experiments. Models that are based on these parameters are the OML model (Berkowicz et al., 1985), to some extend the UK-ADMS (Carruthers et al., 1992), the HPDM (Hanna and Chang, 1993), the Dutch Nationale Model STACKS (Erbrink, 1995) and the OPS model (Van Jaarsveld and De Leeuw, 1993). To come to a reliable calibration of these dispersion modules, good independent data sets are needed. Although much effort has been put into setting up well-documented dispersion experiments, e.g. Condors experiments (Briggs, 1993), Prairie grass experiment (Haugen, 1959), measurements at Cabauw (Van Duuren and Nieuwstadt, 1980) and others, there is still a strong need for more measurements. Secondly, some well-known schemes are developed to calculate σ_y and σ_z directly from the turbulence parameters σ_v, σ_w and related time scale T_l. Directly measured values of turbulence intensity and time scales are recommended as input parameters. The advantage is clear: no additional corrections for surface types (moisture content, albedo, roughness-length and corrections for application over water surfaces) are necessary. Instead of measured values, more or less theoretical schemes may be used to obtain values of σ_v, σ_w and T_l when measurements are absent. For these schemes it is necessary to know these surface parameters. The dependence on height of turbulence can easily be implemented to some extent. These features are implemented in (among others) the UK-ADMS and STACKS.

While all these models use **the gaussian plume concept** as a starting point with the concentrations calculated from the height of the plume axes and the determining parameters σ_y and σ_z (see Figure 3), other non-gaussian models became applicable because of the growing computating capacity of non-mainframe computers and later the more user-friendly workstations. The so-called Monte Carlo models, (or Lagrangian particle models) are capable of handling non-homogeneous turbulence. The models are based on the idea of describing the individual motions of many particles in terms of mean and turbulent behaviour. Most models of this type give solutions of the Langevin equation. **The profiles of wind speed, wind direction and relevant turbulence parameters** such as intensity, time scale and possibly higher moments (skewness, kurtosis) should be given **as functions of height**. Because little is known with enough accuracy about those profiles (except for very convective conditions over flat terrain), the output of those models is not very well verified (see. e.g. Wratt, 1987; Irwin and Paumier, 1990).

The work of Thomson (1984) belongs to the basics in the field. While gaussian models can be applied effectively only over relatively flat terrain, Monte Carlo models can be applied in non-uniform terrain as well as treating the more subtle non-gaussian aspects over flat terrain. Applications in coastal zones are presented by Flassak and Moussiopoulos (1992), Eppel et al. (1992). Valley dispersion is modelled successfully by Lange (1990) and Matamala and Pilinis (1991) in the Grand Canyon. Non-homogeneous turbulence above flat terrain is worked out by Baerentsen and Berkowicz (1984) and Brusasca et al. (1989).
A detailed overview is given by Zannetti (1990). The incorporation of buoyant sources was developed by Van Dop (1992), Beniston et al. (1990) and Hurley and Physik (1993) and in a simpler way by Anfossi et al. (1993).

Figure 3 The gaussian plume concept.

A further technique to calculate dispersion is the Large Eddy Simulation (LES). The basic idea of LES is to solve the Navier Stokes equations for the energy containing eddies in a grid of cells. While the Monte Carlo models need turbulence profiles as input, the large eddy models generate those profiles themselves, with only the geostrophic wind field and surface conditions as input. By considering the motions of particles in this framework, the dispersion is calculated. Because a large number of cells must be followed with a small time step, calculations are expensive and can not be done to obtain concentration statistics over long periods (e.g. a year). Examples of such calculations were recently presented by Nieuwstadt (1992), Nieuwstadt and De Valk (1987), Nieuwstadt and Bouwmans (1994) and Henn and Sykes (1992).

All these different models reflect the reality that atmospheric processes are very complicated. Each model can only handle a restricted subset of processes, depending on the purpose of the model and the available input parameters. This determines the model's applicability and usefulness.

Both the description of meteorology and dispersion have been improved considerably. In our opinion the improvement of meteorology is most urgent. In earlier days the stability of the atmosphere was estimated with Pasquill/Gifford/Turner schemes, but the use of more physically related parameters has led to considerable progress in dispersion modelling.

In the above we have seen that profiles of wind and turbulence play important roles in modelling air pollution using different types of models (gaussian, lagrangian or eulerian).

2 DIFFERENT FORMULATIONS FOR VERTICAL PROFILES

2.1 Criteria

The main criteria to select procedures for calculating vertical profiles are:

- general applicability (viz. not restricted to one location or one situation)
- input should be available from routine measurements
- preferably refereed (and accepted) concepts (e.g. similarity concepts)

With the similarity theory it is possible to calculate the profiles of wind speed and turbulence, which are crucial factors in modelling air pollution. Until similarity theory was developed, experimental power law profiles were used to obtain wind profiles and no useful concepts for turbulence profiles were available. The Monin-Obukhov theory became better-known by the exploration of its concepts in large field experiments, such as the Kansas (in 1968) and Prairie Grass (in Nebraska in 1957) experiments (see Haugen, 1959; Businger et al., 1971; Kaimal et al., 1972) and the Minnesota experiment (Kaimal et al., 1976). Although the Monin-Obukhov theory strictly applies to the surface layer (roughly 10% of height of the atmospheric boundary layer), it has also been applied to greater heights with reasonable success.
The core of the atmospheric mixing layer ($0.1h<z<0.9h$, with h the mixing height and z the height of measurement) is characterised by an intense vertical mixing during the daytime. Variables like potential temperature, wind velocity and moisture are nearly constant with height. Due to the earth's proximity, surface-layer ($z<0.1h$) characteristics are different. Wind speed diminishes and becomes zero at ground level. Vertical movements are attenuated by the obstruction of the ground. Hence, σ_w and T_l (in the vertical direction) decrease. Hence, for some parameters the formulation of the vertical profile differs between the surface layer and the mixed layer.

2.2 Wind profiles

2.2.1 Theoretical background

Let us first define some dimensionless numbers, which are of great importance in ABL modelling. Presenting dimensionless quantities has some important advantages over the original forms in which parameters are measured. Dimensionless combinations of variables reduce the number of dependent variables and are independent of the units used. Their use facilitates the comparison of observations from different places and often contributes to the discovery of simple functional relationships. Also, in handling wind tunnel models for instance it is important to use the same geometric and flow scalings as in reality. This can be achieved through the use of dimensionless numbers.
Height z can be made dimensionless by dividing it by z_0. However this is only useful near the ground, Instead we introduce a nondimensional form of the height z, valid for

describing the wind profile in the core of the boundary layer well above the roughness elements, where the wind profile is determined by friction and by the Coriolis force. Friction (momentum flux) is described by the friction velocity u_* and we obtain the non-dimensional height: zf/u_*. f is the Coriolis parameter, being defined as $2\Omega\sin(\phi_l)$. Since the ABL is driven at its top by the geostrophic wind (defined by the horizontal pressure gradient) the difference between u and U_g is determined by friction; hence, it is logical to scale u-U_g with u_* and $(u$-$U_g)/u_*$ is supposed to be a function of fz/u_*, for the x-component of the surface wind and in the neutrall case (no buoyancy force) we find:

$$\frac{u-U_g}{u_*} = F_x\left(\frac{fz}{u_*}\right) \tag{1}$$

For some function F_x close to the surface (in the surface layer) the Coriolis force is not relevant anymore but the surface roughness is. Height z can be made non-dimensional by dividing it by z_0: z/z_0, and again wind speed can be made non-dimensional as u/u_* we have:

$$\frac{u}{u_*} = f_x\left(\frac{z}{z_0}\right) \tag{2}$$

Differentiation of (2) gives:

$$\frac{\partial u}{\partial z} = \frac{u_*}{z_0}\frac{df_x}{d(z/z_0)} \tag{3}$$

Differentiation of the wind profile (1) gives:

$$\frac{\partial u}{\partial z} = f\frac{dF_x}{d(zf/u_*)} \tag{4}$$

Now we have two equations for the change in wind speed with height: one derived from the situation close to the surface and one describing the situation starting well above the roughness elements. In the region where the requirements for both equations should be fulfilled (far from the geostrophic wind level, $z \ll u_*/f$ and far from the surface ($z \gg z_0$)) the two differential equations should match. The double limit $z/z_0 \rightarrow \infty$ and $zf/u_* \rightarrow 0$ should be equal, but this is only possible when there is an asymptotic non-zero and finite number κ:

$$\frac{z}{u_*}\frac{\partial u}{\partial z} = \frac{1}{\kappa} \tag{5}$$

2.2.2 The surface layer

In (5) the von Karman constant (κ) appears. When this constant κ is moved to the left, this expression is called **the dimensionless wind profile**. This formula tells us that in a layer where z/z_0 is large enough to allow the influence of individual roughness elements to be neglected and zf/u_* is small enough to allow the Coriolis effect to be neglected (no directional windshear), the wind shear only depends on u_*! The von Karman constant κ is an experimental factor, and its value has given rise to a lot of discussions. Now most researchers accept a value close to 0.40. In this report κ=0.40 will be used.
Integration gives the logarithmic wind profile:

$$\frac{u}{u_*} = \frac{1}{\kappa}\ln\left(\frac{z}{z_0}\right) \qquad \textbf{(6)}$$

for neutral conditions. This expression can always be used in neutral conditions when the surface roughness is known and together with the friction velocity u_*.
If the surface heat flux is non-zero Monin-Obukhov theory predicts that the Monin-Obukhov length scale is a measure for the buoyancy influence, because this parameter is the only length scale containing the surface heat flux. The Monin-Obukhov length scale L_* is defined by:

$$L_* = -\frac{u_*^3 c_p \rho \theta_v}{\kappa g\ H} \qquad \textbf{(7)}$$

This assumption, which states that $\zeta = z/L_*$ is the non-dimensional height, leads to a generally accepted formulation for the wind speed profile:

$$\frac{\kappa z}{u_*}\frac{\partial u}{\partial z} = \phi_m(\zeta) \qquad \textbf{(8)}$$

If ζ=0 (L_*=infinite) then $\phi_m(0)=1$. To calculate $\phi_m(z/L_*)$, different more or less empirical functions have been proposed. Panofsky and Dutton (1984) give:

$$\begin{aligned} \textit{unstable}: &\quad \phi_m = (1-16z/L_*)^{-1/4} \\ \textit{stable}: &\quad \phi_m = 1+5z/L_* \end{aligned} \qquad \textbf{(9)}$$

Holtslag (1984) showed some improvements in calculating the wind speed profiles when 1/3 is used as exponent instead of 1/4. The constant 16 sometimes is different; Dyer and Bradley (1982) give a value of 28. Integration of (8) from the height where theoretically

$u(z)$=0, i.e. the roughness length z_0, yields the wind speed profile:

$$u(z) = \frac{u_*}{\kappa}[\ln\frac{z}{z_0} - \psi_m(\frac{z}{L_*})]$$
$$\textit{with}: \quad \psi_m(\frac{z}{L_*}) = \int_0^{z/L_*} [1-\phi_m(\zeta)]\frac{d\zeta}{\zeta} \tag{10}$$

with ψ_m=0 in neutral conditions. In this formula we neglect the term $\psi_m(z_0/L_*)$ - which is normally small - which should appear, strictly. These functions are widely used.
In unstable conditions -applying (10)- the final result of this integral is (Paulson, 1970):

$$\psi_m(z/L_*) = \ln\left[\frac{1+x^2}{2}\left(\frac{1+x}{2}\right)^2\right] - 2\ \arctan(x) + \frac{\pi}{2}$$
$$\textit{where}: \quad x = (1-16\frac{z}{L_*})^{1/4} \tag{11}$$

and in stable conditions:

$$\psi_m = -5\frac{z}{L_*} \tag{12}$$

As for the neutral result (6) these formulations are stricly valid only in the surface layer at heights much less than h. In practice they can work to higher heights, although not for example if 'low level jets' are present. For a discussion about these low level jets see e.g. Laude and Tetzlaff (1986) or Thorpe and Guymer (1977). The main characteristics of the low level jet are described by Arya (1988).

2.2.3 K-theory

The wind profile can also be derived from the equations for horizontal motions. the equations can only be solved when the system is closed. To do so, an assumption, a closing hypothesis must be introduced. The simplest one is to relate the flux of momentum linearly to the wind speed gradient in the vertical:

$$\overline{u'w'} = K_m\frac{\partial u}{\partial z} \tag{13}$$

in which K_m is the eddy viscosity for momentum transfer, also called the eddy exchange coefficient, which is assumed to depend on height and u_*, for example $K_m=\kappa u_* z$ in the neutral surface layer (see Pasquill and Smith, p 41). This is a first-order closure.

K-theory is widely applied to dispersion problems, but to a lesser extent in gaussian models. The diffusivity constants K_m, K_h, K_y and K_z stand for the diffusivity of momentum and heat in the vertical direction and of non-buoyant gases in the *y*- and *z*-direction. Although the *K*-values (for the vertical direction) principally differ, for practical applications they are often assumed to be equal. In particular the assumption K_m=K_h is commonly made and has some theoretical justification. Values of K_m can be derived from the similarity theory as follows. From the wind profile u_* can be solved:

$$u_* = \frac{\kappa}{\ln(z/z_0)} u \tag{14}$$

If this formula is squared and multiplied by air density ρ, we have an expression for the surface stress τ:

$$\tau = \frac{\kappa^2 \rho u^2}{\ln^2(z/z_0)} \tag{15}$$

From (13) we have:

$$\begin{aligned} -\tau &= \rho \overline{u'w'} = \rho K_z \frac{du}{dz} \\ \therefore \quad K_m &= -\frac{\tau}{\rho du/dz} \end{aligned} \tag{16}$$

The wind profile *du/dz* in this equation can be replaced by substituting *du/dz* from (8); using $\theta = u_*^2$ this yields:

$$K_m = \kappa \frac{u_* z}{\phi_m(\zeta)} \tag{17}$$

in which $\phi(\zeta)$ represents stability functions, having different forms for different stability regimes (stable/neutral/unstable). This expression is valid only up to a height of 0.1 *h* (the surface layer); above this height, K_z is commonly assumed to be constant up to a height of 0.8 *h*.
Similar expressions as (17) can be obtained for heat, when θ_h is used as the stability function for heat. K_z can also be obtained by assuming, as discussed above, that K_z=K_h
An alternative for neutral conditions is proposed by Shir (1973):

$$K_z = 0.4 u_* z e^{-4\frac{fz}{u_*}} \tag{18}$$

which should be valid for the whole depth of the neutral boundary layer.

K-theory is can be applied in gaussian models, using:

$$\sigma_y = \sqrt{\frac{2K_y x}{\bar{u}}} \quad and: \qquad \sigma_z = \sqrt{\frac{2K_z x}{\bar{u}}} \qquad \textbf{(19)}$$

These formulae restrict the dispersion parameters to varying as $x^{1/2}$. As can be seen from the well-known Pasquill dispersion curves, the exponent of the curves tend to shift from 1 to 0.5 when ranging from unstable to stable conditions. During this shift the timescale is decreasing and hence the requirement for K-theory to be valid, namely that the travel time is much greater than the turbulence time scale, is gradually being fulfilled. In K-theory $\sigma_z^2 = 2K_z t_k$, in which t_k is the effective travel time. Connecting K-theory with Taylor theory means that both σ's should be equal, resulting in

$$\sigma_z^2 = 2K_z t_k = 2\sigma_w^2 T_L^2 \left[\frac{t}{T_L} - (1 - e^{-t/T_L}) \right] \qquad \textbf{(20)}$$

in which T_L is the Lagrangian time scale at the source height. If we wish K_z to be independent of travel time and $t_k = t$ at large times we can identify K_z with $\sigma_w^2 T_L$ and so we have:

$$t_k = t - T_L(1 - e^{-t/T_L}) \qquad \textbf{(21)}$$

The travel time in K-theory can be considered to be 'corrected' by (21) to make it consistent with statistical theory. The reason for this correction is that K-theory coming from the Fickian diffusion type assumes that there is no correlation in the movements. In other words, K-theory does not consider the spectrum of turbulent fluctuations that results in the autocorrelation function with a certain timescale. An alternative method of correcting K-theory is possible in which t_k=t but K_z varies with t, becoming an "effective diffusivity".

2.2.4 Wind direction profile

When a similar expression as (13) for the *v*-component is formulated, the two equations can be solved simultaneously and the change of the wind is described in complex form. This solution was first found by Ekman (1905) for a problem in the upper ocean and soon modified for the atmosphere.
Not only the wind speed is given, but also the wind direction:

$$u(z) = U_g \; [1 \; - \; e^{-\gamma z} \cos(\gamma z)] \qquad v(z) = U_g \; e^{-\gamma z} \sin(\gamma z) \qquad \textbf{(22)}$$

This 'Ekman-spiral' solution (in which $\gamma=\sqrt{\{f/2K_m\}}$) is however very limited since it assumes that K_m is constant in the whole depth of the boundary layer. The value of K_m is dependent on height and the stability of the atmosphere and therefore on the heating and the cooling of the surface. It gives a useful qualitative view of the clockwise turning of the wind direction, but it is probably not quantitatively accurate in most cases. Hence, another approach is advisable, where experimental data are used, instead of the theoretical approach. In a comprehensive analysis of the turning of the wind direction with height using a long time series of measurements in The Netherlands at Cabauw (made froma 213 m meteo tower), Van Ulden and Holtslag came up with a empirical description. The performance of this description would be better as the theoretical Ekman profile, described above. In their analysis an exponential function is suggested:

$$D(z)=d_1 D(h)\ (1-e^{(-d_2 z/h)}) \qquad \textbf{(23)}$$

with d_1=1.23 and d_2=1.75 are pure empirical parameters from the Cabauw-mast analysis.
In the up-dated Dutch National Model (Van Ham et al, 1996) gives for the maximum turning of the wind direction over the whole depth of the boundary-layer $D(h)$ is:

$D(h)= 0^\circ$	if $h/L_* < -10$ (Nieuwstadt, 1983)
$D(h)= 20^\circ + 25^\circ(1 + 0.18\ h/L_*)$	if $-10 < h/L_* < 0$
$D(h)= 45^\circ$	if $h/L_* > 0$

This formulation is a slightly changed version of Van Ulden and Holtslag's formulation. For very unstable conditions this means no wind direction change in the ABL; the change in wind direction takes completely place in a small region around the top of the ABL and the wind direction within the ABL appears to be nearly uniform.
It must be mentioned here that baroclinic effects on the turning of the wind (and the temperature profile as well) may be stronger than Coriolis forces. Including baroclinicity would probably improve the modelling of the wind direction profile. Baroclinicity can turn the wind more than Coriolis force: it can be up to 180 degrees (while Coriolis effect typically results in turns up to 45 degrees) or even a backing of the wind is possible. The shape of the profile also can be much altered by baroclinicity depending on both its intensity and especially its direction. One should have available some geostrophic wind information which will be lacking in general. We will not address baroclinicity here, interested readers are referred to Arya (1975), Hoxit (1975) and Joffre (1985).

2.3 Formulations of vertical temperature profiles

2.3.1 The surface layer

Temperature profiles can be formulated in a way analogous to the formulations of the wind profile (Panofsky and Dutton, 1984):

$$\frac{\theta(z)-\theta(0)}{\theta_*} = \frac{1}{\kappa}\left[\ln\frac{z}{z_0} - \psi_h\left(\frac{z}{L_*}\right)\right] \quad \textbf{(24)}$$

$$\theta_* = -\frac{H}{\rho c_p u_*} \quad \textbf{(25)}$$

ψ_h is the equivalent function for heat to the function ψ_m for momentum introduced above. It is derived by integrating an empirical function ϕ_h (analogous to ϕ_m) The form of ϕ_h is similar to ϕ_m but, in unstable cases, the exponent is often taken to be 1/2 instead of 1/4 (see equation (9)). In some practical applications the two are assumed equal. The calculation of θ_* from temperature and wind at two heights is performed in a similar way as u_* and L_*; an iteration gives both parameters (L_* and u_*, or θ_* and u_*). For recommendations top calculate L_* and u_* we refer to the report of working group 1.

2.3.2 Above the surface layer

Temperature profiles in the core of the ABL and immediately above the ABL are of interest for plume rise calculations and inversion rise modelling. For the core of the ABL ($0.1<z/h<1$), no generally applicable formulas are found. Above the ABL certain models assume a constant inversion strength (HPDM uses $d\theta/dz$=0.010 K/m for nighttime conditions and $d\theta/dz$=0.005 K/m for daytime conditions). Since the growth of the ABL (dh/dt) is very sensitive to this temperature gradient, a better way would be the derivation of $d\theta/dz$ from radiosoundings. This more detailed approach is sometimes possible. In The Netherlands a detailed data set, derived from routine balloon soundings has been used to determine an averaged (over 3 years) morning (06h00) temperature profile for the first 1060 m of the atmosphere for different seasons and wind directions (Erbrink, 1995). It was found that the **amplitude** of the profile is clearly a function of month-number M (ranging from 1 to 12) and wind direction D (in degrees):

$$\theta_z(M,D) = T_a + \theta_{z,norm}\left[7+1.8\ C_s + (4+1.5C_s\cos(\frac{D-90}{180}\pi)\right] \quad \textbf{(26)}$$

$$\textit{with:} \qquad C_s = \cos(\frac{M-1}{6}\pi)$$

which, of course, can only be used for Dutch (or similar) circumstances.

The normalized mean potential temperature profile $\theta_{z,norm}$ is obtained by normalizing the mean potential temperature profile for all z values by setting θ=1 K at 1060 m. The mean potential temperature profile is obtained by taking the average profile over all 6h00 soundings over three years. The scatter in the modelled-observed plot can be reduced by adding a correction term for the difference in maximum ground level temperature on the previous day and the actual temperature at 06h00 $\Delta T_{max\text{-}6}$: $\theta_z'=\theta_z - 2.5 + 0.35\ \Delta T_{max\text{-}6}$, thus making use of measured cooling down effects during the night, resulting in a value of the correlation coefficient of 0.67.

An alternative way could be the use of gridded forecasted or analyzed profiles from mesoscale models. These data are sometimes routinely available from National Meteorological offices and can be used as input for dispersion models.

2.4 Turbulence profiles

Some reviews to describe the turbulence profiles in the atmosphere have been given in the past. We mention Hanna (1982), Weber et al. (1982), Gryning et al. (1987), and Zannetti (1990). In general their formulations make use of the scaling parameters L_*, u_*, h or related dimensionless forms (such as z/L_*). Many of the initial formulations of the turbulence profiles are based on the analysis of the Minnesota experiments. More recent formulations are also based on water tank experiments of Willis and Deardorff and Large Eddy Simulations (LES) and some other field experiments.
A number of formulas for σ_v and σ_w have been proposed by various authors. Compared to the wind profile formulations most of the formulations for σ_v and σ_w are more empirical functions, derived from field experiments and dimensional analysis. We shall summarize some of these here, separately for unstable, neutral and stable atmospheres.

2.4.1 Unstable atmospheres

Panofsky et al. (1977) give as a result of the Minnesota data (measuring height in the range 4-32 m; sampling time ~1h):

$$\begin{aligned}\sigma_v(z) &= u_*(12-0.5\frac{h}{L_*})^{1/3}\\ \sigma_w(z) &= 1.25u_*(1-3\frac{z}{L_*})^{1/3}\end{aligned} \qquad \textbf{(27)}$$

The latter formulation is the simpler of the two σ_w formulations proposed by Panofsky et al. and is also based on some wind tunnel data. The σ_v formualtions also agree with the Deardorff water tank experiments.

Hanna (1982) based on Panofsky (1977) and Irwin (1979, reviewing some experiments) recommended the following formula:

$$
\begin{aligned}
&\sigma_v = u_*(12 - 0.5\frac{h}{L_*})^{1/3} \\
&\sigma_w(z) = 0.96 w_* \left(\frac{3z}{h} + \frac{|L_*|}{h}\right)^{1/3}, && z/h<0.03 \\
&\sigma_w(z) = \min[0.95 w_* \left(\frac{3z}{h} + \frac{|L_*|}{h}\right)^{1/3},\ 0.763 w_* (\frac{z}{h})^{0.175}], && 0.03<z/h<0.4 \qquad \textbf{(28)} \\
&\sigma_w(z) = 0.722 w_* (1 - \frac{z}{h})^{0.207}, && 0.4<z/h<0.96 \\
&\sigma_w(z) = 0.37 w_*, && 0.96<z/h<1 \\
&\text{with } w_* = u_*(-\frac{h}{kL_*})^{1/3}
\end{aligned}
$$

These relations are often used in practical applications (see e.g. Weber et al., 1982; Renkowski and Wiesner, 1994). Gryning et al. (1987) recommend:

$$
\begin{aligned}
&\sigma_v(z) = u_*[0.35(-\frac{h}{kL_*})^{2/3} + (2 - \frac{z}{h})]^{1/2} && (1) \\
&\sigma_w(z) = u_*[1.5(\frac{z}{-kL_*})^{2/3}\exp(-2\frac{z}{h}) + (1.7 - \frac{z}{h})]^{1/2} && (2)
\end{aligned}
\qquad \textbf{(29)}
$$

Equations (29.1) and (29.2) are based on an empirical model by Brost et al. (1982). Equation (29.1) is also based on Caughey (1982). Equation (29.2) have been tested by Irwin and Paumier (1990, CONDORS experiment, z=10 m, averaging time 20 min). The performance of these formulations is quite fair. For the whole ABL Holtslag and Moeng (1991) proposed a another fomulation which was used for the surface layer in the Dutch National Model and performed better than (29):

$$
\sigma_w^{\,3} = [1.6 u_*^{\,2}(1 - \frac{z}{h})]^{3/2} + 1.2 w_*^{\,3}(\frac{z}{h})(1 - 0.9\frac{z}{h})^{3/2}. \qquad \textbf{(30)}
$$

This expression gives about the same results for the rest of the mixed layer as Gryning et al.'s formulae for unstable conditions, but is superior in describing the dispersion characteristics in the surface layer for the Prairie Grass experiments.

Paumier et al. (1986) mention the following formulae (tested and referred to by Wratt (1987) in New-Zealand with the following: z=8.6 m and 56 m, averaging time 10 min, sampling time 1 h).

$$\sigma_v(z)=[0.38w_*^2+2.7u_*^2\frac{(1-z/h)^2}{(1+2.8z/h)^{2/3}}]^{1/2}$$
$$\sigma_w(z)=[1.54w_*^2(\frac{z}{h})^{2/3}\exp(-2z/h)+1.457u_*^2(1-z/h)^2]^{1/2}, \tag{31}$$

Wilczak and Phillips (1986) give:

$$\sigma_v(z)=0.74w_*\{0.5[2-(\frac{z}{h})^{1/2}]+0.3(\frac{z}{h})\}^{1/2}$$
$$\sigma_w(z)=\{Cw_*^2(\frac{z}{h})^{2/3}[1-0.91(\frac{z}{h})]\}^{1/2} \tag{32}$$

with C=1.8 (Deardorff, 1974, laboratory water tank study; Caughey and Palmer, 1979, Minnesota and Ashchurch experiment), or C=2.5 (BAO data, 1983, 8 levels of sonic anemometers up to 300 m, averaging time 20 min).

2.4.2 Stable atmospheres

Based on Minnesota experiment Hanna (1982) proposed the following formula which was tested by Wratt (1987):

$$\sigma_w(z)=\sigma_v(z)=1.3u_*(1-\frac{z}{h}), \tag{33}$$

Gryning et al. (1987), based on Nieuwstadt (1984) recommends:

$$\sigma_v(z)=u_*[2(1-\frac{z}{h})]^{1/2}$$
$$\sigma_w(z)=u_*[1.7(1-\frac{z}{h})^{3/2}]^{1/2},\ h/L_*>1 \tag{34}$$

Paumier (formulations tested by Wratt, 1987) proposed:

$$\sigma_v(z)=1.643u_*\frac{1-z/h}{(1+2.8z/h)^{1/3}}$$
$$\sigma_w(z)=1.207u_*(1-z/h), \tag{35}$$

2.4.3 Neutral atmospheres

Hanna (1982), based on Wyngaard et al. (1974) suggested.

$$\sigma_v(z)=\sigma_w(z)=1.3u_*\exp(-2\frac{fz}{u_*}), \qquad \textbf{(36)}$$

Panofsky and Dutton (1984) propose:

$$\begin{aligned}\sigma_v&=bu_*, \quad b=1.92\pm0.05\\ \sigma_w&=cu_*, \quad c=1.25\pm0.03\end{aligned} \qquad \textbf{(37)}$$

for the surface layer only.

In an analysis of Erbrink (1995) Gryning' et al.'s (1987) formulations for unstable and the formulation of Panofsky for stable turned out to be slightly better than others, when compared with a dataset of surface observations over all stabilities in the Netherlands. The expressions for both σ_v and σ_w values were tested with the (surface) data from the Detailed Hourly Meteorological (DHM) data set. This data set is specially made for application in the major part of the Netherlands with measurements from hourly surface observations and 6-hourly balloon soundings (at 0h00, 6h00, 12h00 and 18h00). The length of this data set is from March, 1990 until March 1993. The routine surface observations are carried out at a meteorological site near Wageningen and the routine balloon soundings in de Bilt (the home of the Dutch meteorological office), both in the center of the Netherlands. The values of σ_{vf}, σ_{vs} and T_l are derived from lateral wind measurements applying digital filtering techniques, as is described in Erbrink (1989) and Erbrink and Bange (1992).

To obtain realistic values of the time scale, it is important to filter out the highest frequencies with a second-order low-pass filter with time constants of 1 and 5 s. The slow fluctuations are filtered out by calculating the moving average with averaging times ranging from 5 minutes (stable) to 30 minutes (unstable). From these fast fluctuations the intensity (σ_{vf}) and Eulerian time scale (T_e) is calculated. The remaining (slow) lateral wind fluctuations (giving σ_{vs}) have different causes and are difficult to predict with boundary layer relations. The lagrangian time scale T_l itself is not measured, but is derived from T_e, using $T_l=\beta T_e$. In some applications (e.g. Pasquill and Smith, 1983) β is thought to be inversely dependent on turbulence intensity. In most cases β is treated as a constant, many values are reported for β, ranging from 2 up to 10 (Moore et al., 1985; Lee and Stone, 1983); we have taken β=3, which fits in best with the results from dispersion experiments (Van Duuren and Erbrink, 1988). The correlation coefficients for the comparison between measured and predicted values and standard errors (SE) for σ_v are listed in Table 1. In this comparison the measured values of σ_{vf} are used (derived with our method of digital filtering of wind measurements), because this parameter is expected to reflect turbulence better than σ_{vs} or σ_v. The results for most schemes are remarkably good, implicitely confirming the usefulness of our method to estimate atmospheric turbulence.

Table 1 Correlation coefficients R and standard errors (SE) for the comparison between measured and computed values of σ_v. For σ_v-measured σ_{vf}-values are used.

	Values of R and SE (m/s)						
	Method	Stable		neutral		unstable	
		2722 hours		663 hours		1173 hours	
		R	SE	R	SE	R	SE
σ_v	Panovsky	0.70	0.14	0.76	0.16	0.73	0.21
	Gryning	0.72	0.14	-	-	0.81	0.19
	Hanna	0.72	0.14	0.76	0.16	-	-

2.5 Time scales

Very few data are available for the integral Lagrangian time scale T_l. Most formulations are based on the spectral analysis of Kaimal (1972). Hanna (1982) uses tower (spectra) data to deduce:

$$T_l = \frac{z/\sigma_w}{2(1+15fz/u_*)} \tag{38}$$

which is assumed to be valid for all three turbulent components and for neutral conditions. For other stabilities Hanna recommends somewhat different expressions. These expressions reflect the height dependency of the time scale. Erbrink (1995) found a practical method[1] to measure the time scale and correlated his data to the formula of Hanna: the agreement is very poor. Cenedese (1994) found from water tank experiments where fine particles are tracked with laser doppler techniques that the height dependency of Hanna's formula is poor for convective conditions. In many cases, a constant value for the time scale is used; Gryning et al. (1987) recommend constant values of 30 and 300 s (for stable and unstable conditions respectively), following Draxler (1976); the HPDM model uses a value of 15,000 s for the *y*-direction, which is clearly meant not to reflect the turbulent but the low frequency meander σ_{vs}. In contrast, Erbrink and Scholten (1995) recommend after a detailed analysis to use a height-independent function:

$$T_{l_{y,z}} = 26\ \frac{\sigma_v}{u} \left[\ln(\frac{z_r}{z_0})\right]^2 \tag{39}$$

where β=0.4 u/σ_v is assumed; the observation height z_r for σ_v and u is normally 10 m. The height dependency of T_l is not studied in their work, but it is thinkable that T_l varies with

[1] Both σ_v, σ_w and the time scale of turbulence T_l are derived from simple lateral wind measurements. This is done by separating the fluctuations of the wind direction into a stability-dependent and a non-stability dependent part.

height since σ_v is height dependent; other research in water tanks results in the opposite (see appendix B).

We will now discuss the analysis of Erbrink and Scholten in more detail. In this analysis Erbrink used the results of 4 measuring stations (in the Netherlands), at which the COPS-method (Calculation of Pasquill Stability; Erbrink, 1989) turbulence parameters (σ_{vs}, σ_{vf} and T_e) were measured at least over the period of a year. Those sites vary in roughness lengths: the marine measuring site 'Meet Post Noordwijk' (MPN) with z_0=0.03 cm (estimated from surface type using Figure 9.6 of Stull, 1988); Abbenes with z_0=5 cm (computed from several months of u_* measurements with a sonic anemometer); Wageningen with z_0=30 cm (wind from city-directions) or z_0=10 cm (other wind directions) and Kollum (a very flat open terrain in the north part of the country) with z_0=3 cm. The roughness length at Wageningen and Kollum are determined by Hanna's (1981) method of estimating z_0 from σ_θ measurements:

$$z_0 = z\ \exp(-\kappa c_v\ /\sigma_\theta) \tag{40}$$

Recommended values for c_v range from 1.64 to 2.0, but the question is which part of the turbulent spectrum should be included in the value of σ_θ. To calibrate this formula with our measured values of σ_θ with COPS (which are actually filtered σ_θ values, see section 2.4.3), we used z_0=5 cm for the Abbenes measurements obtained with the sonic, and calculated the averaged value of c_v during neutral conditions (being defined as $|L_*|>1000$); this results in c_v=1.40. Then (22) is applied to obtain wind direction dependent z_0 values at the other sites. Time scale values (as T_e) are plotted against $(\ln(z/z_0))*(\sigma_v/u)$ in a log-log plot. The result is shown in Figure 4; all data points merge together resulting in the empirical fit:

$$T_e = 65\ \left[\frac{\sigma_v}{u}\ \ln(\frac{z}{z_0})\right]^2 \tag{41}$$

The vertical clustering is due to the stepwise way of calculating T_e and σ_v from the wind fluctuations; the averaging time to filter out the slow fluctuations is adjusted to the calculated value of T_e in the first step, as is described in Erbrink (1991). This does not influence the results significantly. T_e is measured at the standard height of 10 m, except at MPN where the measuring height was about 20 m (depending on the sea surface level). Including the dependence on both z/z_0 and σ_v/u seems to generalize all measurements to a great extent. Although the time scale is thought to be dependent of height (see for example the work of Cenedese et al, as reported in Appendix C) we could not find a consistent formulation. Hence, we think it is best to use the calculated time scale values at a fixed height of 10 m, since time scales measured according to Erbrink's method are not available beyont heights of 10 or 20 m. The ratio z_r/z_0 in (39) and (41) is meant more to account for roughness differences rather than to different heights.

Figure 4 Combined plot of all data from 4 measuring sites with different roughness lengths. Measurements over several years, all stabilities. Left: $\ln(T_e)$ vs $\ln(\sigma_V/u)$; *: MPN, x: Kollum; ○: Abbenes, □: Wageningen. Right: $\ln(T_e)$ vs $\ln\{\sigma_V/u)*\ln(z/z_0)\}$. The large symbols apply to average values in both left and right plots.

This is recommended for the whole depth of the ABL, except for the surface layer. For the surface-layer (typically being the first 50 to 60 m of the atmosphere) an alternative method with a higher performance than (39) for the Prairie Grass data is found (Flesch et al. 1995):

$$
\begin{aligned}
T_l &= \frac{z}{2\sigma_w}\left(\frac{1}{1+5z/L_*}\right), \quad \textit{for } L_*>0, \\
T_l &= \frac{z}{2\sigma_w}\left(1-6\frac{z}{L_*}\right)^{0.25}, \quad \textit{for } L_*<0
\end{aligned}
\qquad \textbf{(42)}
$$

This formula (42) is especially valuable for applications in vertical diffusion (T_{lz}), for lateral we do not have evidence to give a preference for (42) or (39) and (41). For simplicity we recommend the all (39), (41) and (42) for both lateral and vertical time scales. In Figure 5 the concentrations obtained with the advanced gaussian dispersion model STACKS are shown using Gryning's σ_w and the height independent function of Erbrink (39) for T_l on the one hand and Holtslag and Moeng's formulation for σ_w together with (42) for T_l on the other hand. These formulations (42) are useful for the surface layer and give about the same results as (39) at the top of the surface layer (say: 50 m).

This reasoning about time scales seems a bit inconsistent, because values measured at the level of 10 m are used for heights above the 50 m level; while for the surface layer (42) is used. In fact (39) and (41) are derived from the horizontal fluctuations and (42) is selected mainly for the vertical component (Prairie Grass data do not consider horizontal dispersion). We think that the measured horizontal time scale is a good estimator for a constant time scale in the bulk of the mixing layer (in both horizontal and vertical direction), but vertical time scales will decrease close to the ground. The basic thinking is that an appropriate description for the vertical fluctuations is most important; assuming local isotropy the horizontal time scale is a derivate from the vertical time scale.

Figure 5 Calculated crosswind integrated concentrations of the Prairie grass experiments, compared with predictions of the gaussian model STACKS. Left: applying Grynings functions for σ_w and Erbrinks function (39) for T_l; right: applying the formulation of Holtslag and Moeng for σ_w together with (42) for T_l.

3 DATASETS AND ADDITIONAL VALIDATION

3.1 Criteria

Data sets for validation purposes are available in many forms. We will make a difference here between **"recognized" datasets** and others; between data sets from **field measurements**, from w**ind tunnel experiments**, from **water tank experiments** and from **Large Eddy Simulations (LES)**. Although the latter do not concern real observations they are treated in many comparisons on the same level as 'real' observations. The reason for this is the 'belief' that LES are able to reproduce the atmospheric flows and turbulence correctly for both homogeneous and non-homogeneous atmospheric conditions. We use the term "recognized" data set to distinguish them from the many data sets that are poorly described and hardly known in the international scientific world (their use is often limited to only one site and/or even one research group). Recognized data sets are widely used and available in a well-documented form (such as the EPRI-Kincaid dispersion dataset).

Observations from field experiments can be derived from:
* Mast measurements
* Acoustic Profiles (sodars)
* Tethered balloon soundings
* aircraft measurements

In the following we attempt to survey those datasets which are suitable for developing and assessing profile formulations. We also summarize the results of three experimental studies carried out by working group members.

3.2 Recognized Data sets

In COST 710 we found the following (recognized) data sets
* CONDORS at the 380 m meteorological mast near Boulder (Moninger et al., 1983; Eberhard et al., 1985 and Kaimal et al., 1986),
* MADONNA - Meteorology and diffusion over non-uniform areas (Weber et al., 1995)
* MESOGERS-84 experiment (Weil, 1988)
* EFADA: a study on desertification in Spain (Bolle et al., 1993)
* PRAIRIE GRASS experiment (Haugen, 1959), KANSAS experiment (Kaimal et al., 1972)
* MINNESOTA experiment (Kaimal et al., 1976),
* CABAUW datasets: routine measurements at the 213 m meteorological tower in the centre of the Netherlands (available from KNMI, de Bilt, NL),
* The laboratory experiments of Willis and Deardorff (1978)

* Large Eddy simulations, reported by Nieuwstadt (1992), Nieuwstadt and Bouwmans (1994) and Henn and Sykes (1992). A detailed comparison of 4 LES-models of Mason, Moeng, Nieuwstadt and Schumann was recently performed for both the convective boundary layer and the neutral ABL. (Andren et al., 1994 and Nieuwstadt et al, 1996).

3.3 Independent validation with other data sets

A number of studies to validate some of the above profile formulae were carried out by the working group using the following (unrecognized) data sets of:

Swiss sodar and masts data sets

Italian data sets (water tank data; full scale data from experiments near Rome)

Belgian data sets (meteorological mast in Mol)

Since many formulations lean on a restricted number of data originating from one or two specific sites with local restrictions (following from site specific meteorological conditions, terrain features and so on), some additional verifications have been done with independent data sets. The results of these verifications are reported in separate appendices of this report and their findings are used in the final recommendations. In the rest of section 3 we summarise this work.

3.3.1 Results from Swiss mast and sodar datasets

In Switzerland the meteorological office has available several data sets from mast and sodar in the vicinity of Payerne (a site in between mountainous Jura and Alpine Berner-Oberland). The results of the validation for wind speed and σ_w profiles are reported in appendix A. The aim of this study was to evaluate the ability of models to reproduce observed wind speed and σ_w profiles in the ABL. The models combine theoretical formulas for the lower atmosphere and different parameterizations of the relevant processes in the surface layer. Different formulations need different input parameters (such as radiation data or temperature profiles, and winds at different heights) but they calculate, in principle, the same output. This output is compared to the measured values. The wind speed and the σ_w profile models, originally developed for flat and homogeneous terrain, have been used here in a gentle hilly site. The signature of the topography effects has been revealed by slight modifications of the surface wind profile. The study of this effect is beyond the scope of this report.

Wind profiles

In unstable conditions, the main parameter which controls the profile extrapolation is u_*. The differences between the measured and the calculated wind speeds are positive and, at first order, independent of L_* and z. It is shown that the heights of the anemometers significantly affect the results for the Payerne site. A judicious choice of z_0 or performing

measurements in the upper part of the surface layer (which gives larger values of u_*) produces the best results.
In stable conditions, both L_* and u_* control the profile extrapolation. The differences between the measured and the calculated wind speeds are in general negative and L_* and z dependent. It is shown that the choice of the radiation schemes affect significantly the results. For very stable conditions, the temperature gradient method fails compared to the one based on the global or the net radiation method. Using one of these last two methods and a wind gradient measured close to the surface (smaller u_* values) gives the best results.
Regarding the sonic anemometer, measured u_* and L_* produce good results in the unstable regime but are not adequate in stable regime.
In general, it should be emphasized that extrapolation should be restricted to the mixing height. Typically, it is a few hundreds of meters high for most stable cases.

σ_w *profiles*
In unstable conditions the main parameters which control the profile extrapolation are u_*, h and L_*. The differences between the measured and the calculated values of σ_w are L_* and z dependent. This dependency is obvious for the slightly unstable but less clear for strongly unstable. The temperature gradient method associated with wind measurements in the lower surface layer (smaller u_* values) is the optimal combination. Using the values of u_*, and $L_*$from the sonic data do not produce good results in the unstable to neutral regime.
In stable conditions, the main parameter controlling the profile calculation is u_*. The differences between the measured and the calculated values of σ_w are positive and, at first order, independent of L_* and z. Close to neutral conditions, the sensitivity to the precise height of the anemometers is high due to different u_* values. σ_w values measured with sodar are never smaller than 0.15 m/s. This phenomenon could limit the validity of the sodar data in very stable conditions.
The comparison between measured and calculated σ_w values is better when sonic anemometer data are used in the stable regime, thanks to larger u_* values. In general, it should be emphasized that application of the σ_w calculations should be restricted to the mixing height.
The standard configuration considered here for the wind profile (z_0 value and wind speed at 11 m) is probably the most common set-up in meteorological networks. The sonic measurements would have been certainly more adequate as a reference but such devices are still uncommon. Regarding the sodar data, sodar is usually considered as a valuable instrument for wind speed profile measurements, but the performance is less good for σ_w measurements. Hence, results for σ_w profiles should be considered more in a relative rather than in absolute sense. Low σ_w values are probably overestimated by sodar.

3.3.2 Results from Italian water tank data and full scale dataset

In Italy a dataset from a full scale field campaign in the Rome area and data from water tank experiments were used to test models for wind profiles and turbulence profiles. In the

appendix B the results of the validation for wind speed and σ_w profiles are reported, together with the descriptions of full scale and laboratory water tank experiments.
Vertical profiles of wind speed and σ_w have been measured in a coastal site and compared with Monin-Obukhov similarity theory (MOST) estimations. Starting from measurements close to the surface, vertical profiles have been estimated by means of MOST. Sodar and mast measurements of wind speed and turbulence have been compared with the modelled profiles. In addition water tank results of σ_w and time scales profiles have been included.

Wind speed
The calculated scaling parameters when making use of the MOST are strongly affected by the characteristics of the breeze and they are not able to reproduce the wind speed observations at the upper layers, where the sea breeze wind rapidly decreases. With this in mind, it is clear that the similarity theory has to be applied with caution in all that situations in which sea and land breeze occur. A suitable similarity theory to describe the Thermal Internal Boundary Layer produced by the breeze circulation should be carried out. Introductions and ideas have already been reported in literature (Raynor and Watson (1991), Hanna et al. (1985) and Hsu (1986)).

Turbulence
The only parameter involved in the similarity for σ_w profiles seems to be the friction velocity. As for the wind speed profiles, this unexpected behaviour can be due to the effect of the breeze circulation. In this case the breeze makes the vertical profile of turbulence uniform along the vertical direction. It has to be underlined that a systematic underestimation is present in our sodar measurements. In fact, because of low burst repetition frequency, the higher frequencies are cut. In addition, the lower sodar measurements (44 m) have to be considered with caution (the antenna reverberation may affects this measurement).
It is concluded that from both the observed wind speed and turbulence profiles, a discrepancy with the suggested laws appears. A possible overcoming of this inconsistency can be achieved by improving the schemes utilised in these methods, i.e. by introducing other scaling variables able to characterise the ABL also in the case of sea and land breeze regimes.

Conclusions
Vertical profiles of wind speed and σ_w values have been measured in a coastal site and compared with Monin-Obukhov similarity theory (MOST) estimations. Up to 30 m mast sensors were employed while for the upper layers a sodar has been utilised. From the analysis of the results we can state that the MOST does not adequately describe the Thermal Internal Boundary Layer produced by the breeze circulation. A more realistic description might be given by considering other scaling variables able to take also account of these breeze features.
Furthermore, water tank simulations of the CBL have been analyzed by means of LDA and PTV techniques. Vertical profiles of σ_w and time scales have been derived. Both the

Eulerian and Lagrangian time scales show a strong vertical inhomogeneity. Splitting the Lagrangian statistics into upward and downward moving particles, yields completely different time scale profiles, For the lagrangian time scale insufficient evidence has been found to recommend a clear height dependency.

3.3.3 Results from Belgian data sets (Mast)

One application of profile calculation schemes is in the calculation of stability indexes in terms of winds and temperature at certain levels in situations where the available data are at different levels. Here we report on an investigation into the utility of various methods of calculating the Bultynck-Malet stability class (see Bultynck and Malet, 1972) using data other than that directly referred to in the definition of these stability classes. The methods used make assumptions about the surface energy balance as well as the profile shapes and so the work is related to the ideas discussed by working group 1 as well as to those presented in this report. In the appendix C the results of the validation for wind speed and temperature profiles are reported.

A comparison is made of the wind speed and temperature profiles measured along a 120 m meteorological tower with the profiles obtained by the following equations:

$$u(z)=\frac{u_*}{\kappa}\left[\ln\frac{(z-z_d)}{z_0}-\psi_m(\frac{z}{L_*})\right] \tag{43}$$

where z_d is the displacements height.

$$\frac{\theta(z)-\theta(0)}{\theta_*}=\frac{1}{\kappa}\left[\ln(\frac{z}{z_0})-\psi_h(\frac{z}{L_*})\right] \tag{44}$$

where u_*, L_*, ψ_m and ψ_h are calculated with the KNMI software library for the calculation of surface fluxes (Beljaars and Holtslag, 1990).

The aim is to find out how well profiles, derived from routine synoptic meteorological data or observations along a small meteorological tower ($z \leq 30$ m), compare with measured profiles along a higher (and more expensive) meteorological tower, and to determine the sensitivity of the calculation scheme to surface and vegetation dependent parameters.

Seven combinations of wind speed, temperature and cloudiness data from:

* a synoptic meteorological station situated nearby the 120 m meteorological tower
* the lower levels of the 120 m meteorological tower forthe year 1985,

have been used as input data to the routines of the KNMI library. The data are measured over a relatively open, flat and sandy region.

First, profiles are calculated using the KNMI library default values for the soil and vegetation dependent parameters: albedo, Priestley-Taylor parameter and bulk heat transfer coefficient in the soil. With these default values:

* Wind speed values (z=69 m) are well reproduced (correlation > = 0.95) using a single wind speed observation (z=12 m above zero displacement length) and cloudiness, or using a single wind speed and the temperature measured at two heights (z=8m, z=24 m).
* Potential temperature values (z=114 m, z=8 m) are reproduced well (correlation > 0.80) only if temperatures for two heights (z=8 m, z=24 m) are specified. Input combinations without measured temperature profile are found to give poor results (correlation < 0.70).

Secondly, the ratio of the potential temperature gradient and the square of the wind speed at 69 m are used to calculate the bulk Richardson number R_b. The values of R_b are mapped upon 6 classes, corresponding to two classes of stable, one class of neutral and three classes of unstable atmospheric stability. The joint frequency tables of measured and calculated R_b classes are used to evaluate the suitability of the calculated temperature and wind speed profiles using different input parameter combinations to assess the dispersion capabilities of the atmosphere. Depending upon the input data used, the calculated R_b values are in the same class as the measured R_b value for 44% to 71% of the time for the investigated year 1985. The best results are obtained with the wind speed and temperature measured at two heights as input data, with the additional condition that the upper heights for wind speed and temperature are the same.

4 RECOMMENDATIONS

From our studies and experimental analysis, we give recommendations for the wind speed, temperature and turbulence profiles. We realize that some of the recommendations are more the results of our scientific consensus rather than strongly encouraged by thorough experimental validations. This is especially the case for the temperature profile in the surface layer and the turbulence profiles above the surface layer. We are aware of the fact that too few experimental data are available to recommend these formulations for different terrain types. The reader should be careful in using these formulations since the accuracy is sometimes very limited. But we feel that our findings are good starting points for scientists who have to make a choice for their application in dispersion models.
The use of Monin-Obukhov similarity theory (MOST) is recommended, for MOST is the only theory tested widely and applied to describe wind speed, temperature and turbulence profiles in atmospheric boundary layer processes. This does not mean at all that all problems have been solved as has been made clear in the work done during this COST-action. Summarizing our findings we think it is best to use the following expressions for the profiles of:

4.1 Wind speed:

$$u(z) = \frac{u_*}{\kappa}[\ln\frac{z}{z_0} - \psi_m(\frac{z}{L_*})] \qquad \textbf{(45)}$$

with ψ_m=0 in neutral conditions. For unstable conditions:

$$\psi_m(z/L_*) = \ln\left[\frac{1+x^2}{2}\left(\frac{1+x}{2}\right)^2\right] - 2\ \arctan(x) + \frac{\pi}{2}$$
$$where: \quad x = (1-16\frac{z}{L_*})^{1/4} \qquad \textbf{(46)}$$

and for stable conditions:

$$\psi_m = -5\frac{z}{L_*} \qquad \textbf{(47)}$$

These formulations should be restricted to a height of 200m; above this height the wind speed can been assumed to be constant.

4.2 Wind direction

No recommendations can be given. It is best to analyze historical measurements along a mast (or done by profilers) and fit the data as dimensionless functions of *z/h* and *h/L* being the best available height and stability parameters at the time being. This approach has been worked out by Van Ulden and Holtslag and seems to be satisfactory and is applied in the Dutch National Model. They give the right sort of behaviour and may be useful for climatological studies but will be unsatisfactory on a case by case basis due to other influences that dominate the wind direction profile. Instead of mast data, it is advisable to use routine output of weather forecast models on the regional scales. These models produce vertical profiles for relevant parameters in regular grids; these computed profiles maybe better for locations far from WMO-stations with daily balloon soundings.

4.3 Temperature

$$\frac{\theta(z)-\theta(0)}{\theta_*}=\frac{1}{\kappa}\left[\ln(\frac{z}{z_0})-\psi_h(\frac{z}{L_*})\right]$$

for the lower parts of the ABL (say up to 100 m). This is useful for the purpose of stability determination. For the upper parts of the ABL this formula is not believed to give the proper results. For this part of the ABL we recommend to analyze historical measurements from routine balloon soundings to estimate the average profile for modelling the plume rise, inversion penetration or inversion rise (for convective conditions). Instead of mast data, it is advisable to use routine output of weather forecast models on the regional scales.

4.4 Turbulence

For unstable conditions (0>L>-1000):

$$\sigma_v(z)=u_*[0.35(-\frac{h}{kL_*})^{2/3}+(2-\frac{z}{h})]^{1/2} \qquad \textbf{(49)}$$

$$\sigma_w{}^3=[1.6u_*{}^2(1-\frac{z}{h})]^{3/2}+1.2w_*{}^3(\frac{z}{h})(1-0.9\frac{z}{h})^{3/2}. \qquad \textbf{(50)}$$

For neutral conditions (|L|>1000):

$$\sigma_v(z)=\sigma_w(z)=1.3u_*\exp(-2\frac{fz}{u_*}) \qquad \textbf{(51)}$$

For stable conditions ($0<L<1000$):

$$\sigma_w(z)=\sigma_v(z)=1.3u_*(1-\frac{z}{h}) \quad \textbf{(52)}$$

For the Lagrangian time scale:

$$T_{l_{y,z}}=26\ \frac{\sigma_v}{u}\left[\ln(\frac{z}{z_0})\right]^2 \quad \textbf{(53)}$$

where $\beta=0.4\ u/\sigma_v$ is assumed and where z denotes the measuring height for σ_v and u (normally 10 m).
For $z<50$ m preference is given to:

$$\begin{aligned} T_l &= \frac{z}{2\sigma_w}\left(\frac{1}{1+5z/L_*}\right), \quad for\ L_*>0, \\ T_l &= \frac{z}{2\sigma_w}\left(1-6\frac{z}{L_*}\right)^{0.25}, \quad for\ L_*<0 \end{aligned} \quad \textbf{(54)}$$

4.5 Limitations, uncertainty and future work

All recommendations are given for flat and homogeneous terrain. For wind speed profiles the results are good to fair. For turbulence the quality of the profiles is fair to sometimes questionable and for temperature the results are sometimes fair, but more often questionable. However, no better formulations are found.
The KNMI Surface Flux Software Library (Beljaars and Holtslag, 1990) has been used with data-sets from Belgium, Italy and Switzerland, and its use is recommended by Working Group 3. However, this does not mean that all problems concerning vertical profiles are solved. The KNMI Surface Flux Software Library can be used with meteorological input data of varing quality, ranging from cheap routine meteorological observations (cloudiness, wind speed at 10 meters) over data from meteorological towers of varying height and/or net radiation.

The formulations we found can only be recommended for the height of the surface layer; extension to the whole depth of the boundary layer will be the only practical way to obtain profiles, but in many cases deviations from the calculated values should be expected. Validations and experimental fits are performed with mast data, which do not extend beyond heights of 200 m in general. Above the mixing layer (usually referred to as the inversion height) no recommendations can be given, other than analyzing measurements or output of routine forecasting models, especially when output from local area models is used, which have less coarse resolution in the vertical direction.

In the COST 710 context, different input combinations have been used by the Belgian and Swiss contributors. Both agree on that a proper determination of the surface roughness length z_0 is important when using the library, and that it could be necessary to make z_0 dependent on meteorological conditions, a so-called 'effective' z_0. When a wind speed profile is used as input, the heights of the anemometers have an influence on the resulting values for u_* and L_*. The surface roughness may be a function of height since the integrated effects of terrain features are 'felt' by the wind sensor. A greater height means that the influence of a larger area is included in the wind profile. Whether this influence is significant or marginal should be evaluated on the basis of dispersion model output.
At greater heights the external influence of other effects (topography, baroclinicity, non-stationarity) on the profiles of temperature and wind may become stronger, resulting in a restriction on the surface layer formulations recommended here.

The Swiss data analysis is somewhat difficult to synthesise because of a lack of clear reference data. The values of u_* and L_* measured by the sonic anemometer may be considered to be, during some stability conditions, too local measurements for allowing an extrapolation towards altitudes of several hundred meters. The Swiss work however clearly shows that the optimal parameter configurations for calculating vertical profiles for wind speed and for calculating σ_w are different, and vary with the stability conditions.

This is also the conclusion of the Belgian data analysis, where the calculated wind speed and temperature profile is compared directly with the measured values (here, u_* and L_* are used in an implicit way). Most remarkably here is the difference between the measured temperature profile and the calculated one, using solar elevation and cloud cover. A temperature gradient or a measured radiation quantity seems to be required as a minimum. The wind speed profile is well reproduced in the Belgian study using a single wind speed and a value for the surface roughness length z_0 and the zero displacement z_d.

The Italian work finally points out that the KNMI library must be used with caution -or is not applicable for- locations where (strong) sea-land breezes occur. They found that the scheme of Stull was the most appropriate for describing the vertical profile of σ_w under such conditions, whereas the Panofsky formula is not appropriate. In the Swiss study, the Gryning et al. formulation has been used, but the measured σ_w used in this study were considered to be 'not having the required precision to distinguish between different formulations'.

The recommendations of this working group are consequently that:

- work should be continued on the basis of the KNMI surface fluxes software library
- a more systematic evaluation should be made of the minimum required input data to obtain vertical profile estimates of a quality required by dispersion models
- that this be done for different locations in Europe, because some redundancy in location characteristics is needed to avoid local particularities dominating the results
- data measured along meteorological towers are preferably used as reference material, as indirect measurements from sodars or other wind profilers should actually to be interpreted as ‘modelled' values, so their performance should be scrutinized in the same way as the “modelled” profiles that this working group has dealt with;
- developing and using a standard set of statistical tests, graphical representations and data handling protocols agreed upon by the workers in the field so that data from different places become comparable.
- All datasets investigated here are restricted to rural areas. For the urban environment, similar research should be carried out.
- The possibility to of using profile data from routine weather forecasting models should be investigated to see if they can improve the quality and the amount of data used in air pollution modelling.

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