Uncertainties in met pre-processing for dispersion models

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There are various quantities which are not routinely measured and which need to be estimated for dispersion modelling. The process of estimating these from routine meteorological observations is known as 'met pre-processing'. The two main quantities needed are the depth of the boundary layer and some measure of stability. Traditionally, older Gaussian plume type models used the Pasquill (1961) stability categories. More modern models tend to use the Monin-Obukhov length scale *L*, although in fact this concept dates from an earlier time in the 1950s. At the time Pasquill developed his scheme he was well aware of the earlier work by Monin and Obukhov, but his aim was to produce a quick and easy empirical approach for use when other measurements were lacking. This paper discusses met pre-processing in connection with the estimation of these stability measures and the estimation of boundary layer depth. It then discusses the origins of uncertainties in this pre-processing. A particular complication in met pre-processing concerns the treatment of urban areas and some lessons on uncertainties are drawn from recent urban field experiments.

1 INTRODUCTION

Dispersion models have widespread application. They have a long history in environmental impact analyses. More recently they have found a new application in local air quality management. The early models have their origins in Gaussian plumes, wherein the concentration is assumed to be normally distributed in space about the plume centre-line and the plume spread is described by the dispersion parameters σ_y and σ_z. These are the root mean square plume widths in the crosswind and vertical direction. The dispersion parameters, and especially σ_y, generally vary with averaging time; typically values might be given for a 10 minute or a 1 hour averaging time. The NRPB series of reports, starting with the Gaussian model in Report R91 (Clarke 1979), adopted this approach, as did US EPA models such as the ISC model. More recently models such as ADMS and AERMOD were developed to provide a more up to date description of dispersion. These models take much more account of recent developments in our understanding of turbulent motions, especially in unstable conditions. Convection is not a symmetrical process; updrafts tend to be narrower in cross-section and faster in ascent than the corresponding downdrafts and this leads to non-Gaussian concentration profiles. It is important to realise that the above models (and indeed most other types of model such as Eulerian grid models or Lagrangian particle models) are essentially statistical in character. They do not claim to predict the actual concentration field in a given situation but rather the ensemble average that would be expected if one could repeat the scenario many times with essentially identical external conditions. Such repetitions would differ by random turbulent fluctuations.

In general dispersion models require information on aspects of the meteorology which are not routinely measured; the process of estimating the quantities required from routinely available data is known as 'meteorological pre-processing'. The two main aspects where such estimates are needed concern the stability and the depth of the boundary layer. This talk outlines the nature of meteorological pre-processing and considers reasons for uncertainties caused by this part of the dispersion modelling process. A significant part of this uncertainty concerns the effect of urban areas and some discussion is given below of results from recent field studies of urban meteorology in Birmingham.

2 METEOROLOGICAL DATA PRE-PROCESSING

The main purpose of met pre-processing is to estimate the stability and boundary layer depth.

2.1 Stability measures – Pasquill stability class and Monin-Obukhov length

In dispersion modelling, the term stability is used to characterise the relative importance (and sign) of thermal effects and wind shear in producing turbulence. On a clear sunny day with light winds, the production of turbulence by thermal convection will be much greater than the production due to wind shear and the stability will be 'very unstable'. Conversely at night the radiative cooling of the ground will lead to a thermally stratified flow which will act as a sink for the turbulent energy produced by the wind shear – the stability will then be 'stable'. In strong winds the production of turbulence by the wind will be so large that the thermal effects, even if large in themselves, will have only a small relative effect; in such cases the stability will be 'neutral'.

Pasquill (1961) presented a simple scheme for classifying stability into stability classes labelled A to F. The stability class is determined from wind speed and cloud cover at night and from wind speed and whether the incoming solar radiation is 'strong', 'medium' or 'slight' by day. The scheme was revised by Gifford and by Turner (see e.g. Turner 1970, Gifford 1976), partly in order to give a precise interpretation of Pasquill's strong, medium and slight insolation in terms of solar elevation and cloud cover. Over the years a considerable number of further variations on Pasquill's original approach have been proposed with one of the more significant being Smith's introduction of a more stable class G (see Pasquill and Smith, 1983). The practical value of the approach has stood the test of usefulness for many years, both in models such as R91 and ISC and as a useful teaching tool when introducing the subject of atmospheric dispersion to students of environmental science. However it should be emphasised that Pasquill himself regarded his stability category approach as a method of making rough estimates 'for use in the absence of wind fluctuation data'.

The Monin-Obukhov length *L* is a more quantitative method of estimating stability. It requires two quantities not routinely measured by national meteorological networks: the friction velocity u_* and the flux of sensible heat *H*. It expresses the relative importance of thermal effects and wind shear in creating turbulence, and has the form of a length scale because, in unstable conditions, it takes convective thermals some distance to accelerate to the point where they generate significant turbulence in comparison to the wind shear. Also, in stable conditions, the suppression of turbulence by thermal effects generally only becomes significant above a certain height. *L* is defined to be negative in unstable conditions and positive in stable, and its magnitude is a measure of the height above which thermal effects become important, with shear production being dominant below |*L*|. The Monin-Obukhov length is used instead of Pasquill stability in modern dispersion models such as ADMS and AERMOD.

To define *L* we need to consider some surface layer theory. Near the surface, but above the roughness elements, when shear production is dominant, the wind speed obeys a logarithmic wind profile (Oke, 1987; Equation 2.10)

$$u_z = \frac{u_*}{k}\ln\left(\frac{z}{z_0}\right)$$

Here *k* is the von Karman constant, which has a value of about 0.4, and z_0 is the roughness length. Friction velocity u_* is defined by

$$u_*^2 = \sqrt{\overline{u'w'}^2 + \overline{v'w'}^2}$$

where *u′*, *v′*, *w′* are the three fluctuating components of the turbulent flow velocity. A 3-dimensional ultrasonic anemometer can measure *u′*, *v′*, *w′* and a fluctuating temperature θ'. The Monin-Obukhov length is defined as

$$L = -\frac{\theta u_*^{\,3}}{kg\,\overline{w'\theta'}} = -\frac{\rho C_p \theta u_*^3}{kgH}$$

where θ is the absolute temperature in the surface layer, *g* is the acceleration due to gravity, ρ is the density and C_p the specific heat capacity of air at constant pressure. The overbar in the vertical flux of temperature fluctuations $\overline{w'\theta'}$ denotes averaging of the products of fluctuations in temperature and vertical wind velocity. The flux *H* of sensible heat from the surface to the air and $\overline{w'\,\theta'}$ are related by $\overline{w'\,\theta'} = H/(\rho C_p)$. In the UK the sensible heat flux *H* is usually in the range -50 to 200 W m^{-2} and the friction velocity u_* is always positive, typically 0.05 to 0.5 m s^{-1}. Ideally, both u_* and *H* should be measured, and *L* calculated. Such data are rarely available and this leads to uncertainty when estimating *L* (see Section 2.2).

L has the opposite sign to *H*. Thus *L*>0 in stable conditions and *L*<0 in unstable conditions. By definition *L* can have values in the range $-\infty \leq L \leq +\infty$ but values of *L* smaller in magnitude than about 1 m are rare. *L* is positive and small in stable conditions with light winds at night. *L* is small and negative (of order -10 m) on strongly convective days, is of order -100 m on windy days with some solar heating, and tends to infinity in the neutral case with purely mechanical turbulence.

Pasquill stability and the Monin-Obukhov length are trying to describe essentially the same thing. It is not surprising then that there is an approximate relationship between the two stability measures. Golder (1972) investigated this relationship and showed that the Pasquill stability class can be estimated from *L* and the surface roughness length z_0. This approximate relationship is also summarised by Seinfeld (1986).

2.2 Calculating stability from standard meteorological data

The main inputs for calculating stability are governed by availability of data. The measured variables tend to be common to all simple models and rely upon synoptic observations for hourly values of wind speed and direction (generally at 10 metres above ground), cloud amount, time of day, and time of year. Some approaches also require information on humidity, recent rainfall and/or take account of cloud height (thin high cloud has less influence on the radiative fluxes than low cloud).

In the simplest Pasquill-Gifford-Turner stability models one can immediately derive the stability class from this information. More complex models (e.g. R91) will attempt to estimate the surface sensible heat flux as an intermediate parameter and perhaps also u_* (see e.g. Smith's Appendix A in Clarke 1979). Since *H* and u_* are also the parameters needed to evaluate *L*, the pre-processing needed in such approaches is quite similar to that in more modern models based on *L*. During the day the typical approach is to estimate solar elevation, and use this and cloud cover to estimate incoming solar radiation. Then parametrizations of the other terms in the radiation and surface energy budgets lead to an estimate of *H*. The friction velocity u_* is then found using the wind speed and an estimate of roughness length by iteratively solving the equation for the surface layer wind profile (cf Seinfeld, 1986). At night the energy and radiative balances are more subtle and simple physically based parametrizations are harder to find. As a result more empirical approaches are usually adopted. A common approach follows Venkatram (1980) and Holtslag and van Ulden (1982) in estimating the surface layer temperature scale empirically and then obtaining H, u_* and L by iteratively solving the equation for the surface layer wind profile.

2.3 Calculating boundary layer depth from standard meteorological data

A variety of approaches are used to estimate boundary layer depth. For example R91 has a simple look up table of 'typical values' of boundary layer depth for each Pasquill stability class, but it also has a nomogram for daytime boundary layer depth which is based on the Tennekes, Carson and Smith model for the growth of convective boundary layers (Tennekes 1973, Carson 1973, Carson and Smith 1974). The Tennekes, Carson and Smith model is probably the most widely used scheme for daytime boundary layer depths and also forms the basis for the approaches used in ADMS and AERMOD. At night, approaches based on the Zilitinkevich formula ($h^2 \propto Lu_*/f$ where f is the Coriolis parameter) or on the neutral Ekman layer formula ($h \propto u_*/f$) are commonly used or some interpolation between the two.

2.4 Using stability and boundary layer depth information in dispersion models

We saw above how the Pasquill stability class and Monin-Obukhov length are basically trying to represent the same thing and that, in some models at least, the approaches used for estimating them are conceptually very similar. This might lead one to think that the modern models such as ADMS and AERMOD are not so very different from the older models. In fact this is not the case and the modern models are a substantial advance over the older approach. However the difference is more in the way the stability information is used than in the change from Pasquill stability class to Monin-Obukhov length.

In Pasquill-type models the plume spread is independent of the height of the source whereas in the modern approaches the source height (relative to both $|L|$ and h) plays an important role. If we consider vertical dispersion, then we expect the dispersion to be close to that in neutral conditions (with the same value of u_*) whenever the source height is much below $|L|$; in contrast if the source height is much above $|L|$ the dispersion will be influenced by stability. Also, as the plume spreads in the vertical, it might for example spread out of the neutral layer below $|L|$ and begin to be influenced by stability. In addition, the spread will vary with the plume height relative to h; this contrasts with older models where the only influence of h was often simply to act as a 'lid' on the dispersion. Horizontal dispersion is a bit more complex – mainly due to the fact that large convective eddies which fill the boundary layer can influence horizontal spread even at heights less than $|L|$ – but similar issues arise here too.

3 UNCERTAINTIES AND PROBLEMS IN MET PRE-PROCESSING

There are a number of causes of uncertainty in the results of met pre-processing. Firstly there is the nature of the input data. For example different anemometer designs have different starting speeds. Also some sites report only 10 minute average winds (so called 'spot winds') while others report hourly averages as well. However the main focus of this talk is uncertainties arising from the pre-processing procedure itself and this we will now consider.

3.1 Limitations of parameterizations

It is clear from the discussion in section 2 above that there a many different approaches possible, which will inevitably lead to different answers. However there is also a degree of convergence in the approaches used by the more modern models. For example the ADMS and AERMOD models use substantially the same approach and, for some parts of the process, the same algorithm. The main differences in the calculation of H and L concern the daytime Bowen ratio (the apportionment of available energy between the sensible heat flux and evaporation), the handling of dawn and dusk transitions, and the treatment of very stable situations. For boundary layer depth h there are two main differences. Firstly AERMOD can use radiosonde ascents to describe the vertical temperature profile while ADMS generally uses a default value for the lapse rate above the boundary layer (although the user can alter this value if information is available). Secondly different approaches are used in matching between the algorithms used for the extremes of stability and for neutral conditions. Apart from the radiosonde issue, these are all issues where either we don't know enough to decide on an ideal parametrization or where the physics is such that one needs more information than is available in routine observations in order to improve the parametrization (for example the moisture content of the ground and the state of any vegetation would be necessary to estimate the Bowen ratio more reliably). Hence the differences in approach and results should not be unexpected. The use of radiosondes should improve boundary layer depth estimates in principle, but of course the coverage is generally not as dense as one would like and, in the UK, is predominantly coastal.

3.2 Comparing newer models

Hall et al. (2000) compared several important models, ADMS, AERMOD (AERMET is the AERMOD pre-processor), and ISC. Their Figure 38 shows for example the met pre-processor outputs for boundary layer height and the reciprocal of the Monin-Obukhov length, $1/L$ (which is more convenient to plot than L because it varies continuously with stability). The scatter when the two models are plotted against each other is very marked. However it seems likely this scatter can all be explained by the differences outlined above. With $^1/_L$ there are horizontal and vertical lines of points at the very stable and very unstable extremes which suggest that the models are imposing limits on how stable or unstable conditions can get. Unsurprisingly the limits in the two models are different. For boundary layer depth there are a considerable number of points where AERMET predicts deep boundary layers and ADMS shallow ones. Although we have no direct evidence for this, it seems likely that this (as well and some of the differences in the $1/L$ plot) is due to differences in the day/night transitions. In AERMOD the flow is unstable whenever the sun is above the horizon whereas in ADMS the flow can be stable when the sun is very low in the sky. In addition the bulk of the boundary layer depths for AERMOD are greater than those for ADMS. Again we have no direct evidence for why this is, but it seems plausible that it is due to the default assumptions made for stability above the boundary layer. (As far as we can tell AERMOD makes no such assumptions but requires radiosonde input (see Cimorelli et al. 1998); however Hall et al.'s study did not use any radiosonde data and so it seems likely that the packaged version of AERMOD which they used had various defaults added to cope with the absence of radiosonde data). The point is that, although the pre-processors have a similar basis, when formulated as working codes some differences in outputs are likely.

The differences found by Hall et al. do look rather alarming at first glance. However it is important to realise that they are not always of great importance for the dispersion. For example turbulent velocities in convective conditions are sensitive only to $L^{1/3}$. Also uncertainties in boundary layer height may or may not be significant, according to the problem in hand. If the plume is close to the inversion, a change in the depth could affect plume penetration and reflection which may dramatically alter ground level concentrations downwind. However if the plume is well below the boundary layer height, there may be little effect seen at the ground. In fact when concentrations are compared, the general conclusion from Hall et al.'s work is that ADMS and AERMOD are more similar to each other than to older generation models such as ISC (as one would hope).

3.3 Comparing older models

Hall et al. suggest that the uncertainties may be a greater problem for the newer models than for the old. However when the older schemes are examined, it becomes apparent that considerable uncertainty exists there too; indeed it could be argued that the differences between models are greater for the older models. Farmer (1984) compared the frequencies of Pasquill categories that were diagnosed for a UK site by means of several different schemes. The frequency of Category A varied in the range 0.2-0.7%; B 3.5-7.0%; C 12-18%; D (day) 25-33%; D (night) 21-36%; E 7.1-21%; F 4.2-7.2%; and G 0.2-7.3%. Maes et al. (1995) compared the concentrations calculated from five models, four of which use stability classes: the models showed very different annual average concentrations to the North East of a 150 m high buoyant source. At 2-5 km, where concentration was largest, the values varied from one model to another by a factor of two or more. Also in the R91 report there are two procedures for estimating boundary layer depth – a day-time nomogram and a look-up table based Pasquill stability class; the results from these approaches can differ by more than a factor of 2. Finally, we note that Carruthers et al. (1994) showed examples where the centre-line concentration differs by a factor of 6 between two stability class models, R91 and ISC.

3.4 Other uncertainties

Clearly, users of dispersion models must take account of uncertainties which arise in dispersion calculations. Met pre-processing is one of several sources of uncertainty. Other causes include the choice of the meteorological data source or site, the period of data selected, and variations in emissions data including the heat and momentum of the release as well as the pollutant mass emission rate.

Urban conditions introduce further uncertainties into met pre-processing. For example urban effects can reduce the frequency and severity of stable conditions at night. Many current models don't account for this type of effect at all, and those that do often use a simple fix up to limit how stable conditions can get (e.g. by limiting the value of L).

4 EXPERIENCE FROM BIRMINGHAM URBAN EXPERIMENTS

The study was designed to build upon the earlier study of roof-top winds and street canyon dispersion in Leek, Staffordshire, UK by Manning et al. (2000). The purpose of the study was to increase our understanding of urban meteorology (cf Oke, 1987), with the ultimate aim of providing an improved assessment of urban boundary layer stability, which could lead to better air quality forecasts. Derwent et al. (1995) highlighted the significance of the urban heat store for air pollution, and Middleton (1998) used this in stability diagnosis.

Measurement campaigns were conducted in Birmingham, UK, during April/May 1998, January/February 1999, and July/August 2000. The location was Dunlop Tyres Ltd., a large tyre factory. Standard meteorological observations were also taken 8.5 km away at the Met Office station at Coleshill, just outside the city. Instrumented masts of 15 m, 30 m, and 45 m were set up at Dunlop Tyres, secured by guy ropes. At these heights, measurements were taken of turbulence, wind and temperature. Derived from these were sensible heat flux, friction velocity, roughness length and Monin-Obukhov length. Short-wave and long-wave radiation, pressure, rainfall and humidity were also measured. Since the exchange of heat between the atmosphere and the underlying surface is important, temperatures were also measured at 2 m above the surface, on the surface (grass or concrete) and down to 1 m below the surface. Results from the first two trials were described by Ellis and Middleton (2000a, 2000b, 2000c).

4.1 Roughness length

The roughness length z_0 is an important variable for dispersion models. Martano (2000) suggested a scheme to obtain surface roughness length z_0 from a single ultrasonic anemometer, whilst Grimmond et al. (1998) also discussed the derivation of aerodynamic roughness for urban areas. They cited an earlier review of 50 studies that estimated z_0 using different methods. Many are based upon an analysis of the city's geometrical properties or morphology. In the Birmingham data, Rooney (2001) has shown that the roughness length z_0 was in the range 0.5-2 m, fairly typical of values used in dispersion models for cities. He analysed the effects of wind direction, according to fetch and land-use type. His results are of especial relevance to this paper, for they show quite clearly the range of values for z_0.

4.2 Comparison of wind speeds

Other results from these experiments compared the winds measured in the Dunlop factory site at 15 m, 30 m, and 45 m, with the usual synoptic 10 m winds from the station at Coleshill. The effects of the city in retarding the wind were clearly seen when the winds within the factory site (F) were compared with the well-exposed synoptic site (S) at 10m. For example, Figure 1 shows the F_{15} v. S_{10} results from the 1998 trials. The experiments showed that:

- Wind directions were generally similar within and outside the urban area.
- Slower factory winds at 15 m: $S_{10} = 1.18F_{15}$ (1998) and $S_{10} = 1.24F_{15}$ (1999).
- Similar factory wind speed at 30m: $S_{10} = 0.95F_{30}$ (1998).
- Faster factory wind speed at 45m: $S_{10} = 0.82F_{45}$ (1999).

The roughness length was 0.5-2 m and the displacement height 2.5 m approximately.

Preliminary results suggest that the Analysis wind speed at 10 m from the Met Office's Numerical Weather Prediction (NWP) model was:

- NWP within 3% of synoptic 10 m wind speed $NWP_{10} = 0.97S_{10}$ (1999).
- NWP faster by 20% than factory wind speed at 15m $NWP_{10} = 1.22F_{15}$ (1999).

The latter results for the weather forecast model are as expected, in that the resolution of the model is not sufficient to resolve the city effects. As the 'New Dynamics' model* comes into use and resolution is increased, greater account will be taken of changes in surface properties.

4.3 Comparison of heat fluxes

The urban temperature was ~1°C greater than the rural temperature, with some time lag. The 2000 trial is now being analysed; it had the addition of a sonic anemometer at 15 m at Coleshill, in order to enable the heat flux and turbulence to be compared at the rural and city sites. In Figure 2 it was noticeable that the urban heat flux was on average much nearer to zero overnight than at the rural site i.e. conditions were neutral rather than stable in the July data studied. These results are being used to investigate ways of improving the diagnosis of atmospheric stability; they have also shown deficiencies in modelling the early morning transition at the factory site.

4.4 Summary of Birmingham work (to date)

These results convey an important message. Dispersion models (and indeed weather forecast models) cannot represent all the details of the underlying surface, and the flow over it. Considerable simplification of the problem is required. There is always some uncertainty to be associated with this inherent lack of resolution of surface inhomogeneities. However, whilst recognising that the hourly winds plotted at two different sites show mutual scatter, it is very encouraging that the graphs have a straight line form and the scatter is fairly well defined. This means that a population of wind values at Coleshill could be used to model a population of winds over the city by allowing for changes in roughness, whilst recognising there is uncertainty in the data that arises from the nature of the problem. Likewise the heat flux data point to simple schemes for adjusting stability diagnosis in the urban atmosphere.

5 CONCLUSIONS

There are uncertainties in any dispersion model. Users should allow for this when applying the results of dispersion calculations. One of the contributions to the uncertainty concerns the 'pre-processing' of the meteorology. Even when models make similar physical assumptions, differences can arise due to sensitivities (e.g. at the dawn/dusk transition) or due to default values assumed for unmeasured

* A new numerical weather prediction model being developed in the Met Office.

quantities. This is true of models using the old stability category methods, and the newer algorithms based upon Monin-Obukhov length. The complexities of turbulent flows over real surfaces are not fully described by any dispersion model, as models must simplify the world they represent.

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7 REFERENCES

Carruthers D. J., McHugh C. A., Robins A. G., Davies B. M., Thomson D. J. and Montgomery M. (1994). The UK Atmospheric Dispersion Modelling System: Comparison with data from Kincaid, Lillestrom and Copenhagen. Proceedings of the Workshop on Harmonisation within Atmospheric Dispersion Modelling for Regulatory Purposes, Manno, Switzerland, published by the European Commission.

Carson D. J. (1973) The development of a dry inversion capped convectively unstable boundary layer Q. J. Royal Meteorological Society, Vol 99, pp. 450-467.

Carson D. J. and Smith F. B. (1974) Thermodynamic model for the development of a convectively unstable boundary layer Advances in Geophysics, Vol 18A, pp. 111-124.

Cimorelli A. J., Perry S. G., Venkatram A., Weil J. C., Paine R. J., Wilson R. B., Lee R. F. and Peters W. D. (1998) AERMOD: Description of model formulation US EPA web site http://www.epa.gov/scram001/

Clarke R. H. (1979) A model for short and medium range dispersion of radionuclides released to the atmosphere National Radiological Protection Board, Report R91.

Derwent R. G., Middleton D. R., Field R. A., Goldstone M. E., Lester J. N., Perry R. (1995) Analysis and interpretation of air quality data from an urban roadside location in central London over the period from July 1991 to July 1992 Atmospheric Environment, Vol 29, pp 923-946.

Ellis N. L. and Middleton D. R. (2000a) Field measurements and modelling of urban meteorology in Birmingham, UK Met Office Turbulence and Diffusion Note No 268 Met Office, London Rd, Bracknell, Berkshire RG12 2SZ, UK.

Ellis N. L. and Middleton D. R. (2000b) Field Measurements and Modelling of Surface Fluxes in Birmingham, UK. Paper for COST 715 Working Group 2 Expert Meeting Antwerp, Belgium. To be published by EC.

Ellis N. L. and Middleton D. R. (2000c) Field measurements and modelling of urban meteorology in Birmingham, UK Presented at 3rd Symposium on the Urban Environment, AMS, in Davis California, 14-18 August 2000.

Farmer S. G. (1984) A comparison of methods of estimating stability category and boundary layer depth using routine surface observations Met Office Internal Report SITN 37 Met Office London Road Bracknell Berkshire RG12 2SZ.

Gifford F. A. (1976) Turbulent diffusion-typing schemes: A review Nuclear Safety, Vol 17, pp. 68-86.

Golder D. (1972) Relations among stability parameters in the surface layer Boundary Layer Meteorology, Vol 3, pp. 47-58.

Grimmond C. S. B., King T. S., Roth M. and Oke T. R. (1998) Aerodynamic roughness of urban areas derived from wind observations Boundary Layer Meteorology, Vol 89, pp. 1-24.

Hall D. J., Spanton A. M., Dunkerley F., Bennett M. and Griffiths R. F. (2000) An inter-comparison of the AERMOD, ADMS and ISC dispersion models for Regulatory Applications. Environment Agency Technical Report P362, ISBN 1 85705 340 0.

Holtslag A. A. M. and van Ulden A. P. (1982) Simple estimates of nighttime surface fluxes from routine weather data K.N.M.I. Scientific Report W.R. 82-4.

Maes G., Cosemans G., Kretzschmar J., Janssen L. and Van Tongerloo J. (1995) Comparison of six Gaussian dispersion models used for regulatory purposes in different countries of the EU. Int. Journal Environment and Pollution, Vol 5, 734-747.

Martano P. (2000) Estimation of surface roughness length and displacement height from single-level sonic anemometer data Journal of Applied Meteorology, Vol 39, pp. 708-715.

Manning A. J., Nicholson K. J., Middleton D. R., Rafferty S. C. (2000) Field study of wind and traffic to test a street canyon pollution model. Environmental Monitoring and Assessment, Vol 60, pp. 283–313.

Middleton D. R. (1998) A new box model to forecast urban air quality: BOXURB Environmental Monitoring and Assessment, Vol 52, pp 315-335.

Oke T. R. (1987) Boundary Layer Climates Second Edition Reprinted 1993 Routledge

Pasquill F. (1961) The estimation of the dispersion of windborne material The Meteorological Magazine, Vol 90, pp. 33-49.

Pasquill F. and Smith F.B. (1983) Atmospheric Diffusion, Third Edition Ellis Horwood

Rooney G. G. (2001) Comparison of upwind land use and roughness length measured in the urban boundary layer. Boundary Layer Meteorology, Vol. 100, pp. 469-486.

Seinfeld J.H. (1986) Atmospheric chemistry and physics of air pollution John Wiley and Sons

Tennekes H. (1973) A model for the dynamics of the inversion above a convective boundary layer J. Atmos Sci., Vol 30, 558-567.

Turner D.B. (1970) Workbook of atmospheric dispersion estimates Office of Air Programs Report AP-26 US Environmental Protection Agency

Venkatram A. (1980) Estimating the Monin-Obukhov length in the stable boundary layer for dispersion applications Boundary Layer Meteorology, Vol. 19, pp. 481-485.

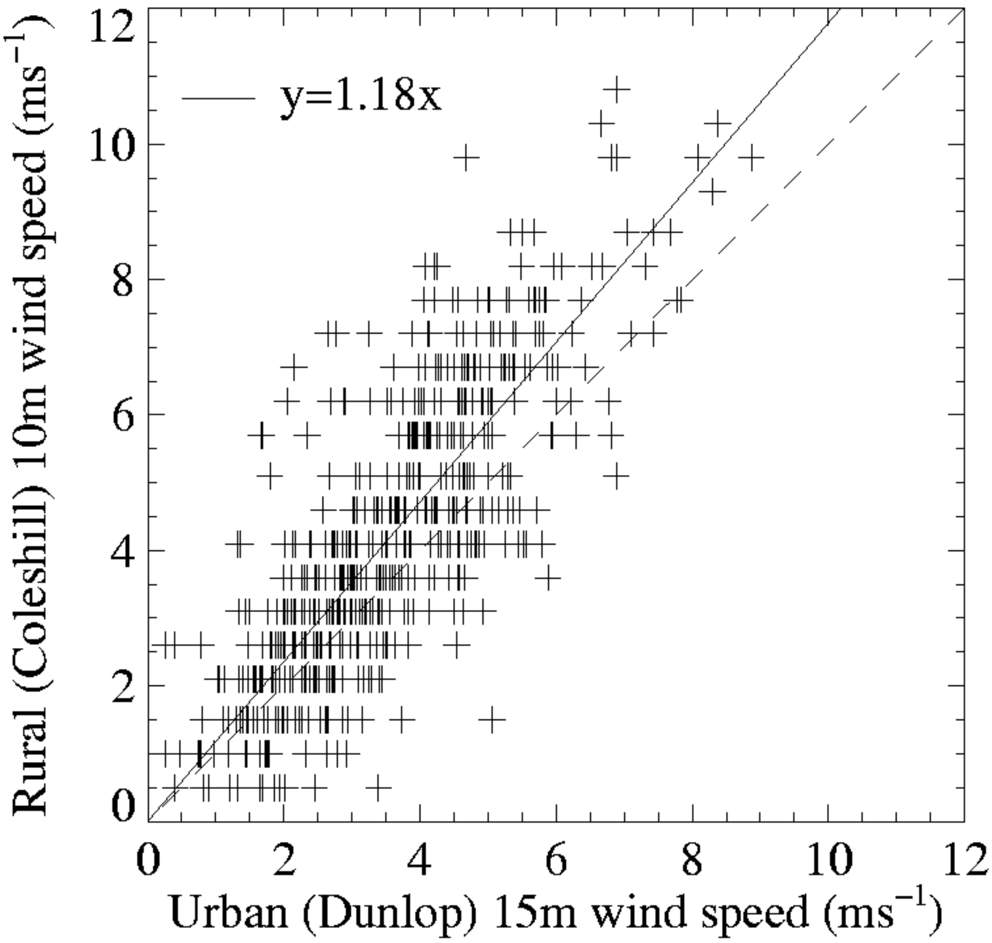

Figure 1 Synoptic data for wind speed at 10 m from the rural Coleshill station plotted against sonic anemometer mean wind speeds from 15 m mast at the Dunlop Tyres Ltd factory site. The regression slope (1.18) of faster synoptic wind shows the effects of roughness elements in the city slowing the wind at the Dunlop site. Results shown are for 1998 and are taken from Ellis and Middleton (2000a) Figure 6a. The synoptic winds appear in discrete layers because they are archived to the nearest knot (approximately 0.5 m s^{-1})

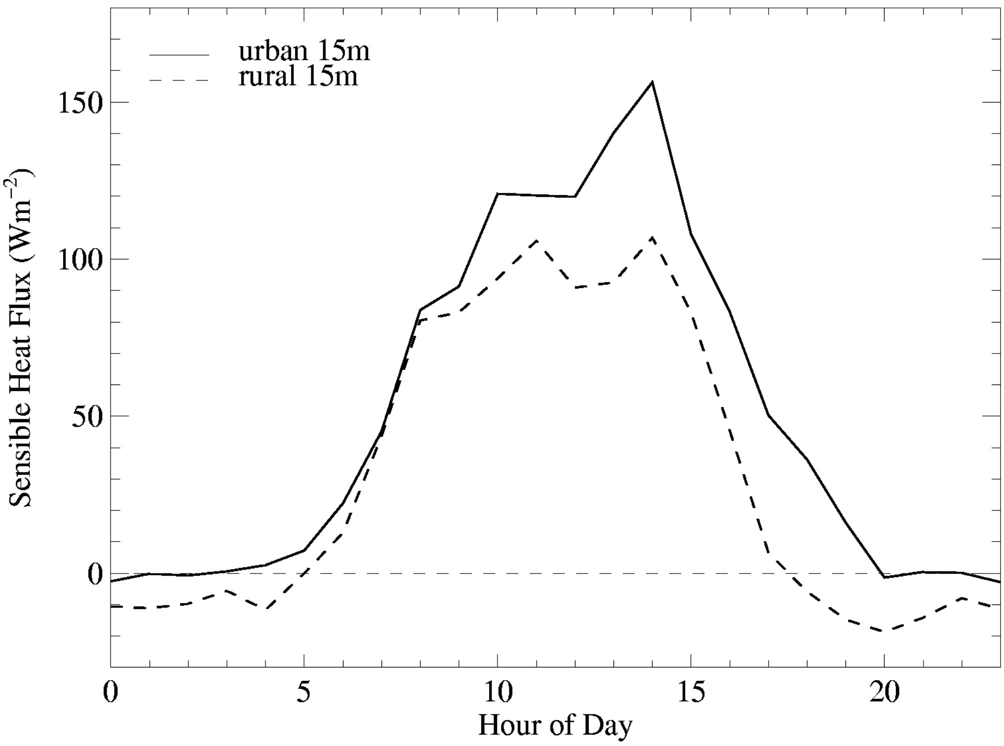

Figure 2 Comparison of sensible heat flux measured using sonic anemometers on 15 m masts at Coleshill synoptic station and at Dunlop Tyres Ltd factory site plotted by Nikki Ellis (personal communication). Results are averages by hour of day. Period was from 10:00 UTC on 7 July 2000 to 17:00 UTC on 28 July 2000. The urban heat flux at the Dunlop site was on average larger during the day. At night the average remained practically zero (neutral) whilst the rural synoptic site had a negative average heat flux (stable) reaching a minimum of -30 W m^{-2} at 20:00 UTC. The difference may be due to both the urban heat storage effect and anthropogenic energy.