Improved Air Quality Forecasting
Invest to Save Report ISB51-01.

MATCHING URBAN LIDAR DATA TO DISPERSION MODELS

By

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ABSTRACT

This report ISB51-01 was produced under Project 51 of the Invest to Save Scheme, or ISB. The aim of the study is to improve air quality forecasting through the use of lidar data with dispersion models that can be used to make such forecasts.

Air quality forecasts (i.e. a few days ahead) and Projections (i.e. for local air quality management, some years into the future) are essential tools in informing the public about poor imminent air quality, and in managing air quality. Air quality forecasting relies upon semi-empirical parameterisations within numerical models for the description of turbulent dispersion. Air quality Projections use the same models, together with information upon future emission control policies. There is scope to improve air quality forecasting through improved information about turbulence measured in an urban atmosphere. For example, turbulence is sensitive to the local stability and surface drag, yet current schemes for turbulence have very limited descriptors of urban effects on turbulence. This report describes the type of data required when turbulence parameters are calculated in a dispersion model. It describes the type of data flowing from a lidar in single mode, and a pair operating in dual mode. The report specifies how the lidar data should be processed to meet the objective of improving air quality forecasting through a better definition of urban turbulent flow fields. A new approach is presented that offers better air quality forecasts from using twin lidar, remote sensing technology to measure the atmosphere over an urban area. This report provides the information that will be used during the Project to define an optimum trade-off between the lidar scan dwell time and size of vertical steps.

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1. INTRODUCTION

Boundary layer measurements of turbulence are usually made using sonic anemometers. They act as rapidly responding point sensors of the three orthogonal components of velocity. By careful alignment of mast and sensor, disturbances to the flow are minimised. By processing the sonic data, turbulence quantities are derived. This Project explores the deployment of lidar systems in an urban environment. The present report investigates the requirements of the data processing software in order to meet the needs of the dispersion modelling community. Dispersion modellers and lidar operators have very different conceptual approaches to the same flow; a primary task in this work is to bring these conceptual models to a convergence of view. For example, a fast response on a sonic is very slow for lidar optics; spatial averaging along the lidar sample volume is quite different to temporal averaging applied to a sonic's time series record. This work is an essential first step in the Project. It defines the foundations for building consistency between lidar measurements and dispersion model met pre-processing. Software can then be developed to process the lidar signals for use with the dispersion modelling work.

2. DISPERSION MODEL VARIABLES

The aim of this Project is the improvement of air quality forecasting through the use of lidar data. To achieve this goal it is necessary to test and improve the way dispersion models describe the atmospheric stability and mixing depth. It will also be necessary to test some of the equipment and ideas in single mode operation as a first step. This means considerable attention must be paid to the flow field as represented in dispersion models i.e. the descriptions of mean flow and turbulent fluctuations. The mean flow advects material away from the source; the fluctuations ensure the spreading and dilution of the pollutant.

Meteorological pre-processors in dispersion models have to represent the complexity of the dispersing atmosphere in simplified form; this Project will seek to improve these descriptions where indicated by lidar field data. In improving the modelling of the underlying physics, the lidar data will thus be potentially improving any relevant dispersion model, and will not be inherently biased towards any one model. This wider usefulness of the lidar data will meet a contractual requirement.

2.1 NAME Lagrangian Model

The UK Met Office NAME model was developed as a long range model for radio-nuclides. It has been extended to the dispersion and chemistry of sulphur and nitrogen compounds, leading to aerosol formation via heterogeneous reactions. Currently its application to model the production of ozone is under development. Another version is being upgraded to handle dispersion at small scales, such as near buildings. An important application of NAME has been its use for air quality forecasting under the National Air Quality Bulletin System. Forecasts of air quality are released to public via the media such as the BBC.

The NAME model was described by two Met Office reports, by Ryall and Maryon (1996) and Maryon et al. (1999). Recently the plume rise scheme has been improved, as described in Webster and Thomson (2001). The model is a Lagrangian particle model. The particles are dispersed by advection and a random velocity, calculated according to the numerical weather prediction velocity field and the stability-dependent turbulence statistics respectively.

The key dispersion parameters are as follows:

Wind profile $u(p,t)$, $v(p,t)$, $w(p,t)$
Potential temperature profile $\theta(p,t)$
Standard deviations of wind fluctuations $\sigma_u(p,t)$, $\sigma_v(p,t)$, $\sigma_w(p,t)$
Turbulent kinetic energy dissipation rate $\varepsilon(p,t)$
Lagrangian integral time-scale $\tau_L(p,t)$

Where each variable is a field of values at the grid-point *p* and time-step *t*.

The components *u*, *v*, and *w* may be defined in terms of the co-ordinate grid directions within the numerical weather prediction model, travelling east, north and vertically (earth's radius) respectively. The co-ordinates may be latitude, longitude, and a height co-ordinate. The latter can be terrain following and/or combining pressure relative to mean sea level pressure or surface pressure. Thus *eta* co-ordinates may be found, as well as the more conventional metres. See the NAME documentation (above).
(NB: These co-ordinates might change with the New Dynamics Unified Model.)

Within the model it is worth noting that the turbulent quantities near to a large point source are changed by a buoyant, momentum plume, and could be the subject of another lidar study, once the techniques are proven in the field.

NAME requires a full three dimensional field of meteorological variables, whereas plume models such as ADMS, AERMOD, ISC, AEOLIUS, BOXURB etc. require meteorological data from a single point.

2.2 ADMS Plume Model

The key dispersion parameters are as follows (for further details see the ADMS Model Documentation from the CERC web site, or from D J Thomson at the Met Office):
Wind profile $u(p,t)$, $v(p,t)$, $w(p,t)$
Potential temperature profile $\theta(p,t)$: when not available the model assumes an overlying stable profile into which the convective layer grows as a function of solar heating of the ground during the day. Measurement of this growth of the convection would thus be valuable.
Standard deviations of wind fluctuations $\sigma_U(p,t)$, $\sigma_V(p,t)$, $\sigma_W(p,t)$ are calculated from the diagnosed parameters including Monin Obukhov stability in the surface layer.
Turbulent kinetic energy dissipation rate $\varepsilon(p,t)$
Lagrangian integral time-scale $\tau_L(p,t)$.
Since the model can calculate a wind profile according to local stability and local surface roughness length, any measurements of differences in wind profile or turbulence profiles between smooth and rough sub-strates would be valuable.
The ADMS met pre-processor software was developed by Dr D J Thomson in the Met Office.
ADMS is described by:
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.
See also web site www.cerc.co.uk

2.3 AERMOD Plume Model

The key dispersion parameters are as follows:

Wind profile $u(p,t)$, $v(p,t)$, $w(p,t)$
Potential temperature profile $\theta(p,t)$
Standard deviations of wind fluctuations $\sigma_U(p,t)$, $\sigma_V(p,t)$, $\sigma_W(p,t)$
Turbulent kinetic energy dissipation rate $\varepsilon(p,t)$
Lagrangian integral time-scale $\tau_L(p,t)$

AERMOD requires these inputs: date & time, sensible heat flux, friction velocity, convective velocity scale, temperature profile above mixing layer (potential temperature gradient), mixing heights for convection & mechanical motions, Lonin Obukhov length, surface roughness length, Bowen ratio (ratio of heat carried by to that by), albedo, wind speed and

direction at given height, ambient dry bulb temperature and its measured height. Upper air data are: date & time, height, direction, speed and temperature.
AERMOD is described under:
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/

2.4 US EPA Gaussian Models

The key dispersion parameters in these older models are as follows (for further details see the user manual for PCRAMMET available from the US EPA web site):
Wind profile $u(p,t)$, $v(p,t)$, $w(p,t)$ but expressed as mean velocity and direction measured at a well exposed anemometer at 10 m above level open ground (rather than as three orthogonal components over a range of heights). In the RAM model, a power law wind profile is calculated, with power selected by stability. Other variables in RAM are invariant with height.
Potential temperature profile $\theta(p,t)$ is only used within the model (using values set by stability class) for the Briggs' plume rise scheme.
Stability Class, according to Pasquill-Gifford-Turner approaches, diagnosed using latitude, time and date, sunrise/sunset, solar radiation or cloud cover, and mean 10 m wind speed. Then plume standard deviations are calculated according to stability class and distance downwind. The scheme in urban areas follows the Briggs' urban dispersion parameter curves, as described by Turner (1994).
Standard deviations of wind fluctuations $\sigma_U(p,t)$, $\sigma_V(p,t)$, $\sigma_W(p,t)$ are not computed in these models, but could be used to check stability class diagnosis via derived standard deviations of elevation and azimuth angles.
Turbulent kinetic energy dissipation rate $\varepsilon(p,t)$ is not used. Lagrangian integral time-scale $\tau_L(p,t)$ is not used.
PCRAMMET
Using as inputs: date, time, lowest cloud ceiling height (hundreds of feet), wind direction (tens of degrees), wind speed (both at 10 m), dry bulb temperature (nearest °F), and cloud cover (tenths), precipitation (type & mm). Upper air data needed are local minimum mixing height (morning) and maximum (afternoon) mixing height.
Giving as outputs: date, time, wind speed (m s^{-1}), temperature (K), Pasquill Gifford stability, and mixing height (m) for rural/urban cases; friction velocity, (m s^{-1}), Monin Obukhov length, roughness length, precipitation.

US EPA models are downloadable and described on:
US EPA web site http://www.epa.gov/scram001/

2.5 VDI Puff Lagrangian Model

The key dispersion parameters are as follows:

Wind profile $u(p,t)$, $v(p,t)$, $w(p,t)$
Potential temperature profile $\theta(p,t)$
Standard deviations of wind fluctuations $\sigma_U(p,t)$, $\sigma_V(p,t)$, $\sigma_W(p,t)$
Turbulent kinetic energy dissipation rate $\varepsilon(p,t)$
Lagrangian integral time-scale $\tau_L(p,t)$

2.6 Boxurb Model

The model uses a stability scheme after F B Smith, requiring:

Mean 10 m wind speed and direction.
Low, medium, high, and total cloud cover (oktas).
Sunrise/sunset, time and date.

It estimates an urban heat store, the urban sensible heat flux and diagnoses friction velocity and Monin Obukhov length. It also requires the urban mixing depth, or boundary layer, whichever is lower.

2.7 Aeolius Model

Model inputs are wind speed, wind direction, temperature and pressure.

The model assumes a neutral logarithmic wind profile to extrapolate 10 m well exposed wind speed and direction to the nominal roof height, and down to street level. Data to better define the wind speed profile down near and amongst buildings are needed.

The model also uses a simple horizontal vortex flow within the street aligned along the street axis; this also requires study.

AEOLIUS is documented and downloadable from met office web site: www.metoffice.com/environment

3. URBAN LIDAR PRINCIPLES

We now review the data products produced from pulsed lidars and the specification of typical lidar outputs. Signals are obtained in back-scatter from naturally occurring atmospheric aerosols whose particle size distribution (up to a few microns) is such that they faithfully follow the local flow-field. The systems produce sight-line Doppler, and incidentally back-scatter strength, with a range resolution determined by the laser pulse-length. In the TEA laser upgraded system this sight-line range gate will be of order 100-200m in length. The equivalent transverse dimension is much smaller, typically less than a metre, since the system is designed to emit a diffraction-limited collimated beam from its 15cm aperture. The laser pulse repetition frequency will be some tens of Hz, possibly as high as 100Hz, and it will be necessary to integrate the signal over tens to one hundred pulses, dependent on the atmosphere. Thus observation rates are likely to be about 1Hz. It is possible in principle to measure spectral width of the signal as well as the peak velocity; this is a measure of variability of sight-line velocity within the range gate and thus of turbulence on the gate scale. Such a width requires a higher signal to noise ratio than velocity estimation so is more limited in range and will not be a data product available initially. The maximum range capability is highly dependent on atmospheric transmission, which at this wavelength of 10 microns is strongly humidity dependent, as well as on the back-scatter strength. For the upgraded system under optimal conditions this maximum range is expected to be about 10km. There is also a minimum range of a few range gates caused by the finite detector recovery time after saturation by instrumental narcissus. Various scanning patterns can be applied to the emission as appropriate to the specific investigation.

Further basic literature on pulsed lidars and their operation is available from QinetiQ at Malvern. Two papers by Constant et al. (1989) and Vaughan and Forrester (1989) from Malvern describes the short range (single mode) CW lidar. The comparison with the MRU Cardington balloon borne probes is also described in the paper by Constant et al. (1989)

3.1 Doppler Effect

The Doppler effect is shown in French (1971) pp. 274-276 to depend only on the component of source velocity in the direction of the observer. Provided the distance from the source to the point of observation is large compared to the wavelength (which is true for the lidar: 0.2-10.0 km compared to10 μm) then successive wave fronts may be assumed parallel. For light of frequency f_0 emitted by a source moving with a velocity u having component $u\cos\theta$ towards the observer, the Doppler shifted frequency f_1 (French, 1971, Equation 8-14) is given by:

$$f_1 = \frac{f_0}{1 - \frac{u\cos\theta}{c}}$$

The velocity of light $c = f_0\lambda_0$. In the lidar, the light is returned by back-scattering from a moving particle, which behaves as a moving mirror, so the image point moves at twice the particle velocity. The velocity component in the Doppler expression is then $2u\cos\theta$. The Doppler frequency is:

$$f_1 = \frac{1}{1 - \frac{2u\cos\theta}{c}}\frac{c}{\lambda_0}$$

The change in frequency Δf:

$$\Delta f = \frac{1}{1 - \frac{2u\cos\theta}{c}}\frac{c}{\lambda_0} - \frac{c}{\lambda_0}$$

Rearranging:

$$\Delta f = \frac{c}{\lambda_0}\left(\frac{c}{c - 2u\cos\theta} - 1\right)$$

And

$$\Delta f = \frac{c}{\lambda_0}\left(\frac{2u\cos\theta}{c - 2u\cos\theta}\right)$$

The velocity of light is many orders of magnitude greater than the velocity of atmospheric aerosol, $c >> u$, and the Doppler lidar frequency shift reduces to:

$$\Delta f = \frac{2u\cos\theta}{\lambda_0}$$

Radiation with wavelength λ_0 is returned with a change in frequency Δf by aerosol particles whose velocity component toward the lidar is $V = u\cos\theta$:

$$\Delta f = 2\frac{V}{\lambda}$$

Equivalently, $\Delta\omega = 2\mathbf{k}.\mathbf{v}$ gives the result directly.

The observer in the lidar system is a combined telescope and heterodyne detection system. When the returning light is mixed by the detector with light of the original frequency, a signal is generated with this Doppler shifted frequency, Vaughan and Forrester (1989). With the CO_2 laser operating at $\lambda = 10.59$ µm, the Doppler shift is 189 kHz per 1 m s^{-1} shift in velocity up or down the direction of the beam (i.e. 1.89 MHz per 10 m s^{-1}). For any given direction of view, the lidar can only measure the velocity component V of the aerosol particle to or from the detector. Additional frequency mixing in the detector is required before the lidar can distinguish the sign of V, as discussed later (Section 3.2.3).

3.2 Inclined Lidar Beams

The lidar beam can be rotated in elevation and azimuth, and this is a powerful feature of the instrument. The Project is centred on the use of the long range pulsed Doppler lidars at Malvern and Salford, which use time gating to resolve range. The principles involved in using inclined lidar beams will be made clearer by reference to an earlier instrument known as LDV1, a focussing instrument with short range. Special cases are now discussed.

3.2.1 Aligned Down Wind

Consider a lidar beam that is inclined above the horizon, but is pointing down-wind, such that a vertical plane through the beam is parallel to the mean wind direction. Hence we see that a strong but uniform horizontal wind of say 10 m s^{-1} when sampled by a CW lidar pointing at elevation angle θ above the horizontal would return a signal with frequency $1.89\cos\theta$ MHz. At

30° elevation it is 1.64 MHz, at 60° elevation, this becomes 0.945 MHz; and for a vertical beam, with $\cos\theta = 0$, the strong horizontal wind would return a frequency shift of zero. In general the beam cannot be aligned identically with the mean flow direction.

3.2.2 Aligned In Any Direction

Consider a lidar beam that is inclined above the horizon at angle θ, but now the azimuth of the beam is changed by rotation to an angle ϕ between the vertical plane of the beam and the mean flow direction of our uniform horizontal wind. The above frequency shifts are now multiplied by $\sin\phi$ giving a combined frequency shift:

$$\Delta\nu = 2\frac{V}{\lambda}\cos\theta\sin\phi$$

If the vertical plane of the beam is at right angles to the mean flow, $\sin\phi = 0$, and there can be no frequency shift.

These results mean that for a uniform horizontal flow, the single beam CW lidar set-up will return a zero Doppler shift in two special cases where $\cos\theta\sin\phi = 0$:

1. If the beam is vertical ($\theta = 90\,°$)
2. If the beam is directly across the mean flow ($\phi = 0\,°$)

In practise the flow varies in time and space, and these special cases do not arise. However they do serve to illustrate the limitations inherent in analysing data from a Doppler lidar beam that is held at fixed elevation and fixed azimuth.

3.2.3 Up or Down the Beam Axis

Notice also that if the lidar beam were pointed in diametrically the opposite direction, against the flow, exactly the same Doppler shift in frequency would be returned by the heterodyne detector. Without some additional information the frequency shift does not reveal whether the wind was approaching or receding. It is usual to build some additional frequency shift into the instrument to avoid this limitation. This is done by imposing a frequency offset on the reference beam before combining it with the returning Doppler shifted back-scattered beam, as in Vaughan and Forrester (1989).

3.2.4 Scanning for Components

The continuous wave lidar LDV1 measures the wind velocity resolved along the lidar beam. Vaughan and Forrester (1989) describe a simple modification to determine the wind field in the vicinity of the measuring station. The principle lies in scanning the beam so that the elevation θ and azimuth ϕ are varied. With the wind components *u*, *v* and *w* resolved along the beam direction, their combined radial velocity (i.e. to or from the lidar) becomes:

$V_m = u\sin\theta\sin\phi + v\sin\theta\cos\phi + w\cos\theta$

As the beam is scanned a least squares fit to this function was used to solve for the horizontal wind speed $\sqrt{u^2+v^2}$. As discussed in Vaughan and Forrester (1989), various scan configurations can be used to optimise the solution for the problem at hand. Their paper should be consulted for more details of the geometry of scanning.

In effect, the radial velocity (to/from the lidar) is taken from a series of different locations at slightly different times, and a composite picture of the flow built up by assimilating the scans as being one sample data-set. The distance to the sampling point is varied by adjustment of the focal length of the telescope, thereby changing the distance along the beam, and thus the height above ground according to the elevation angle θ. Scanning makes it possible to gain more information on the wind components than that provided by the radial velocity component toward the lidar, which is all that a fixed elevation and fixed azimuth would provide. However the samples must cover a geometrically defined volume or surface described in the atmosphere, such as a cone, and they must cover a sufficient sampling period to accumulate enough scans for the least squares fitting process. Once a period of scans are taken and solved, provided the flow field was reasonably homogeneous in space and time, the mean wind speed and direction at each height may be used for evaluating the profiles of velocity versus height that appear in dispersion models. It is not clear how turbulence measurements can be derived however. The need to scan repeatedly means that a significant time must elapse (when compared with the time-scale of the turbulent eddies) before each estimate of *u*, *v*, and *w* can be made.

In terms of the current work, measurements by the LDV1 CW lidar up to 150 m will be very useful for gaining information about the mean wind versus height. More rigorous study is required to understand whether there is scope to derive turbulence quantities from the LDV1, and under what assumptions these quantities would be physically meaningful.

3.3 Dispersion model requirements

It is judged to be of the greatest priority to identify the height of the boundary layer at regular intervals throughout the field campaigns; apart from the flow information this is often associated with a marked reduction in back-scatter, which is readily measured by lidar. Whilst it may appear a simple variable, there is evidence that over cities multiple layers may exist in the atmosphere. Consequently calculations about vertical mixing are sensitive to the height of any temperature inversion. NAME has been found to be especially sensitive to this variable.

It is desirable to monitor the u, v, and w components of the wind over an extended time-scale so that fluctuation statistics, variances, and spectral power distributions can be calculated from the time-series. The height dependence of this parameter set is important because in an urban area, sources near the ground tend to dominate local air quality. The existence of an urban roughness sub-layer, extending from (or possibly including) the urban canopy (according to definition) up to several times building height also needs verifying. The layer is one where according to Raupach (1981), Roth (2000) and Rotach (1999) the Reynolds stress (and friction velocity) is not constant, but varies with height. It is very important to confirm the proposed existence of this layer during the experiments; it is dominated by the transport of momentum in a vertical direction towards the lower atmosphere, and is affected by the roughness, so it influences the mean wind profile, the variation of turbulent quantities with height, and the vertical mixing of pollutants. Urban dispersion is very sensitive to changes in the vertical mixing. However it is also expected that the x and y dependencies are significant when observing the effect of the urban-rural transition and in observing the effects of inhomogeneities in the horizontal plane. The horizontal homogeneity assumed in similarity formulations may not be valid over an urban area. Rotach (1993) has emphasised the importance of spatial averaging when seeking quantities characteristic of the roughness as a whole: he attempted this by averaging his Zurich data from different wind directions. With the lidar systems we have a unique opportunity to obtain spatial averaging over a wide area, which a point anemometer cannot do. However whether the lidars can provide data low enough to be truly within the urban roughness sub-layer will be a matter for experiment (it will be site dependent, according to field of view). This is why the parallel of LDV1 is seen as scientifically very valuable for these trials. For the present Project, we need to measure the components of the mean flow. Where possible it is desired to calculate other parameters of importance in dispersion, such as the eddy dissipation rate and the Lagrangian time-scale.

Bozier, Collier and Davies (2001) at Salford University have prepared a paper on their technique for evaluating eddy dissipation rate from gathered (single-mode) lidar data. Their methods will be reviewed in a subsequent report. Aerodynamic roughness length, zero plane displacement height, friction velocity, and profiles of velocity variance, momentum flux, and sensible heat, were derived. They concluded that their work did not fully use the 3-dimensional scanning offered by the lidar; the present Project seeks to greatly improve 3-dimensional coverage through joint deployment of two lidars.

A list of dispersion model variables appears in Table 3.

3.4 Scan Patterns

The lidar must be told where to look. The scan pattern defines the vertical elevation angles and the horizontal azimuth angles, along with the sequence in which they will be visited by the lidar beam(s) during field trials. There are many ways in which to configure the angular path traced out upon the celestial hemisphere when collecting data.

There is a need to choose the scan pattern to allow measurement of the three components of wind flow at different heights and times. For each height the dwell time needs to be sufficiently long to ensure the low frequency components of the flow field are adequately sampled, while the sample rate must be high enough to get the high frequency components. From the discussions three scan patterns were proposed:

i) Velocity azimuth display, at one or both lidar locations, which will give 360 degree coverage along an inclined sight-line. This will determine the general flow direction and observe features such as height dependent wind-shear for incorporation in (ii).

ii) Co-ordinated range height indicator (RHI) scans from both lidar locations in vertical planes roughly at right angles. The observation column of intersection can be moved around in plan on a circle whose diameter contains the two lidars. Using the data from (i) the column can be positioned so that u and v components of the wind are directly output by each system. Alternatively columns can be positioned in rural and urban locations to glean information on the urban/rural boundary. Evidence of the growth of the urban roughness sub-layer and the urban internal boundary layer from rural conditions will be valuable.

iii) RHI scans by both lidars in the vertical plane containing the two lidar locations to acquire the w component. There will be an inherent error in the accuracy of the measured vertical wind component inversely proportional to the sine of the scan angle to the horizontal. At lower heights, where the sine of the angle is smallest, the error will be greatest. Unfortunately it is at lower altitudes that the measurements need to be the most accurate. On the other hand at low elevation angles the height resolution arising from the finite pulse-length improves. A number of solutions to this problem were presented including:

- Pointing the lidar beams vertically. The main issue here is the minimum height that can be measured will be of the order of a couple of range gates and the vertical measurement would be made over the lidar station rather than at the sampling point mid away between the two lidars.

- QinetiQ Malvern has a short range, CW laser Doppler velocimeter (LDV) which could make the vertical measurements. However because LDV is a different lidar it has a different data recording system and again there is the issue of collocation between measurement points. Nevertheless this may prove to be a valuable instrument to deploy in an urban area where changes in the friction velocity are expected with height, especially in the urban Roughness Sub-layer, up to say 2 to 5 times building height.

Operation is likely to consist of patterns 1,2,3 following each other sequentially. It is noted that there is no need to precisely synchronise the scanning of each lidar provided that the data is stamped with time and pointing direction. The advantage of making the measurements with the twin lidars in this way is that only a carefully selected subset of a potentially huge database will need to be analysed to acquire the information to make an improved air quality forecast.

3.5 Temporal scale of measurements.

For the dispersion model it would be desirable to have frequency sampling at 10 Hz near the ground but with decreasing sampling frequency as height increases. The shorter the sampling time then the less stable the statistics: the optimum time for any given point is a few minutes. (Dispersion models normally need inputs every 15 to 60 minutes, currently from anemometers that sample at a 4 Hz rate). However, long dwell times at many points will give rise to an experiment duration so long that the atmospheric flow pattern cannot be regarded as stationary. Thus the duration of the dwell time impacts upon the spatial resolution of the scan pattern. It may be preferable to resolve the vertical column in sizeable steps to ensure adequate dwell times.

Suitable dwell times will not be known for the upgraded lidar until the modified systems have been tested. An advantage of increased output power is that the dwell time should be shorter than that quoted for the original pulsed lidar systems. Once this knowledge has been gained, the scan pattern can be accurately determined. It is the objective of a later Milestone report, MS4 due in September 2002, to relate these specifications.

3.6 Site location

The sites for the trial need to be identified early in the Project to allow time for permission to be obtained to work there and for the Met Office to run mesoscale forecasts at likely site(s). The separation of the lidars is dictated by the scale length of the urban/rural boundary and by the maximum operable range, which is dependent on atmospheric conditions. The prime site contender is still west London though other conurbations need to be considered. Two other contenders are Birmingham and Leicester (or Reading, of similar size). However it is uncertain if the latter conurbations would be large enough to generate the effects it is intended to measure. MRU Cardington have been asked to investigate possible trial sites; this is needed for the present Project, and to meet the needs of other Met office work. There are some advantages of economy if the site for lidar trials and other work were the same. It is not finalised whether all work needs to be at one city, although changing location usually entails extra work.

For a successful trial it has been identified that the urban lidar sites will need the following features:

- Separation of 5 to 7 km, dependent on the performance of the upgraded systems
- On the prevailing windward side of a large conurbation able to 'look' over the conurbation and detect rural urban transition.
- A line of sight between the two would be desirable or failing that a third independent object is visible to both lidars for alignment purposes
- There is no large hill or other significant obstructions such as trees near either of the lidar locations which would prevent operation at low elevation angles
- Usual field site requirements of power, permissions, security etc.
- Reasonably near (or containing) a suitable area for deployment of surface flux instrumentation on masts up to 45 m.

4. LIDAR SYSTEMS

4.1 Continuous Wave Lidar LDV1

This instrument is called LDV1 and was developed at Malvern in the 1980's and 1990's. It is described in Constant et al. (1989) and Vaughan and Forrester (1989). LDV1 transmits a continuous wave beam and measures the Doppler shift due to moving aerosol back-scatterers that appear in the focal plane of the instrument. Unlike a pulsed instrument, the spatial resolution is not determined by any time gating of the returning beam. Aerosols are present throughout the beam, but the signal is strongest near the focus, where the phase front is planar and best matches the reference beam (and by difference from which the Doppler beat or heterodyne signal is derived). Its maximum useful sight-line range is currently approximately 150m. This is set by the focussing behaviour of such a system, rather than laser power. The sight-line weighting function is of Lorentzian form centred at the focal region and of width proportional to the square of the range to focus; the FWHH at 100m range is 40m. The distance from the instrument to the sampling position along the beam is set by adjusting the telescope that focuses the outgoing beam and collects the returning light. The spatial resolution therefore decreases as the lidar is focussed to greater distances. The device is best set up with a fixed set of 5 sampling heights, and left to take samples every couple of minutes. The integration time is usually just under 2 minutes, sufficient for the near real-time data processing. A related instrument LATAS was used to measure wind shear in the vicinity of an aircraft during flight.

The instrument measures the velocity component in the direction of the beam; by adding an additional known frequency shift the sign or sense of movement toward or away from the telescope can be resolved. (Without such a shift, the Doppler effect only measures a difference in frequency but it is not possible to say whether the reference or Doppler shifted frequency was greater, so the sign of velocity is undefined.)

It is recommended that the LDV1 system be deployed during the field campaigns in the present Project. It will help to span the gap between the top of available masts (45 m) and the pulsed lidars. It will also help in cross-checking the various instruments. However the LDV1 has limitations arising from its focussing principle:

1. The FWHH range resolution at a range of 45m is about 10m, degrading to 200m at 180m range. At longer ranges there is effectively no downrange resolution.
2. It follows from the above that there is no point having such fine steps in range; a non-linear increment in focal distance will be more practical.
3. The included cone angle is 60 degrees for a beam at 30 degrees from the vertical, so the vertical distance resolution is improved by sec 30 (1.155).

A set of adjacent masts can provide measurements using fast 3-D sonic anemometers on the top of each mast at 45 m; 30 m; and 15 m. These will measure the three orthogonal components at each sampling point using fast sensors. They have minimal shadowing by the mast when mounted on the top. Allowing for the need to achieve overlapping measurements, and recognising the increasing focal uncertainty, the LDV1 might be set to the following nominal heights and focal range (1.155 times nominal height):

Decreasing the height by a factor 2/3 each time gives a logarithmic spacing:
Sonics at 13.33 m; 20 m; 30 m; 45 m.
These will fit comfortably on existing masts (15 m, 30 m, and 45 m), although some shadowing by the mast is implied at the 20 m and 13 m positions. These heights would also be useful in the search for Z_* where u_* has a maximum, starting within the urban roughness layer. The LDV1 can be set to correspondingly greater heights (allowing for beam inclination) and by virtue of its increasing focal depth gradually changing from a quasi-point measurement to a quasi linear measure over some distance. Philosophically, LDV1 provides a link between the mast sonics and pulsed lidars. It could be set to observe at these heights (measured vertically):
LDV1 vertical at 30 m, 45 m, 68 m, 101 m, 151 m.
The corresponding focal distances are1.155 times larger:
LDV1 focus at 35 m, 52 m, 78 m, 117 m, 175 m.
The focal range varies from circa 3 m at the lowest setting, to circa 10 m at the 52 m focus, to circa 200 m at 175 m.

A field validation trial at MRU Cardington could provide valuable experience in optimising measurement settings.

4.1 Pulsed Single Lidars

Pearson and Collier (1999) describe the principles behind the Salford and the Malvern high power pulsed lidars, and the current Project aims to upgrade these systems and eventually deploy them in dual lidar mode at an urban site.

Pulsed Doppler lidar operates near the quantum limit of detection of the infra-red 10.6 μm beam. The lidar detects the returns from spatially distributed aerosols and these returns become rapidly decorrelated, on a time-scale of micro-seconds. Their Doppler shift is such that a single pulse can suffice for the Doppler analysis. Pulse accumulation is useful to improve signal strength. High pulse rate allows accumulation of signal over 1 second, during which time the sampled motion of the atmosphere might be regarded as constant. Pearson and Collier (1999) then define the signal to noise ratio, the performance in Doppler resolution, and the basic hardware layout for the lidar. The system emits a pulse of duration about 1μsec, from which the velocity component is measured via the Doppler shift; atmospheric return strength, i.e. back-scatter coefficient, may also be deduced. Off-line processing detects the spectral peak (i.e. frequency and hence the central velocity) of the Doppler signal. In 1999, they reported 1 s of data took 3 s to process. Their paper shows examples of the signal strength decreasing with range, the velocity measured under different wind directions around Malvern, and the increased uncertainty in the peak velocity measured at the larger range where the return power is small. The possibility of measuring lee waves around a hill is mentioned.

4.1.1 Salford and Malvern Pulsed 10 μm Lidars

These are based upon the instrument described in Pearson and Collier (1999), but they are now being upgraded for use in this Project.

4.2 Dual Radar Mode Lidar

The velocity component along the beam provides a limited picture of the mean flow and the turbulence, even when a conical scanning is used as described earlier with LDV1. The scanning method with a single lidar assumes homogeneity and continuity of the flow. In order to reduce the number of unknowns, and to derive 3-dimensional information about the flow, two lidars will be set up a few kilometres apart. They will be programmed to scan a defined locus of overlap. The Essex University flow visualisation software will be written to derive additional information. It will calculate quantities used within dispersion models (reviewed in this report). The Project Implementation plan provides additional information.

4.3 Summary of Lidar Characteristics

Table 1 summarises the main characteristics of the continuous wave lidar LDV1, after Vaughan and Forrester (1989) and Constant et al. (1989). It also summarises the single mode pulsed lidar, and the proposed dual mode of operation using two pulsed lidars as in this Project. All these lidar options are contrasted with the characteristics of an ultrasonic anemometer. On studying their characteristics and regions of overlap and their differences, a new concept for the urban meteorological experiment emerges. The evidence collected here suggests that there is a great benefit to be obtained by combining these systems in the field. Possible measurement heights were mentioned earlier. We plan to combine the sonic anemometers on a mast up to 45 m with the CW LDV1 measurements starting near 45 m (for calibration inter-comparison throughout the trials) and rising towards 180 m, which approaches the first practical and measurable gate from the pulsed lidars. The latter then extend the measurements out to a planned range of perhaps 10 km. The actual heights achieved in the trial are however a function of the elevation angles. The combined deployment ensures that a measurement is achieved from 15 m up to perhaps 1 km, without requiring the deployment of a large tethered balloon, something that presents many problems in a large city. This range of measurement brings the data directly into the altitude range described by surface parameterisations within urbanised mesoscale modelling. It covers the full height range of urban dispersion models. Ideally, measurements will go up to the inversion at the top of the boundary layer.

Table 2 explains the interaction between sampling frequency, the number of samples needed for calculation of turbulence statistics and spectra, and the total elapsed sampling period that is implied. Using integral and even powers n of 2 to define the necessary number of sample values, possible sample sizes are then 2^n. Column 2 is calculated for powers of 10, 12, 14 and 16. Frequency of sampling f was set to values between 1 Hz and 100 Hz. The period T in column 4 is just $1/f$. The duration of sampling S (seconds) is then $S = 2^n T$ or more conveniently in minutes we have $S = 2^n / f \times 60$, and this is in column 5. Since integral sampling duration's are much more convenient, they are column 6. When the Table 2 was calculated, with smaller values of n, some high frequencies yielded sampling duration's that were too short (i.e. S < 1 minute) to be of use in turbulence statistics and were ignored here. Table 2 will be of value when designing the trade-offs between scanning and sampling duration subject to the constraint that the turbulence statistics should be consistent with the definitions implicit in dispersion models e.g. with regard to averaging in space/time.

4.4 Examples of Lidar Results

A number of papers have interpreted lidar results; a comprehensive review has not been carried out. Here we cite some examples by way of illustration.

4.4.1 Single pulsed Doppler lidar

Bozier et al. (2001) operated their lidar over Salford. Their range was roughly 50 m to 1 km. They claim that mean wind, cross wind component, vertical velocity, and momentum fluxes may be derived using wind velocity along the beam (radial). As implied earlier, scans in different directions are needed to achieve this. Through ensemble averaging the cross terms in fluctuations vanish; they arrive at $\bar{u}$, $\bar{w}$ from one plane of scans; $\bar{v}$, $\bar{w}$ from the other (orthogonal) plane. Their method followed the work of Gal-Chen et al. (1992) which is quoted in their paper. The errors in mean flow reported by Bozier et al. (2001) seem to be around 25% for horizontal, or 25-100% for vertical flow if using low inclination angles. As noted in Section 4.5.2 below, other methods of analysis exist, and the system is being upgraded. With the mean components determined, variances are derived. When a single beam is inclined, mass continuity must be invoked to estimate a vertical velocity. The upgraded Salford lidar will have a mirror for vertical viewing. The Malvern lidar can already do a full overhead scan. When operated with the beam pointing vertically, vertical velocity is obtained directly (Doppler shift being along the beam direction). The measurement errors will then be smaller. There is a trade-off between beam inclination and vertical resolution that is achieved. By obtaining spectra, fitting a -5/3 law, the kinetic energy dissipation rate was estimated. This is done at low altitudes where this type of spectrum is expected. The (local) friction velocity was derived directly from the turbulence components representing shear stress. Results for roughness length, displacement height, and fluxes of momentum and sensible heat were also presented. Considerable work will be needed during this Project however to establish whether the values obtained from lidars will meet the requirements of improving dispersion models for forecasting air quality.

4.4.2 Spatial analysis of lidar back-scatter maximum cross-correlation

Buttler et al. (2001) have used lidar to measure highly resolved three-dimensional flows over Barcelona. Their lidar (operated at 1.064 μm) measured the back-scatter signal from different directions. Their data processing used a maximum correlation technique to measure the movement of inhomogeneities in the back-scatter field (representing aerosol concentration inhomogeneities). The distance between successive but correlated signals when divided by the elapsed time yielded the mean flow velocity. This method of spatial correlation analysis has its origins in the use of remote sensing to measure movements of clouds, pack-ice and sea-surface velocities. Their results were compared with Doppler lidar wind profiles and with radiosonde data. Appendices in Buttler et al. (2001) give the theory. It was important to use a low elevation angle, so that the lidar back-scatter was dominated by the urban aerosol.

It raises the question whether the technique could be applied to the back-scatter signal from the Malvern lidar, to complement to its Doppler signal.

5. VARIABLES USED IN DISPERSION MODELS

5.1. Introduction

Sensors for boundary layer measurements fall into two categories, as discussed by Kaimal and Finnigan (1994) Chapter 6. In situ sensors on masts and on the surface and within the ground were deployed in the Met Office urban field experiments in Birmingham; see Ellis and Middleton (2000). In situ sensors are the method of choice for surface studies owing to their accuracy and resolution for such quantitative work, Kaimal and Finnigan (1994). Remote sensors offer increased range and a spatial scanning capability, but according to these authors (ibid) are constrained in minimum range and spatial resolution. They suggest that

> "Used in combination, however, the two types of sensor provide a more complete description of the flow field being studied than either of the two can provide separately."

This is a most important point. It emphasises the underlying scientific relevance to the Met Office research programme of work with the Salford and Malvern Lidars in an urban context. The hope is through the ISB funded Contract, new insights for the turbulent flow fields within dispersion modelling to forecast air quality will result. This report examines the purpose, definitions, spatial and averaging properties of the most important variables that are used in dispersion models. It then addresses the Lidar output data to examine ways of matching the one style of data to the possible variables. Some dispersion quantities will not be yielded up from the Lidar output, and must still be measured in more traditional fashion.

5.2 Wind Shear

Both wind-speed and direction change with height. This shear can have a dramatic effect upon a plume and dispersion models vary in their ability to represent such behaviour. They are often limited by the available input data, such as a mean wind at 10 m height, rather than a profile through the atmosphere. Radiosonde data or numerical weather model output can be used to obtain wind profiles. Wind shear can be modelled in typical conditions as is done in the ADMS model. The profiles for direction and velocity are functions of height to the boundary layer depth or base of lowest inversion according to the prevailing stability (defined below), viz:

1. In convective conditions there is much vertical mixing of momentum, and little change of wind direction and speed with height.
2. In stable conditions the vertical mixing of momentum is suppressed, and much sharper gradients occur.

The surface wind is usually ‘backed’ relative to the geostrophic wind in the northern hemisphere i.e. rotate anticlockwise from the geostrophic wind direction to get the surface direction. Typical observed angles (Hanna et al., 1982) between the surface wind and the geostrophic (free stream) wind are as follows:-

unstable 5-10°, neutral 15-20°, and stable 30-50°.

These angles represent a measure of the turning of the wind direction between the free flow above, and the wind near the ground. Wind velocity increases with height, gradually approaching the free-stream flow. The velocity profile depends on the stability (defined below), and as can be seen from measurements using a balloon ascent (or ‘radiosonde’), may not necessarily fit any smooth analytic function. However a convenient formula for the increase in wind speed at greater height in neutral conditions is the logarithmic wind profile

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

When plotting the graph, of $\ln(z)$ versus $u(z)$,

$$\ln(z)-\ln(z_0)=\frac{k}{u_*}u(z)$$

with intercept - $\ln(z_0)$ and slope $\frac{k}{u_*}$.

Here $u(z)$ m s^{-1} is the mean wind velocity at the height z m in a neutral flow near to the surface i.e. within the boundary layer. The friction velocity u_* (m s^{-1}) has been discussed above; it can be measured by fitting this profile for $u(z)$ to observed values of the mean velocity at different heights. Rotach (1993) has emphasised the importance of locally derived values of scaling parameters. Thus friction velocity would be derived from the turbulence data directly (method 2 below; Section 5.2.1). The von Karman constant *k* does not depend on stability and is usually set to 0.4 (dimensionless). In neutral stability, when this equation is assumed to apply, once u_* has been measured and *k* estimated, the wind profile can be used to estimate z_0. It is derived from the intercept of the graph.

Measurements of *u(z)* versus *z* in neutral stability will show reasonable scatter about a straight line of *u(z)* versus ln*(z).* The slope is u_*/k and the intercept where $u(z)=0$ is at a height z_0. The roughness length z_0 (as it is called) is used to describe the effect of how rough is the surface beneath the flow. The roughness length over the sea is typically 0.001 m, over lawn 0.01 m, uncut grass 0.05 m, and 1 m over wooded landscapes (Seinfeld, 1986, page 495). Roughness length is a measure of, but is much smaller than, the size of the roughness elements. Grimmond et al. (1998) have given a detailed comparison of methods to measure the roughness length for urban areas using anemometry. They also mention estimates based upon surface geometry i.e. ‘morphogenetic’ methods.
In stability’s other than neutral, more elaborate functions are used to describe the variation of the wind with height. These functions give deviations from the logarithmic wind profile according to whether the conditions are stable or unstable. To do this they depend upon the Monin Obukhov length *L* (a measure of stability in the surface layer) as in Seinfeld (1986).

For completeness, we note that the friction velocity may be derived from three different ways of measuring the horizontal shear stress, as discussed by Nemoto and Nishimura (2001):

1. Profile method, using logarithmic fit in neutral stability, requiring anemometers (cup or sonic) at several heights (on a suitable well exposed mast).
2. Eddy correlation method, where turbulence components are measured (e.g. by sonic, or hot-wire anemometers) and covariance calculated to give the Reynolds stress. This gives the local friction velocity (below).
3. Direct measurement of drag force on the surface using a drag plate. This method has a long pedigree, and requires a carefully constructed mechanical assembly with strain gauges or other force-transducers.

5.2.1 Local Friction Velocity u_*

The friction velocity may be derived from the vertical profile of wind speed versus height. This works over a uniform horizontal surface. Where the surface structure is more complicated, with roughness elements of differing sizes, as in an urban situation, Rotach (2001) has suggested that u_* is best determined as the local friction velocity. This is done via the eddy correlation of fluctuating wind components measured with an ultrasonic anemometer (or similar device); it represents a direct measure of the local turbulent stress via the covariance:

$u_* = \sqrt{-\overline{u'w'}}$

In parameterising the urban turbulence and wind profiles, it is the local friction velocity that should be used. Rotach (2001) describes:

- a maximum at the top of the urban roughness sub-layer, at $z = z_*$,
- to be constant in the urban inertial sub-layer, at $z_* < z <\approx 0.1z_i$,
- to decrease to small(er) values as $z -\ z_i$,

- and to decrease towards zero as $z - d$ or 0.

It would be a significant step forward to be able to determine in a field experiment the maximum in the local friction velocity u_* and its associated height z_* above an urban area. Likely values of z_* are thought to fall between 2 to 5 times (mean) building height; cf Rotach (2001), and Roth (2000). Here one must distinguish mean building height from the largest building height; a city often has a distribution of heights.

5.2.2 Urban Roughness Scaling Height z_*

Because modern ideas on the urban Roughness Sub-layer suggest that a new quantity z_* should be measured, some trials are also needed to see if the CW lidar instrument LDV1 could identify the height z_* at which the Reynolds stress $\overline{u'w'}$ (or the related friction velocity u_*) passes through a maximum value. If the LDV1 could yield a profile of either quantity this would be a significant achievement in developing our methods to parameterise z_*. The urban roughness scaling height z_* is thought to fall in the region 2 to 5 times building height. See Rotach (in preparation) and Roth (2000). If z_* proves to be a useful new concept to include in dispersion models, it would be valuable to measure it over several UK towns and cities. The lidars could be deployed in a series of campaigns for this purpose. It is not clear whether measurements near enough to the ground can be made using the dual pulsed lidars; however LDV1 and the masts could provide useful lower level data.

5.3 Measurements for Air Pollution Meteorology

In air pollution meteorology there are some measurements which are particularly relevant to the dispersion of pollutants. See also the AMS Workshop account by Hanna et al. (1977). Briefly, the instrumentation includes:

5.3.1 Anemometer and Wind Vane

All anemometers require careful siting; the choice of site may often be a compromise between the ideal and the possible. The cup anemometer and wind vane provide basic data on wind speed and direction near the surface. They should be properly set up according to well established rules of exposure, such as distance from obstacles. A paper chart (anemograph trace) gives some idea of the fluctuations in wind direction.

5.3.2 Bivane

The bivane is a delicately balanced vane designed to have two axes of rotation. It can respond rapidly to eddies in the wind, and yields the standard deviations of fluctuation of wind direction and elevation, □(□) and □(ϕ). Methods to calculate standard deviations of wind direction are discussed by Verrall and Williams (1982), with allowance for the discontinuity when direction passes through 360°. The direction standard deviations, □(□) and □(ϕ), enable plume spread parameters □$_y$ and □$_z$ to be calculated (see discussion of Practical Schemes, in Section A1.10 below). Formulae for the calculation were reviewed by Hanna et al. (1977). They caution that whilst the standard deviation of vertical wind direction □(ϕ) is a good indicator of stability, it is a difficult parameter to measure. The ultrasonic anemometer (below) can also be used to measure these standard deviations (and is more robust), although such equipment may be better employed to measure the Monin Obukhov length *L* (below).

Sedefian and Bennett (1980) compared several schemes for classifying the turbulence regime, including the use of □□(□). The standard deviation method to diagnose stability has been used in difficult situations, like remote valleys, e.g. Leahey and Halitsky (1973) who used bivanes. These standard deviations of direction measured in the field can be used to diagnose the Pasquill stability class. Strictly speaking (ibid), it is necessary to take samples for a period T and apply running means with the correct averaging time ($x/\beta u$ for travel distance x, wind speed u, and β=4) before the calculating standard deviations of wind direction. Some dispersion models such as ISC3, RAM, or CALINE will require the stability class as an input variable in order to describe the rate of plume spreading: the meteorological data must be processed before the model is run. When a model requires the direct input of standard deviation(s) of wind direction, but these are not available, sensible estimates may be made by referring to the values tabulated in Hanna et al. (1982) after the work of F Gifford. The values are as in Table A1.1.

Table 1 Estimates of standard deviation of wind direction (from Hanna et al., 1982)

Description	Pasquill Class*	Standard Deviation (at 10 m) σ_θ degrees
Very unstable	A	25
Moderately unstable	B	20
Slightly unstable	C	15
Neutral	D	10
Slightly stable	E	5
Moderately stable**	F	2.5

Notes:
* Pasquill stability category (or class) is defined later (Practical Schemes, Section A1.10)
** In the more stable conditions it is not easy to define the most appropriate value for the standard deviation of wind direction, largely because stable boundary layers may not be in equilibrium, and meandering of direction is seen in light winds. Larger horizontal standard deviations may be appropriate on occasion.

5.3.3 Ultrasonic or turbulence anemometer

An ultrasonic anemometer uses three axes of measurements, recording the effect of air movement on the time taken for sound waves to traverse a short gap. There are no moving parts and rapid fluctuations in the flow are recorded. A lower frequency ultrasonic anemometer is also available. Ultrasonic anemometers have possible advantages of robustness over bivanes or propellers for use in towns. (Alternatively three lightweight propellers are used to record the three components of the wind, but are less rugged, and are subject to friction in the slowest winds.) Valuable statistics of turbulence are obtained from these instruments. With the components u, v in the horizontal and w in the vertical, their standard deviations are σ_u , σ_v , and σ_w . Fast response measurements of temperature can be derived from the sound velocity in air, which the device also measures. Turbulence measurements in conjunction with rapid measurements of temperature and water vapour (the latter requiring an additional instrument) can be used to measure the sensible heat flux and latent heat flux respectively. The stability parameter L (Monin Obukhov length) may then be calculated from the measured turbulence data (cf Seinfeld, 1986). This method of diagnosing stability via L relies on the measurement of fluxes. Diagnosis of stability via vertical temperature gradients will be considered later (see under Lapse rate).

5.3.4 Upper Air Soundings

Radiosonde balloons are released from several sites in the country on a routine (synoptic) basis. Pressure, temperature, and dew-point (for humidity) define the state of the atmosphere as the balloon ascends. Its speed and direction (that of the winds aloft) are obtained by electronic means such as radar or satellite-based navigation systems, though a theodolite and rate of ascent can be used. Radiosonde data are invaluable in air pollution studies because they give information on the temperature profile, which is important in affecting the vertical motion of pollutants, and can be used to identify boundary layer depth.

5.3.5 Sodar and Lidar

As implied by their names these are radar-type methods, using sound waves or light waves respectively. The general idea is to send a pulse of energy aloft, then to record the time taken for it to be scattered back to earth. Sound is scattered by regions with a lot of temperature fluctuation, light by particles and aerosol. The change in frequency due to Doppler shift reveals velocity data about the upper air as well. Each method has limitations, such as range, which is sensitive to ambient conditions. The data can be very useful in characterising conditions at a new site. Lidar is useful to identify plume rise. Sodar is able to characterise the mixing depth (defined above).

5.4 Energy balance

The energy balance at the surface must be represented in dispersion models in order to diagnose the atmospheric stability. The urban energy balance is not usually done explicitly in existing models, although they all diagnose a stability. In order to make full use of the lidar data some characterisation of surface processes is required. This means the field experiments need to measure the radiation terms, sensible heat flux, and latent heat flux. The ground flux, or temperatures in the ground, are needed.

5.4.1 Insolation

The sun has a high surface temperature (about 5800 K) and emits mostly in the visible and short wave part of the electromagnetic spectrum (400-700 nm). Energy reaches the upper atmosphere at a rate of about 1380 W m^{-2} (the solar constant). Energy reaching the earth's surface (insolation) is in the range 0-1000 W m^{-2} at low-latitudes or in mid-latitude summer, but just 0-200 W m^{-2} in mid-latitude winter (McIlveen, 1992). The amount, which actually reaches the ground, depends on latitude, because this alters the path length through the atmosphere. It also depends on the season and time of day, as well as on local factors like cloud and precipitation. It is measured using an upward looking short wave radiometer. A downward sensor is also needed, because a significant fraction is reflected.

5.4.2 Surface Cooling

Any surface emits long-wave radiation according to its absolute temperature *T*, and its emissivity. For a black body (emissivity is 1), the maximum energy *E* emitted per unit time per unit area of surface obeys Stefan's law (Seinfeld, 1988, page 448)

$$E = \sigma T^4$$

where the Stefan-Boltzmann constant $\sigma = 5.673 \times 10^{-8}$ W m^{-2} K^{-4} .
Gases in the air, cloud droplets, the sea and the land absorb incoming solar radiation. They also emit radiation according to their temperature. Surface cooling is influenced by the radiation loss and the thermal properties of the ground. Ground loss of radiation is measured by a downward looking long wave radiometer; an upward looking sensor is also needed to detect black body radiation from clouds and the air.

5.4.3 Radiation and cloud

Cloud cover is routinely observed during each synoptic observation. The sky is divided into eighths and the amount of cloud in each layer and the height of the cloud base are reported in oktas and feet respectively. In the USA, cloud is reported in tenths. By day, without much cloud cover, the land receives more energy than it emits, so its temperature rises. By night, without much cloud cover, the land emits more than it receives so it cools down. Cloud can reduce the radiation reaching the ground during the day. By night, the clouds represent a surface that is warmer than a clear night sky, so their radiation reduces the rate of cooling of the ground. Daytime convective growth of the boundary layer and night time development of stable conditions by surface cooling are both affected by the extent of cloud cover. This is why the Pasquill stability Class is selected according to incoming solar radiation (day) or cloud cover (night), reflecting the importance of heating and cooling the air in driving or suppressing vertical motion Similarly, an estimate of the Monin Obukhov length is made within some dispersion models and the starting point could be cloud cover in order to calculate the radiation part of the energy balance (below). Ultimately the energy term needed is the sensible heat flux described below.

5.4.4 Energy Balance

The energy received at the surface from the sun during the day depends upon the latitude, time of year, cloud cover, and surface albedo. The albedo is the fraction of the incoming radiation that is reflected back, typically 0.29-0.34 (Strahler and Strahler, 1992). The absorbed energy depends upon (1.0 - albedo), so a fraction 0.7 of the energy striking the surface after passage through the atmosphere and clouds is absorbed in the land or ocean. The albedo is very site specific. In meteorology it is necessary to know how the energy is shared between the air, ground and water. The flow of energy for each method of transfer is described as a 'flux', which means the rate of transfer of energy per unit area per unit time. Semantically, 'net radiation', like 'sensible heat flux' or 'latent heat flux', is also a flux, although the word 'flux' seems to be dropped in common parlance when referring to the 'net radiation'. These energy terms in a simple energy balance at the surface are as follows.

Net radiation R_N is the incoming solar long wave radiation minus the outgoing radiation. Positive net radiation $R_N > 0$ means downward flux is larger than the upward flux, e.g. 60-80 W m^{-2} at noon, but -10 W m^{-2} overnight. Net radiation is strongly influenced by cloud cover.

Sensible heat flux *H* is the energy carried upwards (positive) or downwards (negative) by the turbulent motions of warm and cool air parcels, e.g. 280 W m^{-2} at noon, - 30 W m^{-2} overnight, where positive sensible heat flux denotes upward transport of heat. It is very important in controlling vertical mixing. Fluctuations w' in the vertical velocity and fluctuations θ' in the potential temperature (defined below) can be measured and their covariance calculated. With the vertical heat flux as *H*, density □, specific heat C_p

$$H = \rho\, C_p\, \overline{w'\theta'}$$

where $\overline{w'\theta'}$ is the vertical flux of temperature fluctuations i.e. $H / \rho C_p$.

Latent heat flux H_L is the heat of vaporisation carried upwards (positive) or downwards (negative) by the movement of eddies carrying water vapour.

Ground flux H_G is the rate at which energy is transferred between air and ground, being absorbed by the ground (positive) or lost from the ground (negative).

The simple energy balance at the surface is then (see Arya, 1988, pp. 10-12):

$R_N = H + H_L + H_G$

At night when winds are light, and skies clear, the net radiation is negative, causing cooling at ground level. Vertical mixing in the surface cooled layer is suppressed; mixing is poor. The air is stable. (The cooling can be sufficient for fog to form if the temperature goes below the dew point and wind is light). In the day, light winds and clear skies mean that the positive net radiation warms the ground rapidly, the temperature of air at the surface rises, and convection takes place. Vertical mixing is enhanced. The air is unstable.

5.4.5 Stability

Dispersion models for air quality use the idea of stability as a means of quantifying the amount of mixing or spreading of a plume. Stability is a measure of whether the atmosphere is likely to enhance or suppress turbulent mixing. The Monin Obukhov length is used, assuming the urban area to be horizontally homogeneous. See below after Lapse Rate.

5.5 Urban Surface Energy Balance

In urban areas, the normal terms in the energy balance apply. In addition there may be a change to the water cycle through irrigation or surface drainage. The storage of heat in buildings and concrete structures is very important, as is its slow release as night-fall approaches. Anthropogenic heating is also a contributor to the surface transfers of heat energy. Buildings and vehicles emit significant quantities of energy, perhaps attaining 10 % of solar input.

5.6 Temperature and Pressure

5.6.1 Ideal Gas Law

The familiar equation of state

$$PV = nR_0 T$$

applies to *n* (dimensionless) moles of gas at pressure *P* (N m^{-2}), absolute temperature *T* (K), volume V (m^3), and universal gas constant R_0 (8.31432 J mole $^{-1}$ K^{-1}). In meteorology it is convenient to use the form:

$$R = \frac{R_0}{W}$$

to define the specific gas constant *R* J kg^{-1} K^{-1} with molar mass *W* kg $mole^{-1}$. In ideal conditions, gas at constant volume which is given a change in pressure from P_1 to P_2 experiences a temperature change from T_1 to T_2

$$T_2 = T_1 \frac{P_2}{P_1}$$

In the atmosphere this must be modified as below (to give Poisson's Equation) because the volume of an air parcel is not constant as it ascends and expands.

5.6.2 Hydrostatic Equation

The pressure at the base of a column of fluid is the gravitational force due to the column above, per unit area. In the atmosphere, about 10^4 kg of air press down on each square metre (Lewis, 1991). If the pressure is *P* at height *z* with density the mass of a layer of depth *dz* per unit area is just *dz*. The change in pressure from *z* to $z+dz$ is *dP*:

$$dP = -\rho\, g\, dz$$

$$\rho = \frac{P}{RT}$$

$$dP = -\frac{P}{RT} g\, dz$$

Integrating from z_1 to z_2 gives the hydrostatic equation (McIlveen, 1992).

$$P_2 = P_1 \exp\left\{ -\frac{(z_2 - z_1)}{(RT/g)} \right\}$$

This equation applies to the vertical motion of air parcels, including pollutants. As a parcel rises, its pressure must decrease. The accompanying expansion causes a change in temperature, which may cause changes in state as well. In the atmosphere, measurements show that the absolute temperature *T* varies with height *z* from day to day and hour to hour. The hydrostatic equation enables the pressure difference across two heights to be calculated, assuming the layer of atmosphere has an absolute temperature *T.*

5.6.3 Potential Temperature

In a laboratory a useful standard state is 'STP', standard temperature and pressure, 0 Celsius and 760 mm mercury (1013.25 mb). In the atmosphere a more practical standard is to refer everything to a pressure of 1000.0 mb, 10^5 N m^{-2}. The compression of the air without any change of state (no condensation/evaporation) and with no heat gain or loss is a dry adiabatic process. The potential temperature is the temperature of the parcel when it has been compressed reversibly to 1000.0 mbar in a dry adiabatic process. However the air parcel can change its volume in the atmosphere, so we must allow for the work done by expansion, *P dV*.
Poisson's equation gives the new temperature in terms of the pressure change, allowing for the work done in the adiabatic expansion:

$$\frac{T_2}{T_1} = \left(\frac{P_2}{P_1}\right)^{R/C_p}$$

Define the potential temperature θ to be T_2 at $P_2 = 1000.0$ mbar. Then

$$\theta = T_1\left(\frac{1000.0}{P_1}\right)^{R/C_p}$$

where the air parcel at P_1 , T_1 , has potential temperature θ. The practical value of θ is that it converts T_1 into a temperature at the standard pressure of 1000.0 mb (10^5 Pa). This corrects the temperature for expansion due to the drop in atmospheric pressure with height. Note that *R* and C_p, should be in consistent units. For dry air, *R* = 287 J kg^{-1} K^{-1}, C_p =1004 J kg^{-1} K^{-1}, and the exponent R/C_p =0.286.

5.6.4 Lapse Rate

Lapse rate (Latin: lapsus, fall) is the rate of decrease of temperature with height. Formally

$$\Gamma = -\left(\frac{\partial T}{\partial z}\right)$$

A positive lapse rate $\Gamma>0$ is a decrease of temperature with height.
A negative lapse rate $\Gamma<0$ is an increase of temperature with height.
A zero lapse rate $\Gamma = 0$ has no change in temperature with height.

5.6.5 Environmental Lapse Rate

This is the lapse rate Γ_e as measured in the atmosphere, perhaps using a balloon and thermistor.

5.6.6 Dry Adiabatic Lapse Rate

This is a theoretical ideal which is of much use when discussing the ascent or descent of air. When an air parcel ascends without change of state, it is assumed to cool adiabatically due to the drop in pressure with height. The dry adiabatic lapse rate has the form

$$\Gamma_d = -\left(\frac{\partial T}{\partial z}\right) = +\left(\frac{g}{C_p}\right)$$

The numerical value of Γ_d is 9.8 $^\circ$C km^{-1}. The environmental lapse rate Γ_e is however often about 6.0 $^\circ$C km^{-1}. Ascending dry air would cool adiabatically at 9.8 $^\circ$C per km, whilst the surrounding air might actually get cooler at only 6 $^\circ$C per km. The environmental lapse rate is said to be subadiabatic when it cools more slowly than the dry adiabat and is called stable. It is called a superadiabat when it cools more rapidly and is called unstable.

5.6.7 Moist Adiabat

The atmosphere can contain significant amounts of water vapour. When considering the change of temperature with height, and the behaviour of air parcels as they move in the vertical, it is essential to include the effects of changes of state for the water. The dry adiabatic lapse rate described above is for the special case that no change of state occurs. This can be a useful description up to the height above ground at which temperature is cold enough for condensation to occur. Once above the condensation level (which is a good guide to the height of the cloud base) a moist or saturated adiabat must be used when discussing vertical motion of the air.

If condensation takes place, the latent heat of vaporisation of the water is released as heat and in turn adds to the buoyancy. The saturation adiabatic lapse rate is similar to the dry adiabat, but subtracts a term for the latent heat effect:

$$\Gamma_m = +\left(\frac{g}{C_p}\right) - \left(\frac{L}{C_p}\right)\left(\frac{\partial m_s}{\partial z}\right)$$

where the moist adiabat Γ_m contains an extra term which depends upon the latent heat *L* and vertical gradient of water vapour m_s. Here m_s is the mass of water vapour per mass of air at saturation, itself a function of temperature. When saturation occurs, the temperature decreases more slowly with height than for the dry adiabat. Hanna et al. (1982) give values for the moist adiabat ranging from 9 $^\circ$C per km (cold polar air) to 4 $^\circ$C per km (warm tropical air).

5.6.8 Using the Lapse Rate: Stability

A stable atmosphere arises when it is difficult for air to be moved up or down. In a neutral atmosphere the motion is not affected. In an unstable atmosphere, vertical motion tends to be enhanced. The stability of the atmosphere is dependent on the temperature lapse rate. If air is moved up a distance *dz* from a height *z*, then its temperature decreases from *T* at *z* to become $T - \Gamma_d \; dz$ at height $z + dz$. However the surrounding air will have a temperature $T - \Gamma_e dz$ at the new height. Three cases arise for dry air:

1. If $T - \Gamma_d \; dz < T - \Gamma_e dz$ then the air parcel is colder than its surroundings at the new level. It will be more dense and tend to sink back, suppressing the motion. Conditions are stable.
2. If $T - \Gamma_d \; dz = T - \Gamma_e dz$ then the air parcel cools at the same rate as its surroundings. It will keep the same temperature and the same density as the air around it. The motion is not affected there being no buoyancy difference, so conditions are neutral.
3. If $T - \Gamma_d \; dz > T - \Gamma_e dz$ then at the new level the air parcel is warmer than its surroundings. It will be less dense and tend to continue to rise, enhancing the motion. Conditions are unstable.

The conditions for stability are in Table A1.2.

Table A1.2 Stability conditions in terms of the environmental (i.e. actual) lapse rate.

Stability	Typical conditions	Environmental versus Dry Adiabatic Lapse Rate	Environmental Temperature Gradient	Description
Stable	overnight, clear sky, light wind	$\square_e < \square_d$	$\partial T_e/\partial z > - 9.8$ C km^{-1}	Subadiabatic
Neutral	overcast, windy, day or night	$\square_e = \square_d$	$\partial T_e/\partial z = - 9.8$ C km^{-1}	Adiabatic
Unstable	clear skies, strong sunshine light wind	$\square_e > \square_d$	$\partial T_e/\partial z < - 9.8$ C km^{-1}	Superadiabatic

These conditions also apply for downward motion (as is seen by replacing *dz* by - *dz*). Stable conditions suppress the displacement, whether it be up or down. Unstable conditions enhance the motion in either direction.
Note: in this context we refer to 'static' stability; dynamic stability has a different cause and meaning.

5.6.9 Inversion

There are several causes (see below) for a temperature inversion: radiation, subsidence, frontal (Seinfeld, 1986) page 462. They amount to cooling from below, or warming from above. Scorer (1968) contains photographs to illustrate the effect of atmospheric temperature gradients (or stability) on plume behaviours.
A subsidence inversion is the result of descending air (subsidence) being warmed by compression. The warming of elevated layers of air can be more than lower down and lead to an inversion. Anticyclonic situations have subsidence and a tendency for poor air quality.
Frontal inversion occurs at a front which is where air masses of different temperatures, pressures and humidities meet. In warm or cold fronts the warm air lies above a sloping wedge of colder air, causing an inversion.
An advection inversion occurs when warm air flows over a cooler surface. Warm air off the sea passing over cold land, gives a surface-inversion. If a cool sea-breeze is overlaid by a warm land-breeze, the result is an elevated inversion.
A radiation inversion occurs when on a clear night radiant heat is lost to space and the ground cools rapidly. Air near the ground is also cooled and can become very still; this is a surface-inversion. Air quality can be very poor when conditions are stable. Such radiation cooling causes the familiar ground frost.
In the convective boundary layer the height of the base of the first inversion layer from the ground is the inversion height z_i (Lenschow, 1986; Chapter 1, p. 6). This inversion layer is often quite conspicuous as a sharp increase in θ at some elevation when vertical profiles of temperature and humidity are plotted. It is especially informative to plot height z versus potential temperature θ.

5.7 Averaging the Planetary Boundary Layer

When the matching of the Lidar data to dispersion model variables is considered, it quickly becomes apparent that careful definition of "averaging" is required. The importance of averaging is discussed by Lenschow (1986; Chapter 1, pp. 6-7) In a turbulent boundary layer, there is a seemingly random three dimensional velocity field, along with random scalar fields of temperature, humidity, and pollutants. Averaged quantities are therefore studied. Averaging minimises the apparent chaos in the instantaneous values. In the daytime the averaging is especially important because large convective eddies can strongly bias short-term observations.

Lenschow (ibid) identifies several ways to average:

1. In turbulence theory, as an average over an infinite number of realisations, an ensemble average.
2. In numerical models (and we may add, especially in numerical weather prediction, NWP) a volume average is also used.
3. In experimental data processing, averages may be of the ensemble type, or taken over volume, area, line, or time.

A volume such as in NWP can be very large e.g. a layer of atmosphere some hundreds of metres deep and many kilometres in each horizontal co-ordinate. The NAME model for example running on mesoscale NWP data has its lowest level at approximately 10 metres and the grid is on a $\approx$ 10-12 km spacing; the volume is thus of order $\approx 10^9$ m^3. NAME would receive an updated value for the mesoscale NWP data in this cell every 1 hour. On global NWP data the grid scale jumps to $\approx$ 60 km and time spacing to 3 hours.

The NWP data represent time separated instantaneous sample values drawn from a continuously evolving field. They have implied spatial averaging over model cells; the cells vary in size due to the non-linear height co-ordinate and spherical grid layout. Furthermore, variables such as wind components, temperature and pressure are arranged on staggered grids to suit the numerical formulation of the forecast model. The evolution of the weather fields within the numerical forecast model is not 'seen' by the dispersion model NAME, which can only retrieve archived values; these are restricted by storage and other resource constraints (e.g. more frequent output from NWP means more computer time for the forecast as it runs) to 1 hour (mesoscale) and 3hour (global). A number of dispersion model variables are not output in the NWP data, and must be generated subsequently within NAME. This point will be discussed elsewhere in this report.

A Lidar may take a series of values from light back-scattered from the many aerosol particles within some sampling volume; the volume is approximately a beam cross-sectional area times the gated pulse length (speed of light times gate time window). For a 30 cm diameter beam with gated pulse length 150 m, the implied sampling volume in any instant would be of order $\approx$ 10.6 m^3.

Time averaging on an ultrasonic anemometer may be for 10 minutes or an hour, with velocity recording at a fixed point from 4 to 20 times per second (according to instrument and data logging set up). The sampling time may be designed to match air pollution and traffic recording, which are usually hourly, being reported at each hour-end, or for some time period 2^n seconds to suit spectral analyses. In some experiments both sampling periods may be required. In intensive field campaigns it is therefore best to store all the raw data for subsequent processing and analyses. In routine continuous observations, this may be too costly in resources and near real-time data reductions become essential.

5.8 Richardson number

Richardson derived an equation for the rate at which turbulent kinetic energy was produced (Sutton, 1953, page 152). His ratio is a convenient and dimensionless measure of whether the atmosphere was tending to create turbulence or to dissipate it.

$$Ri = \left(\frac{g}{\theta}\right)\frac{\left(\frac{\partial\theta}{\partial z}\right)}{\left(\frac{\partial u}{\partial z}\right)^2}$$

McIlveen (1992, page 304) points out that when *Ri* falls to less than 0.25, turbulence is expected to appear. Similarly, when *Ri* exceeds 0.25, turbulence is suppressed: conditions are stable. This criterion, or variants of it, are used to locate the top of the boundary layer depth in some models. *Ri* changes sign, becoming negative in unstable conditions because

$\partial\theta/\partial z < 0$. Similarly, *Ri* = 0 in neutral stability. See Seinfeld (1986) page 495 for further information.

5.9 Brunt Vaisala Frequency

The temperature varies with height, but this may differ from the adiabatic lapse rate. Consider an air parcel which is moved to a different height. Its temperature will follow the adiabat, but the surroundings need not. If the surrounding air in the environment at the new height has a different density then a buoyancy force exists. When the buoyancy force is opposite to the displacement, we have a restoring force at work. Simple harmonic motion is possible. The familiar equation for simple harmonic motion becomes:

$$\frac{\partial^2 z}{\partial t^2} = -N^2 z$$

where

$$N^2 = \left(\frac{g}{\theta_e}\right)\left(\frac{\partial \theta_e}{\partial z}\right)$$

Here θ_e is the potential temperature of the environment at the height *z* of interest. *N* defined as above is the angular frequency of the motion, the Brunt Vaisala frequency. For N^2 to have a real root, the oscillations with frequency *N* require stable conditions i.e. the surrounding air must cool at an environmental lapse rate (see above) which obeys $\Gamma_e < \Gamma_d$. The air must cool more slowly with height than the dry adiabatic lapse rate for there to be a restoring force.
The Brunt Vaisala frequency is of significance in air quality problems where there is the chance that vertical oscillations of the air could bring pollutants towards the ground. Simple models do not usually consider a plume whose height may oscillate with gravity waves to the lee of an obstacle such as a ridge. The wavelength (McIlveen,1992, page 360ff, 368ff) is

$$\lambda = \frac{2\pi u}{N}$$

for a wind speed *u* over the obstacle. The wavelengths can be 3-20 km.

5.10 Monin Obukhov length

The Monin Obukhov length *L* expresses the relative importance of shear (speed change with height) and convection in creating turbulence, and has the form of a length scale because it takes convective thermals some distance to accelerate to the point where they generate significant turbulence. This variable is used to describe the stability of the atmosphere in many field experiments. It is an essential parameter in modern dispersion models. Businger in Nieuwstadt and van Dop(1982) has defined the basis of *L*. In attempting to solve the equations governing the flow, some simplifying assumptions about the turbulence have to be made. Otherwise the mathematical problem of 'closure' is met, where there are more unknowns than equations, and which would make solution impossible. In the course of writing down equations for the so-called second-order closure, two terms for the production of turbulence kinetic energy appear. They are the production due to shear in the mean flow, and production due to buoyancy which tends to cause vertical motions. Obukhov sensed that the height above the surface where these two terms were equal might be a useful measure. Near the surface, shear production is dominant; above the height of the Obukhov length *L* buoyancy production is dominant. He solved for the height where the ratio of these terms was equal to one. He made the assumption that near the surface the wind speed obeyed

$$\frac{\partial \bar{u}}{\partial z} = \frac{u_*}{kz}$$

which on integration gives a logarithmic wind profile seen above. The friction velocity u_* m s^{-1} is measured via the wind speed fluctuations, using the shear stress as noted earlier.

He arrived at a definition of the Obukhov length in metres as

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

where $\overline{w'\,\theta'}$ is the vertical flux of temperature fluctuations (measurable using an ultrasonic anemometer with eddy correlation of fluctuations in temperature and vertical wind velocity), *k* is the von Karman constant (typically 0.4), and *g* the acceleration due to gravity. The minus sign has been introduced so that *L* has the same sign as the Richardson number. *L* and *Ri* are >0 in statically stable conditions but <0 in statically unstable conditions.

With the vertical heat flux *H*, density □, specific heat C_p, vertical flux of temperature fluctuation $H/\left(\rho C_p\right)$, and average temperature of air near the surface T_s (i.e. absolute temperature in degrees Kelvin), an alternative form is

$$L = -\frac{T_s u_*^{\ 3} \rho C_p}{kgH}$$

This is convenient when T_s, *H* and u_* can be estimated from routinely observed meteorological quantities (and the other quantities ρ, C_p, *k*, *g* are constants). Arya (1988) page 159 has a useful nomogram for *L* as a function of friction velocity and surface heat flux.

By definition *L* could have values in the range $-\infty \le L \le +\infty$. The sensible heat flux $|H|$ is usually of magnitude 10 to 120 W m^{-2} and the friction velocity u_* is always positive and its magnitude is often 0.05 to 0.25 m s^{-1}; the practical range of $|L|$ is 1 to 1000 m.

1. When conditions are unstable, with upwards positive heat flux *H, L* is negative.
2. When conditions are stable, with downwards negative heat flux *H, L* is positive.

According to Lewis (1991), *L* is positive and small in stable conditions with light winds at night. It is small and negative (about -10 m) on strongly convective days, about -100 m on windy days with some solar heating, and tends to infinity in the neutral case with purely mechanical turbulence. Processing of routine meteorological observations is used to estimate *L* for use by the ADMS or AERMOD models; likewise the Indic model can process measurements of turbulence, temperature and wind speed to obtain *L.*

Incidentally, the Monin Obukhov length can be used to obtain a dimensionless height □.

□=(z/L)

The height □ is the vertical co-ordinate used in Monin Obukhov similarity theory (proposed by these workers in 1954). It is used by the more modern dispersion models in formulae (cf Seinfeld, 1986) that describe as a function of height z the state of the atmosphere near the ground.

For Monin Obhukov in urban areas, remember that local values of the scaling variables are required (Rotach, 1999).

5.11 Convective Scaling Velocity

Near to the surface the flow is strongly affected by the frictional drag. Therefore as noted above the friction velocity is a useful quantity. In strong convection and well away from the surface another scaling velocity is required, because the flow is being driven upwards by the thermals. The convective velocity scale w_* has been used successfully when trying to match laboratory studies of convection in water tanks to observations in the atmosphere. (cf Tennekes, in Nieuwstadt and van Dop, 1982, page 59). The value of w_* is related to the mixed-layer height *h*, the height to which convection has grown, and the surface heat flux *H* (Arya, 1988, page 178) was discussed earlier. Here a temperature difference between the surface air and that higher up is driving the convection with a characteristic velocity w_*

$$w_* = \left(hg \frac{\left(\overline{w'\theta'}\right)}{\theta} \right)^{1/3}$$

where $\left(\overline{w'\theta'}\right)$ is the vertical flux of temperature fluctuations and can be expressed in terms of the sensible heat flux H via $H/\rho C_p$ as seen earlier.

The convective velocity scale w_* has meaning only in unstable conditions when there is upward movement of heat, just as the Brunt Vaisala frequency *N* has meaning only in stable conditions. These parameters have a special role to play in the more modern models when seeking to model dispersion, especially for elevated plumes that are well above the ground.

5.12 Height to Maximum Stress (Depth of Roughness Sub-Layer) z_*

This is the height above ground at which the friction velocity attains its maximum value. It will typically be a few times building height, as discussed earlier. It can be derived from the profile of Reynolds stress.

5.13 Roughness Length z_0

This is the measure of the roughness, as it appears in the wind profile. It can be derived from the wind speed data versus height.

5.14 Displacement Height d

This is a scaling height where the Reynolds stress and mean wind speed fall to near zero.

5.15 Mixing Height

This quantity has great importance for improving urban air quality forecasts. The mixing height is used in several dispersion models as the height from the ground through which pollutants may be expected to disperse. It is a useful concept but fraught with difficulty because in the real atmosphere it is not easily defined. On some occasions there is a well defined single inversion at some height above ground, acting as an effective lid to prevent vertical mixing of pollutants above it. In the case of large buoyancy or momentum driven plumes, the flow may partially or wholly penetrate the inversion, and then sinks back to its proper density determined height as the overshooting motion is damped and decays. On other occasions there may be a ground based inversion, so the layer of air near to the ground is stable, suppressing vertical spread. The possibility of multiple layers in the atmosphere should also be remembered, according to the interactions between the synoptic flow, local mesoscale effects like sea breezes or valley winds, topographic effects, and urban effects. Diagnosis of a boundary layer depth or mixing layer height will not be straightforward in such a situation.

6. SPECTRUM OF TURBULENCE

6.1 Introduction to Spectra

Early work to measure a spectrum of turbulence was inspired by G I Taylor and conducted in the wind tunnel using electrical analogue filtering circuits by Simmons and Salter (1938).

Spectra of atmospheric turbulence were the subject of active study in the 1950s. Panofsky and Deland (1959) provide a clear introduction:

- Power spectra describe the contribution of oscillations with particular frequencies f or wave numbers f^{-1} to the total variance of a variable.

- For turbulence, the variables are the velocities u, v, w in the three Cartesian directions x, y, z along wind, cross wind and vertical respectively.
- Velocity components when given as a function of time at a fixed point yield the Eulerian time spectrum as a function of frequency. Most spectral estimates are from fixed anemometers so represent Eulerian time spectra.
- Velocity components when given simultaneously at many points yield the Eulerian space spectrum as a function of wave number. With a many-point sample, usually measurements are all along a line, giving a one-dimensional Eulerian space spectrum. This is described simply as the "space spectrum" in their paper (ibid). This is the case with a anemometer mounted on an aeroplane flying through the turbulent boundary layer, Panofsky and Deland, 1959. However Lenschow, 1986, Section 4.2 page 10, suggests a moving probe like an aircraft measures time series data. NB: Perhaps this contradiction arises because the more modern instrumentation is much faster in acquiring data, so the time and space changes even as an aircraft flies may now be resolvable.
- When the values are fluctuations of the velocities of a given particle of air with time, they yield the Lagrangian space spectrum. Lagrangian space spectra may be studied by following individual air elements by means of a tracer, or when analysing the statistical properties of diffusion from a continuous source.

With regard to the measurement of spectra, and their practical, even engineering, application, Panofsky and Deland (1959) further explained that:

- Eulerian <u>time</u> spectra are related to easily measured atmospheric variables, but are least useful in applications.
- Eulerian <u>space</u> spectra are most relevant to the reactions of structures to turbulence, particularly aircraft.
- Lagrangian spectra are required for the prediction of dispersion.

Consequently they point out (ibid) that it is important to know if the various spectra are related, and whether point (fixed anemometer) and line (aircraft) observations can be combined to improve understanding.
NB: Lagrangian spectra are not discriminated here into time and space variants.

6.2 Time Spectra and 1-D Space Spectra

Quoting Panofsky and Deland (1959, p. 42):

> "G I Taylor (1938) postulated that time spectra should be equivalent to space spectra in the direction of the mean motion, provided that *t* is replaced by *x/U*, or the frequency *n* by *kU* where *U* is the mean wind speed and *k* the wave number. Taylor further stated that this transformation would be satisfactory provided the level of turbulence is low. Ogura (1953) and Gifford (1956) developed a theory indicating that the relation should be good even when the ratio of the turbulent fluctuations to the mean wind speed is of order one."

(Here *t* is time, *x* distance in the mean flow direction downwind, *U* mean flow velocity, *n* frequency, and *k* the wave number n^{-1}).

Panofsky and Deland (1959, p. 42) quote wind tunnel studies, and atmospheric work, by others, as experimental verification of Taylor's hypothesis for wavelengths up to 200 m. In an urban atmosphere, especially near or amongst the roughness elements, it is not certain if Taylor's hypothesis still applies. This could be significant when lidar data are being transposed into a form consistent with dispersion model met-preprocessors. Its significance or otherwise in the planned experiments with the dual lidars is unclear at the time of writing. However in discussion it is natural to assume some aspects of the flow are at least reproduced over some area, but this implicit assumption may not be true near the urban surface. If the turbulence fields vary rapidly in space it may constrain the use of simplifying assumptions in processing lidar data. The lidar scans on an arc as the beam elevation is changed; for a given lidar gate time, or distance up the beam, the sampling point is changing position horizontally across the city as the elevation is altered.

6.3 Conventions with Spectra

Spectra plot the frequency n (Hz *i.e.* s^{-1}) horizontally. Alternatively, they may plot wave number $k = n^{-1}$ (s) on the horizontal axis. When several orders of magnitude are involved, the logarithm of frequency $\log n$ or of wave number $\log k = -\log n$ is plotted on the horizontal axis. The power (J s^{-1}) associated with each frequency is plotted on the vertical axis.

6.4 Turbulent Kinetic Energy TKE

Kinetic energy is defined as

$$KE = \frac{1}{2}mv^2 .$$

Its dimensions are ML^2T^{-2} and units J.
If the mean velocity is $\bar{u}$ the kinetic energy of the mean flow per unit mass is

$$MKE = \frac{1}{2}\bar{u}^2$$

where the overbar implies that a mean value of the velocity is taken.

Turbulent kinetic energy (J) is defined as $\frac{1}{2}m\overline{u'^2}$.

Turbulent kinetic energy per unit mass (J kg^{-1}) is therefore given by

$$TKE = \frac{1}{2}\overline{u'^2} \equiv \frac{1}{2}\overline{\left(u-\bar{u}\right)^2} \equiv \frac{1}{2}\sigma_u^2$$

Here σ_u^2 is the variance of the velocity fluctuations[1], $\overline{u'^2}$.
The velocity fluctuations u' are measured by subtracting $\bar{u}$ from $u(t)$; and $\overline{u'^2}$ is found by calculating their standard deviation. Note that $\overline{u'^2} > 0$.
In meteorology, it is traditional to consider unit mass of air, implying $m = 1$.
Considering all three components of the flow, with variances of velocity fluctuations σ_u^2 , σ_v^2 , and σ_w^2 gives

$$TKE = \frac{1}{2}\left(\sigma_u^2 + \sigma_v^2 + \sigma_w^2\right)$$

Three methods of deriving the mean velocity $\bar{u}$ were analysed in the Appendix to Sakai et al. (2001):

1. Centred running mean (their equation A1) of an odd number of successive measurements centred upon the time t . There is no phase lag (their equation A2).
2. Block average, where an average over a time interval is used for $\bar{u}$. The block average is often used: $\bar{u} = \frac{1}{n}\sum_1^n u_i$ .The transfer function is their equation A3.
3. Linear trend removal where the mean flow is represented by a fitted straight line across the interval. A phase change is introduced (see their equation A4).

Sakai et al. (2001) describe transfer functions for each method. Consideration must be given to outliners in the data for $u(t)$ when filtering the data to derive $\bar{u}$. The work of Sakai et al. (2001) has significance in the present Project because it highlights the importance of low

[1] It can be shown that $\overline{u^2} = \overline{\left(\bar{u}+u'\right)^2} = \overline{\left(\bar{u}^2 + 2\bar{u}u' + u'^2\right)} = \bar{u}^2 + 2\bar{u}\overline{u'} + \overline{u'^2} = \bar{u}^2 + \overline{u'^2}$ since $\overline{u'} = 0$.

frequencies when measuring fluxes over a rough surface. Lessons drawn from their forest studies are relevant to our urban experiments.

6.5 Eddy Dissipation

This is the process whereby the kinetic energy of the turbulent energies is being degraded into heat energy, or thermal molecular motions. It arises through the effects of molecular viscosity (below) acting as a frictional force upon adjacent layers of fluid having a velocity gradient between them. Before defining the eddy dissipation rate ε, we explain molecular viscosity η and its associated kinematic molecular viscosity ν.

6.5.1 Coefficient of Molecular Viscosity η

The kinetic theory of gases reveals that when an ideal gas undergoes shear, so that one layer of gas moves past another, there is a frictional force. The force arises because molecules transferring between adjacent layers transfer momentum between the layers. The viscosity is a characteristic property of the fluid, being much smaller in a gas than in a liquid. The frictional force F increases with the velocity gradient u/d and with the area A of sheared fluid. The coefficient of viscosity η is expressed as the force per unit area per unit velocity difference between layers, with unit separation between layers:

$$F = \frac{\eta A u}{d}$$

Using mean molecular velocity v, molecular mean free path λ, density ρ, kinetic theory has:

$$\eta = \frac{1}{3} v \lambda \rho$$

In an ideal gas, the coefficient of viscosity derived from kinetic theory is independent of pressure P because λ depends inversely upon ρ. In real gases, the viscosity does vary with pressure, but only by small amounts, unless very high pressures are attained. The theory also shows that the coefficient of viscosity in a gas depends upon the mean free path λ, and hence upon the absolute temperature T of the gas.

As layers of fluid pass each other, work is done by the molecular viscosity. The work done being the frictional force times the distance of action. Within a turbulent flow, as the eddies twist and swirl past each other, velocity shear means that the kinetic energy associated with the turbulence is being constantly degraded into molecular motions i.e. into thermal energy, or heat. Molecular viscosity thus plays an essential role in turbulence by controlling this dissipation of the motions through smaller and smaller eddies into heat. This process increases the entropy of the molecules, so it is not reversible.

6.5.2 Kinematic Molecular Viscosity υ

The coefficient of molecular viscosity η (above) may expressed per unit mass of fluid, by dividing by the density ρ, and this gives the kinematic viscosity $\nu = \frac{\eta}{\rho}$. The kinematic viscosity (Lewis, 1991) will vary with ρ in the atmosphere.

6.5.3 Eddy Dissipation Rate ε

In the Navier Stokes equations for fluid flow, the velocity components may be separated into mean and fluctuating quantities (e.g. $\bar{u}$ and u' respectively). An equation for turbulent kinetic energy production can then be derived. Busch explains that the equation for turbulent kinetic energy includes terms which are dependent upon molecular viscosity. The shape of the turbulent energy spectrum is uniquely determined by the molecular viscosity ν and the turbulent kinetic energy dissipation rate ε. See for example Busch pp. 26-28 in Haugen

Workshop on Micrometeorology; also Batchelor (1956) is cited with reference to Equation 8.14. Thus, by measuring the spectrum of turbulence, and if in the inertial sub-range, it may be possible to estimate the corresponding value for ε. In this region the spectrum is expected to have the usual -5/3 behaviour.

Panofsky and Deland (1959; p. 47) divide the micrometeorological spectrum into 3 parts:

1. Frictional dissipation, with wavelengths of order ≈ centimetre.
2. Inertial sub-range, where turbulent energy is neither created nor destroyed. Theoretical predictions suggest the spectrum $S(n)$ varies as $n^{-5/3}$, where the frequency is n, but it also appears as $n^{-2/3}$ when plotted in the form $nS(n)$. Turbulence here is isotropic; the lateral wind components v, w have 30% more energy than the longitudinal component u.
3. Energy-producing range, where the form of the spectrum is not given by these authors (ibid) , but it must approach a non-zero value in the limit of zero frequency ($n - 0$). n

Since there are three orthogonal wind components, three distinct one-dimensional spectra can be envisaged. Within the inertial sub-range (isotropic turbulence) the spectra tend to be similar. The height of observation z above the ground is an important consideration when interpreting the spectra. Spectra are plotted with a vertical co-ordinate (ordinate) as $nS(n)/\overline{v_z}^2$, where multiplying by the frequency n means that the area on the spectrum between two frequencies represents the variance contributed by motions between them. Here $\overline{v_z}^2$ is a normalising factor, the square of the mean velocity recorded at height z. The quantity $S(n)$ is the energy spectrum for the turbulent velocity fluctuations. Because of the wide range of possible frequencies, the horizontal axis plots the logarithm of frequency. Plotting of spectra in meteorology is reviewed by Kaimal and Finnigan (1994, Chapter 2, pp. 37-39.)

It may seem surprising at first glance that an energy spectrum is drawn from the velocity fluctuations, or rapid deviations from the mean, until one remembers that velocity squared (multiplied by half the mass) is a measure of kinetic energy. The spectrum of turbulence reveals the fraction of the turbulent kinetic energy carried by motions at different frequencies, within the measured range. The highest measured frequency is limited to half the digitising frequency, and the lowest recorded frequency is constrained to have a period less than the duration of sampling.

Calculation of the turbulent kinetic energy dissipation rate ε from the spectral energy at the high frequency end of the spectrum is discussed in Kaimal and Finnigan (1994, Chapter 2, p. 36); see Term V in Equation 1.59 in, page 26 and their Chapter 2; pages 32-65. This high frequency part of the spectrum is called the inertial sub-range.

Kolmogorov had argued on dimensional considerations that the turbulent kinetic energy $E(\kappa)$ for wave-number κ may be proportional to $\varepsilon^{2/3}\kappa^{-5/3}$. The formula (Equation 2.3) in Kaimal and Finnigan (1994, Chapter 2, p. 36) is due to Kolmogorov (1941) and represents the one-dimensional spectrum using wind velocity component u in the wave number form using κ_1:

$$F_u(\kappa_1) = \alpha_1 \varepsilon^{2/3} \kappa_1^{-5/3}$$

Equivalently, using frequency n:

$$F_u(n_1) = \alpha_1 \varepsilon^{2/3} n_1^{5/3}$$

Here α_1 is the Kolmogorov constant of proportionality, between ≈ 0.5 - 0.6.

Hence

$$\ln F_u(\kappa_1) = \ln\left(\alpha_1 \varepsilon^{2/3}\right) - \frac{5}{3} \ln \kappa_1$$

and these spectra when plotted as $\ln(F(\kappa_1))$ versus $\ln(\kappa_1)$ will show in the inertial sub-range a slope of -5/3 with intercept $\ln\left(\alpha_1 \varepsilon^{2/3}\right)$. Since α_1 is known, ε can be determined from this intercept.

The turbulent kinetic energy dissipation rate ε is the amount of energy dissipated per unit time normalised by the mass of fluid. The dimensions of ε are $L^2 T^{-3}$ and units $m^2 s^{-1}$. The variable ε is used in dispersion models, especially random walk Lagrangian models like NAME, because under a typical velocity in the eddies, the smaller eddies dissipate faster, and the spreading of particles depends upon the sizes of the eddies. Near the ground, where strong shear is expected, ε increases rapidly (because it depends upon u_*^3/z).

It could be very useful if the Lidar could measure eddy sizes directly, but a potential problem lies in the relative sizes of the Lidar pulse volume (sampling space) and its relationship to eddy size (which could be greater or smaller). Ideally the three dimensional field of ε might be measured; furthermore direct measurements of ε in the vicinity of dispersing momentum and buoyancy driven plumes would be most valuable for the improving of air quality forecasting. The behaviour of turbulence in the urban Roughness Sub-Layer (RS), where u_* is not constant, may be somewhat different from that in the urban Inertial Sub-Layer, where u_* may be assumed to be constant with height. Consequently, measured profiles of ε will be most valuable. A comprehensive review of atmospheric turbulence over cities has been compiled by Roth (2000). Within the region up to some 2 to 5 times building height, the basis of several approaches to micrometeorological exchange processes is in doubt. He stresses the importance of urban turbulence measurements that are required within the urban canopy and its overlying urban boundary layer. The present Project offers scope to address some of these needs in a way that traditional tall masts on their own would not. In short, much more is to be learned if the masts, LDV1, and dual lidar were to be deployed together. Even on its own however, the dual lidar system offers a capability of spatial sampling for the turbulence that governs urban pollutant dispersion over a city that will be unique.

6.6 Integral Time-scale

Once the lidar has determined the turbulent kinetic energy dissipation rate ε, the other fundamental dispersion parameter is the integral time-scale of the turbulence.

Consider a point sensor that records a velocity $u(t)$ as a series of time or as space separated measurements u_i. Their mean velocity is

$$\bar{u} = \frac{1}{n}\sum_{i=1}^{n} u_i$$

This is an ensemble mean of n values. It represents a time average if the sensor is stationary and the flow moves past it.

It represents a spatial average if the points are spaced out by co-ordinate but are sampled together simultaneously in time. It could be regarded as a line, area or volume average according to the positions of the points sampled.

The turbulent fluctuation is

$$u'(t) = u(t) - \bar{u}$$

Their standard deviation has dimensions of velocity, units m $^{s-1}$, and from n values is given by

$$\sigma_u = \sqrt{\frac{1}{n}\sum_{1}^{n}\left(u_i - \bar{u}\right)^2}$$

(and when the mean value $\overline{u} = 0$, this would become the root mean square value).
The standard deviation of wind velocity fluctuations is a quantity of much importance in dispersion modelling. It is normally implicit that c_u, c_v and c_w are to be calculated by some semi-empirical formulation in the dispersion model, and are Eulerian quantities. That is, they are treated as if derived from three-dimensional anemometer data taken at a fixed point. The spreading of particles is calculated using the standard deviations of the fluctuations in the three wind speed components. Note that the calculation of standard deviation is well defined and is not assuming any particular shape for their probability distribution. The wind speed fluctuations need not be normally distributed; c_u, c_v and c_w can still be used to model dispersion.

The auto (i.e. self) correlation coefficient for a time lag τ is the dimensionless quantity

$$R(\tau) = \frac{\overline{u'(t)u'(t+\tau)}}{\sigma_u^2}$$

When there is a wide range of velocity fluctuations, the variance in the denominator will be large, and the auto correlation $R(\tau)$ will be small.
The integral time-scale of the turbulence is

$$T = \int_0^\infty R(\tau) d\tau$$

with dimensions of time, and units s.
The larger the eddies, the greater distance in space over which the velocity is correlated, and for a fixed point (Eulerian) sensor, the longer is the time for which the velocity values remain correlated. The larger the eddies, the slower does $R(\tau)$ decrease with increase in τ and the larger is its integral time scale, after Hanna et al. (1982; pp. 8 ff.). Since a large integral time scale implies larger eddies, this has implications for the dispersion of pollutants, and is used in the NAME model.

6.6.1 Lagrangian Timescale τ_L

This is given by $t_L \equiv T$ under the assumption that the velocity values $u_i(t)$ were taken moving with the flow (Lagrangian sampling). This implies that the sensor is moving with the mean velocity $\overline{u}$ and is merely sensing the fluctuations in velocity of fluid next to the moving sensor. Once the Lagrangian velocity fluctuations have been measured, their auto-correlation's with increasing lag-times may be evaluated, and from the integral of the auto-correlation with respect to time, the Lagrangian integral timescale is determined. Directly measured values of the time-scale would, with the profiles of ε, fill a very significant gap in the dispersion modelling repertoire. Both variables have an important influence upon the calculated pollutant concentrations.

When Lagrangian measurements of velocity fluctuations are not feasible, an approximate route may be to determine the Eulerian integral time-scale, and to then make some assumption about the relative magnitudes of the two types of integral time-scale. Then the Lagrangian might be estimated via the Eulerian; in practise this route is the more likely.

6.6.2 Eulerian Timescale τ_E

This is given by $t_E \equiv T$ under the assumption that the velocity values $u_i(t)$ were taken at a fixed point, not moving with the flow but allowing the flow to move past the sensor (Eulerian sampling)as if the turbulence were somehow frozen as in Taylor's hypothesis.

6.7 Integral Lengthscale

For completeness, we note in similar vein that the auto (i.e. self) correlation coefficient for a point separation s is the dimensionless quantity

$$R(s) = \frac{\overline{u'(x)u'(x+s)}}{\sigma_u^2}$$

When there is a wide range of velocity fluctuations, the variance in the denominator will be large, and the auto correlation $R(s)$ will be small.

The integral length-scale of the turbulence is

$$L = \int_0^{\infty} R(s)ds$$

with dimensions of length, and units m. NB This *L* is not the Monin Obukhov Length. The larger the eddies, the greater distance in space over which the velocity is correlated, and for a fixed point (Eulerian) sensor, the longer is the time for which the velocity values remain correlated. The larger the eddies, the slower does $R(\tau)$ decay.

6.7.1 Lagrangian Lengthscale

As before, with *s* replacing τ, this is defined using the velocities of particles following the flow.

6.7.2 Eulerian Lengthscale

As before, with *s* replacing τ, this is defined using the velocities of particles measured on a fixed point.

Table 1 Characteristics of sonic and lidar sensors for atmospheric turbulence and mean flow components.

	3-D Sonic Anemometer	Continuous Wave Single Lidar	Pulsed & Gated Single Lidar	Pulsed & Gated Dual Lidar
Principle	Ultrasound time-of-flight	Doppler shift light frequency	Doppler shift light frequency	Doppler shift light frequency
Sampling Frequency	Typically log at 4-10 Hz Preset options only?	Rotate scan at ½ to 2 Hz. One height: 10/minute Five heights: 16 profile/hour	Pulsing at ≈ 100 Hz	Pulsing at ≈ 100 Hz
Spatial Coverage	Point measurement (few cm path)	Focal Plane at Variable Focal Length	Beam Volume for gate open period	Beam Overlap Volume for gate open
Averaging	Temporal average of time series	Volume average of Lorentzian sightline weighting function	Volume average over gate volume (long truncated cone)	Volume average over overlap space (long truncated cone)
Primary Measurement	3 orthogonal components u, v, w; u down-wind	Velocity resolved along beam axis	Velocity resolved along beam axis	Velocity resolved along each of 2 beam axes
Inferred Measurement	De-trend; Mean u, θ; Standard Deviation via Taylor: 'frozen turbulence'	Scan/rotate inclined beam to infer 'frozen flow' over scanned space (continuity)	Scan/rotate inclined beam to infer 'frozen flow' over scanned space (continuity)	Solve 3-D flow field from 2 beams' overlap components & 'frozen flow' - vertical slice
Number of samples for half hour	2^{14} Samples at 10 Hz gives 27.3 minutes	Integrate for 2 minutes to infer flow; Rotate 15 scans/half-hour	Trade-off scan pattern with sampling	Trade-off scan pattern with sampling
Location of Measurements	Mast: 15m, 30m, 45m. Structures obstruct wind. Tethersonde restricted in urban areas.	Range by varying focus Absolute maximum range 200m	Range from 3rd gate range out to 10 km	Range from 3rd gate range out to 10 km but limited to overlap reach of both beams

Table 2 Sampling duration (minutes) and number of data values (2^{14} to 2^{16}) within sample, according to even powers of 2, and the frequency of digitising (Hz).

Power n	Samples	Frequency f	Period T	Duration S	Integer S
	2 To n	Hz	seconds	minutes	minutes
10	1024	1	1	17.06666667	17
10	1024	2	0.5	8.533333333	9
10	1024	4	0.25	4.266666667	4
10	1024	5	0.2	3.413333333	3
10	1024	8	0.125	2.133333333	2
10	1024	10	0.1	1.706666667	2
10	1024	16	0.0625	1.066666667	1
Power n	Samples	Frequency f	Period T	Duration S	Integer S
	2 To n	Hz	seconds	minutes	minutes
12	4096	1	1	68.26666667	68
12	4096	2	0.5	34.13333333	34
12	4096	4	0.25	17.06666667	17
12	4096	5	0.2	13.65333333	14
12	4096	8	0.125	8.533333333	9
12	4096	10	0.1	6.826666667	7
12	4096	16	0.0625	4.266666667	4
12	4096	20	0.05	3.413333333	3
12	4096	40	0.025	1.706666667	2
12	4096	50	0.02	1.365333333	1
Power n	Samples	Frequency f	Period T	Duration S	Integer S
	2 To n	Hz	seconds	minutes	minutes
14	16384	1	1	273.0666667	273
14	16384	2	0.5	136.5333333	137
14	16384	4	0.25	68.26666667	68
14	16384	5	0.2	54.61333333	55
14	16384	8	0.125	34.13333333	34
14	16384	10	0.1	27.30666667	27
14	16384	16	0.0625	17.06666667	17
14	16384	20	0.05	13.65333333	14
14	16384	40	0.025	6.826666667	7
14	16384	50	0.02	5.461333333	5
14	16384	64	0.015625	4.266666667	4
14	16384	80	0.0125	3.413333333	3
14	16384	100	0.01	2.730666667	3
Power n	Samples	Frequency f	Period T	Duration S	Integer S
	2 To n	Hz	seconds	minutes	minutes
16	65536	1	1	1092.266667	1092
16	65536	2	0.5	546.1333333	546
16	65536	4	0.25	273.0666667	273
16	65536	5	0.2	218.4533333	218
16	65536	8	0.125	136.5333333	137
16	65536	10	0.1	109.2266667	109
16	65536	16	0.0625	68.26666667	68
16	65536	20	0.05	54.61333333	55
16	65536	40	0.025	27.30666667	27
16	65536	50	0.02	21.84533333	22
16	65536	64	0.015625	17.06666667	17
16	65536	80	0.0125	13.65333333	14
16	65536	100	0.01	10.92266667	11

Table 3. A list of variables used in dispersion models, measured by a variety of techniques. (Not all of these are measurable by lidar.)

Variable	**Symbol**	**Notes**
Mixing height	z_i	Height of lowest inversion Beware multiple layers
Boundary layer depth	h	Rural value City has internal layer(s)
Wind speed	$u(z),\ v(z),\ w(z),$	Plume models use $u(10)$
Wind direction	$\theta(z)$	Plume models use $\theta(10)$ Beware blowing to/from.
Potential Temperature	$\theta(z)$	Plume models use the lapse rate $\frac{\partial\theta}{\partial z}$
Height	z	Or via pressure (NAME)
Pressure	$p(z)$	Or eta co-ordinate
Mean flow (space or time average)	$\overline{u},\ \overline{v},\ \overline{w}$	Via batch or running means of u, v, w
Turbulent fluctuation	$u',\ v',\ w'$	Via $u' = u - \overline{u}$ etc.
Reynolds stress	$\overline{u'w'},\ \overline{v'w'}$	For urban roughness sublayer and to get u_*
Turbulence	$c_\theta,\ c_\phi$	Std dev of wind angles for plume models cf Pasquill
Turbulence	$c_u,\ c_v,\ c_w$	Std dev of wind velocity component fluctuations
Turbulent kinetic energy	$TKE = \frac{1}{2}\left(\sigma_u^2 + \sigma_v^2 + \sigma_w^2\right)$	Kinetic energy of turbulent velocity fluctuations
Local Friction velocity	u_*	via Reynolds stress for urban roughness layer $u_*^2 = \left(\overline{u'w'}^2 + \overline{v'w'}^2\right)^{1/2}$
Log law for mean wind speed	$\overline{u} = \frac{u_*}{k}\ln\left(\frac{z-d}{z_0}\right)$	Velocity & stress approach zero at height $d + z_0$. Some authors use log law with implicit value $d = 0$.
Displacement height	d	Height adjustment for better log law fit at low heights.
Surface roughness length	z_0	Measure of overall aerodynamic roughness.
Urban roughness sublayer scale height	z_*	Height to which roughness affects turbulence statistics Height of maximum stress
Mean building height	$\overline{h}$	Must be known to interpret results: $z_* \approx \beta\overline{h}$; $\beta \approx 1\text{-}\ 5$. Need to establish β
Von Karman constant	k	$k \cong 0.4$ (dimensionless)
Kolmogorov constant	α_1	Value is $\alpha_1 \approx 0.5\text{-}\ 0.6$
Eddy dissipation rate	ε	Via inertial sub-range part of spectrum $\ln F(\kappa_1)$ versus

		$\ln \kappa_1$ since intercept is $\ln\left(\alpha_1 \varepsilon^{2/3}\right)$; $\alpha_1 \approx 0.5$ - 0.6
Lagrangian integral timescale	$\tau_L = \int_0^\infty R(\tau) d\tau$	Decay time scale for auto correlation coefficient $R(\tau) = \frac{\overline{u'(t)u'(t+\tau)}}{\sigma_u^2}$ for lag τ & velocity variance σ_u^2
Sensible heat flux	H (sometimes Q_H)	Urban or rural heat carried up by convective eddies.
Latent heat flux	λE (sometimes Q_E)	Heat carried upwards by water vapour flux E and latent heat λ.
Ground heat flux	G (sometimes Q_S)	Heat transfer flux into Ground G or Soil S
Building heat fluxes		Analogous to G but into building fabric; plays role in urban heat store effect.
Anthropogenic heat flux		Released from energy use.
Net incoming all-wave solar radiation reaching surface	R or Z (sometimes Q^*)	After passage through cloud
Precipitation		For wetness of surface
Temperature	$\theta(z)$ or $T(z)$ where z is at several altitudes and also below ground level.	Sensor response: Lapse rate: slow. Sensible heat flux: fast. Ground: slow.
Mean temperature	$\overline{\theta}$	Batch/running mean of θ
Temperature fluctuation	θ'	From $\theta' = \theta - \overline{\theta}$, usually via sonic or fast thermometer
Flux of temperature fluctuation	$\overline{w'\theta'}$	Measure w' and θ' rapidly at same point
Flux of sensible heat (above)	$H = Q_H = \rho C_p \overline{w'\theta'}$	Measured via vertical turbulent fluctuations
Brunt Vaisala frequency (stable lapse rate)	$N = \sqrt{\left(\frac{g}{\theta_e}\right)\left(\frac{\partial \theta_e}{\partial z}\right)}$	Associated with buoyancy restoring force (stable lapse rate)
Convective scaling velocity	$w_* = \left(hg\frac{\left(\overline{w'\theta'}\right)}{\theta}\right)^{1/3}$	Associated with speed of convection (unstable)
Monin Obukhov Length	$L = -\frac{\theta u_*^3 \rho C_p}{kgH}$	Stability scale height above which convection dominates turbulence
Dimensionless height	ζ	Defined as $\zeta = \frac{z}{L}$

7 Conclusions

1. This report describes the variables used in the met pre-processors of dispersion models. Variables are listed in Table 3. Variables of particular importance for this study to improve air quality forecasting include:
   - Mixing height (possibly multiple layers),
   - Day/Night transition in stability and turbulence variables,
   - Mean wind profiles & heat flux,
   - Existence & properties of urban roughness sub layer (height z_*), which affects wind and turbulence profiles, and the rural-urban transition,
   - Measurements in summer anticyclonic conditions are needed to study the contrast between strong daytime convection and night time conditions,
   - Measurements in winter anticyclonic conditions are needed to study the effects of temperature inversions.
2. The Project will have the greatest scientific benefit if the lidar trials can be accompanied by surface flux observations on masts and slow ascent radiosonde releases, together with interpretations using high resolution mesoscale meterological modelling on a case by case basis. Such parallel studies are however not a part of the ISB51 Project, and would need separate resourcing.
3. The report discusses recent lidar implementations at Salford and Malvern; both are to be upgraded, and new data processing and visualisation software developed.
4. The principles of using lidar data to derive mean flow and turbulence quantities was illustrated. Although a full literature review of lidar remote sensing was not attempted, this review demonstrates the ability of lidar to measure the relevant wind flow parameters with sufficient accuracy to meet the requirements of ISB-51. Further work is required to fully specify the software and field scanning/sampling programme.
5. The Project represents a unique opportunity to gain flow and turbulence data using lidar remote sensing over a city for the improvement of dispersion models that are used in air quality forecasting. Current experience in the Met Office shows that the required measurement heights and spatial sampling over a conurbation can only be achieved through these remote sensing techniques. An optimal approach combines both surface and remote sensing.

The report is the first Milestone in the ISB Urban Lidar Project No 51.

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