Turbulent transfer relationships over an urban surface. I: Spectral characteristics

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SUMMARY

A comprehensive study of atmospheric turbulence over an urban surface has been carried out. In this Part I of the study the spectral characteristics are analysed. Results involving all the important atmospheric variables, measured at two heights \( z' / z_0 = 21 \) and \( 37 \) under unstable conditions, are presented, normalized by the respective variances/covariances, and within the Monin–Obukhov similarity framework. Although the shapes and location of the peaks of the urban spectra/cospectra are in good agreement with homogeneous surface layer data, the analysis of the normalized dissipation rates and spectral correlation coefficients reveals differences which can be attributed to the rough surface. For example, the dissipation of turbulent kinetic energy is relatively small, perhaps owing to an increase in the transport and export of locally produced turbulent energy; there is a very efficient transfer of momentum which it is suggested may be a result of wake production associated with bluff bodies, and it is found that the transfer of heat and moisture are dissimilar.

1. INTRODUCTION

(a) Research context

In the ‘constant’ flux layer the standard micrometeorological theory has been shown to apply, and turbulent fluxes and statistics can be described and evaluated using methods such as the aerodynamic, Bowen ratio-energy balance or the Monin–Obukhov similarity (MOS) theory. The validity of these frameworks, however, must be questioned when used in the lowest few metres over surfaces that are rough (in the aerodynamic sense) and inhomogeneous (with respect to the source and sink distribution of turbulent fluxes), because several of the assumptions underlying their derivation are probably invalid.

Some progress has been made in the description and parametrization of turbulence in rough but horizontally homogeneous environments such as the surface layer over forests (e.g. Raupach 1979; Garratt 1980; Baldocchi and Meyers 1988; Gao et al. 1989), tall crops (e.g. Raupach and Thom 1981) or bushland (Chen 1991a, b). But the boundary layers over inhomogeneous and rough surfaces in general are not well understood, one reason being the lack of appropriate experimental data which could be used to develop a consistent theoretical or conceptual framework for the turbulent exchange. This raises several questions regarding the measurement and analysis of turbulent fluxes over such surfaces and about their statistics: for example, the sampling strategy to be employed (in space and time), the height of the instruments, the representativeness of the measurements etc.

Our overall objective in this research has been to investigate the properties and dynamics of turbulence over a built-up surface. Considering that most boundary layers of human interest occur over ‘non-ideal’ (in the micrometeorological sense) surfaces and, in view of the rapid increase in urban development with its associated air pollution problems and changes in local climates, such research is of considerable importance. The focus in Part I of this study is on the spectral characteristics: Part II (Roth 1993) will deal with the analysis of the integral statistics.

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(b) Concepts and objectives

It is convenient to subdivide the atmosphere over a suburban surface into two parts. In the lower part, known as the roughness sublayer, the microscale flux fields within and just above the urban canopy are highly variable in space and time, and the flow fields are affected by direct interaction with individual and/or a combination of roughness elements. Effects include local pressure gradients related to form drag on the roughness obstacles and wake production. In the upper part, at some height, $z^*$, the horizontal variability disappears (in the time average) and a ‘constant’ flux layer (surface layer) is present. It follows that, although turbulent processes above height $z^*$ may be treated as horizontally uniform in space and time, the turbulent field below $z^*$ must be considered as three-dimensional.

Consequently sensors for the measurement of turbulence parameters must be located above $z^*$ to make use of standard micrometeorological practice and for the observations to be considered fully spatially representative of the underlying surface. The vertical extent of the ‘constant’ flux layer, on the other hand, is determined by the development of an internal boundary layer which responds to mesoscale land-use changes in the upwind region. Considering that a fully adjusted boundary layer develops only slowly in the direction of the mean wind and, given the large roughness elements and surface patches typical of an urban area, it is possible that the roughness sublayer exceeds the vertical range of the ‘constant’ flux layer (typically 10% of the planetary boundary-layer depth), so that one-dimensional surface layer scaling becomes inappropriate (Oke et al. 1989).

The depth of the roughness sublayer is the subject of debate. On the basis of observations over forests and of measurements taken in wind-tunnels it would appear that the length scales relevant to the height $z^*$ are the horizontal spacing of the dominant roughness elements, their height and the roughness length (Mulhearn and Finnigan 1978; Raupach et al. 1980; Garratt 1980). If we apply the dimensions from the present study site (see below) to the corresponding range of empirical relations suggested we obtain values for $z^*$ between 20 and 100 m. This range encompasses the height of most meteorological towers in urbanized areas including that used in the present study (23 m). Therefore it cannot be supposed a priori that a traditional micrometeorological approach can be used at urban locations. For instance, MOS does not include a length scale, such as the roughness length, $z_o$, which is characteristic of the surface texture or roughness.

Very few studies deal with the observed turbulent structure over built-up surfaces, and previous results indicate that MOS might apply for some variables and statistics. Whereas the shape of the spectra/cospectra is preserved, minor deviations compared to rural reference observations occur in the location of the peaks and in the amount of low-frequency energy (e.g. Coppin 1979, height above ground $z = 34$ m, average building height 10 m; Högström et al. 1982, 6 m above the roof of a 21 m-high building; Clarke et al. 1982, $z = 30$ m surrounded by mainly 2-storey buildings). In the immediate vicinity of the roughness elements (5–10 m above an 18 m-high urban canopy) measurements by Rotach (1991) demonstrate a relatively well-preserved shape of the spectra/cospectra; however, the peaks are shifted towards higher frequencies, probably reflecting the generation of small-scale eddies in the wake of the buildings. This finding is supported by Jackson 1978 (at various heights close to tall downtown buildings) for the alongwind component. Within the urban canyon, Rotach (1991) showed that surface layer scaling is inappropriate. Here the spectra/cospectra become flat with only a small roll-off at the high-frequency end, which is possibly related to intermittent transfer processes coupling the air within the canyon with the flow above the canyon.
The few observations reported in the literature are not sufficient to warrant drawing firm conclusions. The presentation of the results often lacks the proper scaling within the surface-layer similarity framework. It is also unfortunate that very few observations of the turbulent fluxes of momentum and heat, and virtually none of moisture, exist despite their importance in driving the urban energy balance. In addition, there is no information concerning coherence and phase-angle spectra or spectral correlation coefficients, which are important descriptors of the turbulent transfer.

The present study seeks to fill some of these voids. All standard variables (wind components, temperature, humidity and the fluxes of momentum, sensible heat and moisture) are measured at two heights over a suburban surface. The spectral/cospectral characteristics are presented normalized by their respective variances/covariances and within the MOS framework (Monin and Obukhov 1954). Although the present measurements are probably not from a homogeneous surface layer (see below), MOS is used because:

(1) There is no other suitable conceptual framework available for the presentation and description of turbulence data from the roughness sublayer;
(2) It seems natural therefore to use, as an initial attempt, MOS which has provided the most successful framework for describing micrometeorological data in the homogeneous surface layer;
(3) MOS is of considerable interest and merit for the evaluation of the applicability of similarity laws over this type of surface.

2. EXPERIMENTAL ARRANGEMENT

(a) Measurement site and instrumentation

The observational program for the present study was conducted in a suburban area of Vancouver, British Columbia, Canada. The same site has been used for a number of urban climate and micrometeorology projects before (e.g. Cleugh and Oke 1986; Roth et al. 1989; Schmid and Oke 1990; Schmid et al. 1991; Grimmond et al. 1991). The suburban study site (called Sunset) is characterized by a homogeneous distribution of primarily residential houses with a mean height of 8.5 m (Fig. 1). Within a circle of 2 km radius centred on the site about 43% of the active area (approximately 1.5 times the map area) is greenspace, 13% is roof, 11% is paved and 33% is walls (or canyon). According to Steyn (1980) the area surrounding the site has a mean aerodynamic roughness length of 0.52 m and a zero-plane displacement height, \( d \), of approximately 3.5 m in the south-western sector (the sector used for the measurements).

Turbulence sensors were mounted at two levels on a triangular-section, steel lattice, free-standing tower. After allowing for \( d \) the effective measurement heights were \( z' = 11 \) and 19 m respectively \( (z'/z_0 = 21 \) and 37). The sensors at both levels were mounted on a rotatable boom attached to a 2 m-long support arm connected to the tower (Fig. 1). The position of the booms was adjusted between turbulence runs (see below) to ensure that the sensors always faced into the approaching mean wind.

A summary of the instruments used to measure instantaneous horizontal, \( u' \), \( v' \), and vertical, \( w' \), wind speed as well as the temperature, \( T' \), and humidity, \( q' \), fluctuations is given in Table 1. To measure the \( w'q' \) covariance two different instruments were combined (see Table 1). The horizontal distance between the two sensors was about 0.11 m and the humidity sensor was set back by an additional 20 mm to be able to measure
Figure 1. Photographic view of the KD (on the right) and the SAT(1127)/KH(1016) combination (on the left). A SAT(1130)/Lyman-α hygrometer combination was mounted on the same boom a similar distance (about 1 m) to the right of the KD. An identical SAT/KH combination was used at the lower level (see Table 1). The houses seen at the top (looking towards south-west) are typical for the area surrounding the site. (For sensor denotation see Table 1).

<table>
<thead>
<tr>
<th>Height z'</th>
<th>Variable</th>
<th>Instrument</th>
</tr>
</thead>
<tbody>
<tr>
<td>19 m</td>
<td>$w^', u^', w^', T^'$</td>
<td>Three-dimensional Kaijo-Denki [KD].</td>
</tr>
<tr>
<td></td>
<td>$w^', T'/q'$</td>
<td>One-dimensional sonic anemometer and fine wire thermocouple system [SAT(1130)]/Lyman-alpha hygrometer [Ly-α].</td>
</tr>
<tr>
<td>11 m</td>
<td>$w^', T'/q'$</td>
<td>One-dimensional sonic anemometer and fine wire thermocouple system [SAT(1127)]/Krypton hygrometer [KH(1016)].</td>
</tr>
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</table>

Numbers in parenthesis are instrument serial numbers. KD and SAT transducer spacing at 0.2 and 0.1 m, respectively. Hygrometer sensor and detector separation approximately 8 mm. For relative position of the various sensor combinations, see Fig. 1.

Apart from the fast-response sensors an array of slow-response instruments was used to measure mean wind speed, wind direction, temperature and humidity at the two levels, and net radiation near the top. Unfortunately the temperature and humidity
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sensors did not function properly and the respective gradients had to be rejected. Since some of the wind-speed observations at the lower level also proved to be unreliable, a local mean wind speed was derived by finding the average percentage reduction in wind speed between the two levels from some good data, and subtracting this difference from the output from the dependable sonic anemometer at the top level. In general the wind speed at the lower level was found to be 8% lower than at the upper level. This reduction between levels also compares favourably with previous measurements from the present site.

(b) Data processing and observations

Because, inevitably, the bulk of the Kaijo Denki sensor array caused distortion of the velocity field being measured, the instantaneous velocity components were therefore subsequently corrected using the procedures given by Rotach (1991); corrections were made for misalignment of the transducers from a supposedly orthogonal array, for the transducer shadow effect, using the parametrization by Wyngaard and Zhang (1985), and for flow distortion, after Wyngaard (1981).

For the present study two fast-response humidity sensors were employed (see Table 1). Both have their own benefits and drawbacks. The main advantage of the krypton hygrometer is the stability of the water-vapour absorption coefficient, its strong linear relationship with water-vapour density and the fact that the same coefficients apply over a large range of absolute humidities. After independent checks the manufacturer’s calibration was accepted for the krypton hygrometer. Its main disadvantage is the high sensitivity of the krypton line to oxygen (Campbell and Tanner 1985), so, where applicable, the measurements have been corrected for additional oxygen absorption using a procedure proposed by Tanner and Green (1989). Because of the nonlinear relationship of the Lyman-alpha hygrometer between the instrument’s response and the water-vapour density, calibrations were carried out before and after the observations. In practice there was little change in the slopes of the calibration curves between the two calibrations.

Observations were conducted mainly during afternoon and evening hours on eight days in the period 5 to 15 July 1989. Characteristic values for temperature, humidity and wind speed were 18–27°C, 8–12 g m\(^{-3}\) and 1.5–8 m s\(^{-1}\), respectively. The observations represent unstable conditions in the range \(-0.05 > z'/L_o > -1.80\) (with a mean of \(-0.62\)), where \(L_o\) is the Obukhov length,

\[
L_o = -\frac{u^3_*}{g k w^2 T_v}
\]

with \(u_* = (u'w'^2 + v'w'^2)^{1/4}\) (taking into account the possibility that the stress tensor may not be aligned with the mean wind); \(T_v\) is the mean absolute virtual temperature; \(g\) the acceleration of gravity; and \(k\) the von Kármán constant (here taken to be 0.4). A total of 40 (only 33 and 24 for humidity and momentum respectively) 60-min research-grade turbulence runs were obtained.

Data recording and processing was carried out on a PC-XT based data acquisition system. All 15 input channels were low-pass filtered with a cut-off frequency of 10 Hz using 6-order Butterworth active filters (one per channel), amplified where necessary and fed into the PC using a 12-bit successive approximation analogue-to-digital converter. The signals were sampled at a rate of 25 Hz with an associated record length of 60 min. Data processing included linear detrending, computation of spectral/cospectral statistics and averaging of the spectral/cospectral results for presentation displaying equally spaced data points in the log–frequency domain with about 10 points per decade (after Kaimal and Gaynor 1983). To obtain better representation of the low-frequency end the first
few points were not averaged (or only over a few data points), with the result that the
low-frequency side remains statistically unreliable.

3. THEORETICAL BACKGROUND

The presentation of spectra is based on the Kolmogorov–Obukhov hypothesis (e.g. Cham-
gagne et al. 1977). According to the second Kolmogorov hypothesis for the inertial
 subrange (between the production and dissipation scales), and using Taylor’s frozen
turbulence hypothesis to convert wavenumbers to frequencies (i.e. \( k_1 = 2\pi n/U \), where
\( k_1 \), \( n \) and \( U \) are wavenumber, natural frequency and mean wind speed, respectively), the
spectral density, \( S \), of velocity components is given by the expression

\[
nS_{u,v,w}(n) = A_{1,2} \varepsilon^{2/3}(2\pi n/U)^{-2/3}
\]

where \( \varepsilon \) is the dissipation rate of turbulent kinetic energy and \( A_{1,2} \) is a universal constant
for the inertial subrange (Kolmogorov constant). Since the inertial subrange levels for \( v \)
and \( w \) are higher than those for \( u \) by a factor of 4/3, predicted by isotropy, the corresponding
Kolmogorov constants differ in the same ratio. In the present study \( A_1 \)
(for \( v \) and \( w \)) and \( A_2 \) (for \( u \)) were set at 0.68 and 0.5, respectively (Panofsky and Dutton
1984).

Similarly, for temperature and humidity fluctuations, the one-dimensional spectra
can be represented in the forms

\[
nS_T(n) = B_T \varepsilon^{-1/3} N^*(2\pi n/U)^{-2/3}
\]

and

\[
nS_q(n) = B_q \varepsilon^{-1/3} \gamma^*(2\pi n/U)^{-2/3}
\]

where \( N^* \) and \( \gamma^* \) are the rates of destruction by molecular conductivity of \( T'^2/2 \) and
\( q'^2/2 \) respectively. \( B_T \) and \( B_q \) are constants analogous to \( A_1 \) and \( A_2 \) in (1), but these
values are not as well established and in many studies they are chosen as constant and
close to 0.8 for both temperature and humidity (e.g. Paquin and Pond 1971; Dyer and
Hicks 1982). In this study we use the value recommended by Panofsky and Dutton (1984)
for both \( B_T \) and \( B_q \) of 0.78.

Equations (1) to (3) can be solved for the dissipation rates \( \varepsilon, N^* \), and \( \gamma^* \) which are
used in the computation of the respective non-dimensional dissipation rates defined as

\[
\phi_e = \varepsilon k^z/u^3
\]

\[
\phi_N = N^* k^z/u^* T^2
\]

\[
\phi_q = \gamma^* k^z/u^* q^2
\]

where \( T^* = -\bar{w}'T'/u^* \) and \( q^* = -\bar{w}'q'/u^* \), the scaled surface-layer temperature and
humidity, respectively.

Replacing the natural frequency \( n \) with a non-dimensional frequency \( f = nz'/U \) and
substituting Eqs. (4) to (6) into Eqs. (1) to (3) leads to the equations

\[
nS_{u,v,w}(n)/u^2 = A_{1,2}(2\pi n k)^{-2/3} \phi_e^{2/3} f^{-2/3}
\]

\[
nS_T(n)/T^2 = B_T(2\pi n k)^{-2/3} \phi_e^{-1/3} \phi_N f^{-2/3}
\]

\[
nS_q(n)/q^2 = B_q(2\pi n k)^{-2/3} \phi_e^{-1/3} \phi_q f^{-2/3}
\]
The Eqs. (7) to (9) are consistent with Monin-Obukhov scaling, which implies that spectral characteristics are a function only of $f$ and $z'/L_v$ (note that the non-dimensional dissipation rates depend on stability). If we now include $\phi_v, \phi_h$ and $\phi_q$ in the normalization of the spectra (i.e. divide the left-hand sides of Eqs. (7) to (9) by these quantities) we remove the $z'/L_v$ dependence in their equations. This brings all spectra into coincidence in the inertial subrange, whereas the low-frequency parts spread out as a function of stability. This form of normalization has been applied to the spectra presented in section 7.

The representation of the cospectra follows the practice of Wyngaard and Côté (1972). They proposed a model which supposes that the cospectrum between the vertical velocity and another variable at high frequencies is proportional to the vertical gradient of that variable. The cospectra also depend only on $\varepsilon$ and $k_1$, with constants that are stability dependent. Using dimensional analysis, invoking Taylor’s frozen turbulence hypothesis and replacing wavenumber with non-dimensional frequency, we obtain for the cospectra of momentum, heat and water vapour the expressions

$$\begin{align*}
n Co_{uw}(n) &= -\xi_1 \frac{\partial U}{\partial z} \varepsilon^{1/3}(z'/2\pi)^{4/3}f^{-4/3} \\
n Co_{wT}(n) &= -\mu_1 \frac{\partial T}{\partial z} \varepsilon^{1/3}(z'/2\pi)^{4/3}f^{-4/3} \\
n Co_{wq}(n) &= -\delta_1 \frac{\partial q}{\partial z} \varepsilon^{1/3}(z'/2\pi)^{4/3}f^{-4/3}
\end{align*}$$

where $\xi_1, \mu_1$ and $\delta_1$ are the non-dimensional cospectral constants and the overbars denote time average. (Note that the original paper by Wyngaard and Côté does not include the humidity cospectrum. Here it is supposed that humidity behaves in a manner similar to that of temperature, which should be the case in the inertial subrange). Unlike the spectra the cospectra fall off more rapidly with a $-4/3$ slope (compared to $-2/3$).

Using Eq. (4) and the equations for the non-dimensional wind, temperature and humidity gradients defined as

$$\begin{align*}
\phi_m &= (kz'/u_*) (\partial U/\partial z) \\
\phi_h &= (kz'/T_*) (\partial T/\partial z) \\
\phi_q &= (kz'/q_*) (\partial q/\partial z).
\end{align*}$$

Eqs. (10) to (12) can be put into the usual Monin–Obukhov form, viz.

$$\begin{align*}
n Co_{uw}(n)/u_*^2 &= -(2\pi k)^{-4/3} \xi_1 \phi_m \psi_1^{1/3}f^{-4/3} \\
n Co_{wT}(n)/u_* T_* &= -(2\pi k)^{-4/3} \mu_1 \phi_h \psi_1^{1/3}f^{-4/3} \\
n Co_{wq}(n)/u_* q_* &= -(2\pi k)^{-4/3} \delta_1 \phi_q \psi_1^{1/3}f^{-4/3}.
\end{align*}$$

The combinations $\xi_1 \phi_m \psi_1^{1/3}, \mu_1 \phi_h \psi_1^{1/3}$ and $\delta_1 \phi_q \psi_1^{1/3}$ depend on $z'/L_v$ only and can be determined directly from measurements of cospectra in the inertial subrange and $z'/L_v$ without going through the separate functions. Using empirical functions for the individual terms in these products (Wyngaard and Côté 1972; Kaimal et al. 1972) we obtain the following normalization functions for momentum, heat and water vapour:

$$\begin{align*}
G(z'/L_v) &= \xi_1 \phi_m \psi_1^{1/3}/(0.56k^{4/3}) \\
&= -n Co_{uw}(n)(2\pi)^{4/3}f^{4/3}/(0.56u_*^2)
\end{align*}$$
The empirical constants used in the evaluation of these combinations are based on measurements in a homogeneous surface layer. Corresponding values are not available for the urban environment but differences are likely. In addition Kaimal et al. (1972) use \( k = 0.35 \) which affects the size of the denominator on the right-hand side in Eqs. (19) to (21). To bring the cospectra into coincidence in the \(-4/3\) region we divide the left-hand sides of Eqs. (16) to (18) by \( G(z'/L_o), H(z'/L_o) \) and \( Q(z'/L_o) \), respectively, which again, as for the spectra, separates the low-frequency end, according to stability. This is the normalization applied to the cospectra in section 7.

When examining the transfer mechanism the spectral correlation coefficients

\[
R_{ij}(f) = \frac{C_{ij}(f)}{\{S_i(f)S_j(f)\}^{1/2}}
\]

are of special importance, where \( ij \) is any of \( uw, wT, wq \) or \( Tq \).

The coherence spectra are defined as follows:

\[
Coh_{ij}(f) = \frac{\{C_{ij}(f) + Q_{ij}(f)\}^{1/2}}{\{S_i(f)S_j(f)\}^{1/2}}
\]

where \( Q_{ij} \) are the quadrature spectra. The phase-angle spectra are given by

\[
Ph_{ij}(f) = \tan^{-1} \frac{Q_{ij}(f)}{C_{ij}(f)}.
\]

4. Evaluation and comparison of fast-response sensors

Multiple sensors were installed at the upper level (Table 1) for the following purposes:

1) Back-up in the case of malfunction.
2) Intercomparison of different sensors (e.g. KD vs. SAT and Lyman-\( \alpha \) vs. KH).
3) Examination of reliability and repeatability of the turbulence measurements.

The performance of these instruments will be investigated using the spectral information from all the sensors at the upper level. The observations plotted in Fig. 2 are averages taken over all unstable runs (see below).

The \( w \)-signals from the three different sensors agree with each other over a large frequency range (Fig. 2(a)). Differences occur only at the high-frequency end and reflect the limitations set by the sensors’ response times. The onset of the inertial subrange (\(-2/3\) slope) is observed for values of \( f \) of about 2, however, at non-dimensional frequencies greater than about 9 for SAT and 4 for KD sensors the spectra start to drop off faster. Because of line averaging over the sensor path the instruments are not able to resolve the smallest eddy sizes. The present observations agree roughly with theoretical predictions if use is made of the transfer function associated with spatial averaging of turbulent velocity fluctuations. For example according to Moore (1986) spectral distortion is expected to appear for values of \( f \) about 12 for SAT and 6 for KD sensors with a spacing of 0.1 m and 0.2 m, respectively. At the high-frequency end the KD sensor
performs better for the horizontal than for the vertical wind components (Fig. 2(a)) and only experiences appreciable high-frequency degradation for values of $f$ greater than 8.

The SAT temperature sensors agree with each other over the entire available frequency range (Fig. 2(b)), but not too well with the KD temperature sensor which reads higher than the SAT sensor at $f > 0.9$ and which does not follow the $-2/3$ slope. The KD measurements are based on the temperature dependence of the velocity of sound from the same ultrasonic pulse propagation times as are used in the calculation of the vertical velocity, for which, however, agreement with SAT is good (Fig. 2(a)). The KD temperature fluctuations were not corrected for the well-known humidity and horizontal wind contaminations (cf. Schotanus et al. 1983) which probably also affect the spectral shape, but since it seems unlikely that the application of these corrections could eliminate the observed differences the KD temperature observations were subsequently
rejected. On the other hand the high-frequency roll-off measured by the SAT sensors is slightly faster than the inertial subrange prediction. This is unexpected considering that the frequency response is reported to be about 30 Hz (Biltoft and Gaynor 1987).

The current experiments also provide a first opportunity to compare the spectra obtained from different types of fast-response hygrometers. The results from the Lyman-α and KH sensors agree fairly well over most of the frequency range; only at the highest frequencies do they deviate from each other (Fig. 2(b)). According to Massman et al. (1990) three-dimensional averaging over the volume between the source and the detector tubes should result in lower spectral densities (faster roll-off compared to the inertial subrange prediction) at non-dimensional frequencies greater than about 35 (using the dimensions of the present sensors). However, the data in Fig. 2(b) show the opposite effect, with the spectral densities from the KH sensor increasing at \( f > 4 \), and at \( f > 20 \) for the Lyman-α sensor. A common source of high-frequency noise in spectra is aliasing of potential energy contained above the Nyquist frequency, but this ought not to be a problem here because of low-pass filtering. It is interesting to note that both Krypton hygrometers behave in exactly the same way at the high-frequency end (the KH sensors' data from the lower level are not shown here), yet there are no electronic hygrometer problems known to the manufacturers which could cause this kind of high-frequency contamination; it has still not been possible to identify the source of this noise.

Based on the results plotted in Fig. 2 all spectra and corresponding cospectra presented herein are truncated at \( f = 7 \), which approximately corresponds to the non-dimensional frequency at which the errors described above start to make their appearance. In the case of the KH sensors, truncation was made at \( f \approx 4.5 \). Figure 2 also shows that there is excellent agreement between observations from different sensors measuring the same variable; which is an important result that demonstrates the good performance and reliability of the instrumentation used. In the following, the \( w, T, q, wT \) and \( wq \) results from the upper level \( (z' = 19 m) \) are presented as averages from the two SAT hygrometer combinations.

5. SPECTRA/COSPECTRA NORMALIZED WITH VARIANCE/COVARIANCE

In this section the composite spectra/cospectra are plotted against the non-dimensional frequency \( f \) on log scales. Composite denotes the average over all individual spectra/cospectra of a particular variable (or combination of two variables). Following Large (1979) we have estimated the standard deviations indicated on the plots from \( \sigma_n = \sigma / (M^{1/2}) \), where \( M \) is the number of runs, which is a better indicator of the variability of the mean than the common standard deviation, \( \sigma \), (calculated from all data-points within a frequency band). In the following figures, in the interests of readability, \( \sigma_n \) is plotted for only one of the sensors.

Where possible the results from the present study are compared with observations taken at the same site in 1986 under unstable conditions (mean \( z'/L = -0.75 \)) (Roth et al. 1989). The rural reference results in this section are those of Anderson and Verma (1985). Although they are not from 'ideal' surface conditions (soya bean crop), and represent different stabilities, these data were chosen as a reference because this is one of the few studies which includes the complete set of variables normalized by the respective variance/covariances. Compared to the present observations (from unstable conditions) the near neutral results from Anderson and Verma (1985) are, generally, shifted towards higher frequencies. This is in agreement with the stability dependence usually observed in the homogeneous surface layer (cf. Kaimal et al. 1972). In the
following figures the solid lines of $-2/3$ and $-4/3$ slope indicate the proportionality of the spectral and cospectral densities, respectively, to the inertial subrange laws.

(a) Velocity spectra

In Fig. 3(a) the composite $w$-spectra are presented and compared against those from other studies. Agreement with the results from the 1986 study is very good. Compared to the reference, a spectral model developed by Højstrup (1981) based on the Minnesota results (Kaimal 1978), the present data have slightly more low-frequency energy with an associated shift of the peak towards larger scales (also observed by Steyn 1982 at the same site as is in use here). By virtue of Taylor’s hypothesis, $f_m = z'/l_m$, where $l_m$ is the peak wavelength; with $f_m = 0.15$, the dominant eddy scale is computed to be $l_m \approx 6.7z' \approx 127\text{ m}$. The $-2/3$ slope is followed in the region of the inertial subrange.

The $w$-results from the lower level correspond well with those from the top. The lower spectral densities observed at high frequencies are probably the result of a reduction in the rate of increase of wind speed with height observed at this site. If we assume as a working hypothesis that the wind profile is logarithmic and if we make use of the $U$ and $u^*$ data presented in Part II (Fig. 1), then the wind speed reduction between the two levels should be about 20% (compared to the 8% observed). If this value were used (appropriate for the surface layer) the corresponding non-dimensional frequency $f$ would be larger and would result in a shift of the spectrum towards higher frequencies, which in fact is the case for all spectra/cospectra from the lower sensors, as will be seen later.

There is fair agreement in the energy containing range between the composite $u$-spectrum and the 1986 observations (from a limited data-set using a modified Gill twin propeller-vane anemometer) (Fig. 3(b)). The faster roll-off at the low-frequency end compared to the reference (from near neutral conditions) is consistent with the concept that with energy-producing eddies there is a shift towards smaller scales in the wake of buildings upstream. A similar result was obtained by Jackson (1978) in another urban study. The $w$-component is not affected in the same way because it has most of its energy at relatively high frequencies which adjust more rapidly to changes in roughness features.

(b) Scalar spectra

Compared to the data from the 1986 study and the reference, the temperature spectrum measured at the upper level is slightly shifted towards higher frequencies (Fig. 4(a)). It has been suggested that both the $u$ and $w$ components contribute to the fluctuations in $T$ (Lumley and Panofsky 1964). The shift of the $T$-spectral peak could therefore be caused by the shift observed in the $u$-spectrum towards higher frequencies. The observations at the lower level are again in good agreement with those from the top.

The humidity composite spectra (Fig. 4(b)) are the first of their kind for an urban environment. The $-2/3$ slope at the upper level is not reached until values of $f$ of about 2, unlike the rural studies in which the onset is at $f \approx 0.3$ (Elagina 1969; Smedman-Högström 1973; Ohtaki 1985) or at $f = 1.3$ (Anderson and Verma 1985, Fig. 4(b)). The behaviour at the low-frequency end of the humidity spectrum is not well covered in the
Figure 3. Normalized composite spectra of (a) $w$, (b) $u$, and (c) $v$ vs. $f$ on a log-log plot. This study (symbols) compared with a model developed by Højsstrup 1981 (solid line) is shown in (a). The 1986 results (dashed line) are shown in (a) and (b) and the rural reference from Anderson and Verma 1985 (dotted line) in (b) and (c). The vertical lines are $\pm 1\sigma_v$. 
literature and, obviously, the results from the present study deviate from the reference (representing near neutral conditions). Whereas Ohtaki (1985) measured a peak at $f = 0.08$ the observations by Högström and Smedman-Högström (1974) show a point of inflexion at $f \approx 0.1$ and a broad peak at around $f \approx 0.02$ associated with a considerable amount of scatter at the low-frequency end (a feature which is evident in most rural observations). The results from the present study mostly resemble the humidity measurements by Phelps and Pond (1971) taken over the sea, and the results from unstable conditions in Minnesota analysed by Schmitt et al. (1979), which also do not show a low-frequency roll-off. There is again good agreement between the data from both levels.

The observations in Fig. 4(b) suggest that the humidity transfer is affected by large-scale structures. This can be expected over the rough urban surface which is likely to produce well-developed interaction between the surface and the upper portion of the urban boundary layer. In about 50% of the measurement runs (usually from cloudy days with weak convective activity) there were signs that the humidity transfer was influenced
by structures which extended well above the surface layer. The corresponding time traces were dominated by a few large negative deviations (up to 1.7 g m\(^{-3}\)) which lasted for 15 to 60 s and were associated with negative \(w'\)-values (Roth 1991). In these cases evaporation was driven more by the downdraughts of drier air from above.

(c) Cospectra

Considering the extent of the roughness of the urban surface, which should exert a strong influence on the dynamics of the airflow, the momentum-flux results correspond surprisingly well with measurements from surfaces with a low roughness and greater homogeneity (Fig. 5(a)). The overall shape is preserved but it is not as smooth as is usually observed. The position of the peak frequency compares well with other suburban observations by Coppin (1979) but is slightly lower than is generally reported from rural studies. The frequency range \(0.1 < f < 3\) is marked by relatively low energy densities which gives the cospectrum a ‘flat’ appearance. Another ‘dip’ is observed at \(f = 0.05\). Because of its ragged appearance it is impossible to confirm the \(-4/3\) slope at the high-frequency end. The 1986 results (dashed line) are from a very limited data-set using a

![Figure 5. Normalized cospectra of (a) \(u_w\), and (b) \(u_T\); the results from other studies are as indicated.](image)
modified Gill twin propeller-vane anemometer. The shape of the $uT$ composite cospectrum is similar to that of $uw$ (Fig. 5(b)). Wyngaard and Coté's (1972) theoretical prediction of $-2$ for the high-frequency end cannot be confirmed nor is the $-3/2$ slope followed that was observed by Kaimal et al. (1972).

The composite sensible heat-flux cospectra (Fig. 6(a)) agree very well with the 1986 results and the suburban observations of Coppin (1979). The $-4/3$ slope is followed for $f > 3$. Comparison with the reference is fair; however, the location of the peak is still within the corresponding unstable region indicated by Kaimal et al. (1972) (see also Fig. 10(b)). Differences between the two levels are again small.

The first moisture-flux cospectrum measured in a suburban atmosphere is shown plotted in Fig. 6(b). It closely resembles that for heat flux but is marked by more variability at low frequencies. A $-4/3$ slope appears to be followed in the inertial subrange ($f > 2$), yet the cospectral estimates show some variability (more so at the lower level). The relatively organized low-frequency roll-off is surprising considering the shape of the corresponding $q$-spectrum (Fig. 4(b)). This can in part be attributed to the steep low-frequency roll-off found in the $w$-spectrum. Furthermore, the $wq$ spectral correlation coefficients are relatively low at large scales (Fig. 7), which indicates that there is little net humidity transfer. Agreement with the near-neutral reference is not good but is better with the unstable rural results from McBean and Miyake (1972) and Schmitt et al. (1979). The first point at the low-frequency end is negative (but plotted as a very small positive value) and reflects the high occurrence of negative cospectral estimates at low frequencies in the individual moisture-flux cospectra.

The overall shape of the composite $Tq$ cospectrum (Fig. 6(c)) resembles rural observations (e.g. McBean and Elliot 1981). Wyngaard et al. (1978) predict a $-2/3$ slope for locally isotropic turbulence and suppose that there is a constant spectral correlation coefficient in the locally isotropic region. However, as shown in Fig. 7, this requirement is not satisfied, and so this leads to the progressively steeper high-frequency roll-off observed in the present data. The $Tq$ values at the low-frequency end are occasionally negative but are plotted as small positive numbers.

6. SPECTRAL CORRELATION COEFFICIENTS, COHERENCE AND PHASE-ANGLE SPECTRA

The normalized cospectra indicate which scales are important in contributing to the respective transfers but, for investigating how these transfers take place, the spectral correlation coefficients are more useful. With little net transfer the quantity $R_{ij}(f)$ will be small but will reach unity for optimally efficient turbulent transfer. For these reasons the term 'transfer efficiency' has been associated with the spectral correlation coefficient (McBean 1973). These important descriptors of the turbulent transfer are presented in Fig. 7 together with the corresponding coherence (indicating the similarity of the structure of two time-series at a certain frequency, without regard to the presence of a phase shift) and phase-angle spectra (indicating the phase shift between two variables at each frequency).

The spectral correlation coefficients for $uw$ (note that it is $-R_{uw}(f)$ that is plotted) are high throughout the low-frequency end. However, despite these large magnitudes, large scales do not necessarily contribute much to the total transport, as is shown in the small contributions to the total covariance at the low-frequency end of the corresponding cospectrum (Fig. 5(a)). The maximum $-R_{uw}(f)$ of about 0.6 observed in the present study is higher than was found over rural surfaces (e.g. McBean and Miyake (1972) measure a peak value of about 0.35 at $f \approx 0.06$). This very efficient momentum transfer at large scales in the urban environment may result from organized flow structures (e.g.
Figure 6. Same as for Fig. 5 but for (a) $wT$, (b) $wq$, and (c) $Tq$; the results from other studies are as indicated.
Figure 7. Composite spectral correlation coefficients, $R_{uv}$ (top row), coherence spectra, $Coh_{uv}$ (middle row) and phase angle spectra, $Ph_{uv}$ (bottom row) vs. log $f$ for $uw$, $uT$, $wT$, $wq$ and $Tq$ (from left to right). Negative values are plotted for $R_{uw}$ and $R_{uT}$. Diamonds and stars are from $z' = 19$ and 11 m, respectively.
urban heat-island circulation; horizontal or vertical advection of air). At $f = 0.05$ the transfer efficiency is slightly reduced. The small values observed at the highest frequencies are common to all transfers and are a result of instrumental effects and randomization due to the turbulent energy cascade (Phelps and Pond 1971). The $uw$ coherence follows the behaviour of the spectral correlation coefficients. The phase angle is out of phase by 180 degrees for the entire frequency range up to $f = 2$ which is to be expected since a negative $w$ is associated with a positive $u$ or vice versa; at the high-frequency end the phase angle becomes meaningless because the coherence is very small. $R_{wT}(f)$ closely tracks $R_{uw}(f)$; e.g. a pronounced dip at $f = 0.05$ is observed in both.

The $R_{wT}(f)$ observations (Fig. 7) compare well with the rural results of McBean and Miyake (1972) as regards the magnitude and location of the peak. As was the case with the momentum transfer, the latter is at a slightly smaller frequency compared to the corresponding cospectra (Fig. 6(a)). Unlike $uw$, the $wT$ correlation drops off towards the low-frequency end. The coherence values from the present study are higher than those by Ohtaki (1985) who measured a broad peak of about 0.6 in the frequency range $0.01 < f < 0.1$. The phase angles are close to zero, which is to be expected if the surface is a source of sensible heat (positive $w$ associated with positive $T$).

The correlation and coherence results for $wq$ (Fig. 7) indicate a less efficient transfer at all scales compared with sensible heat. McBean and Miyake (1972) observed a peak in $R_{wq}(f)$ of almost 0.6 at a slightly higher frequency compared to the suburban results. The coherence values from the present study are also less than were measured by Ohtaki (1985) whose $wq$ results are similar to his $wT$ observations (see previous paragraph). The overall shape of the $Tq$ spectral correlation coefficients (Fig. 7) is similar to rural measurements done by Phelps and Pond (1971). At the lowest frequencies the $Tq$ correlation tends to reverse sign. A positive correlation can be expected if, for instance, the results for warm and wet conditions are correlated (convection from the surface) whereas a warm (cold) and dry (wet) combination would result in negative correlations. The latter possibility can occur if for example horizontally advected, relatively humid air is imported from the mixed layer into the surface layer.

As noted in the previous section agreement between the data from both levels is very good. The slight shift of the observations at the lower height towards lower frequencies has already been noted and explained before.

7. Normalized within the Monin–Obukhov similarity framework

With appropriate normalization the spectra/cospectra are reduced to a family of curves which spread out according to $z'/L_o$ at low frequencies but converge to a single universal curve in the inertial subrange. Empirical evidence of this kind has been obtained by Kaimal et al. (1972) based on the 1968 Kansas AFCRL measurement program, and will serve as the ‘ideal’ (low roughness, homogeneous fetch) reference in this section. In order to analyse the data from the present study systematically, the turbulence runs were divided into four stability groups on a subjective basis. These groups with their stability limits and the number of runs in each group (in parentheses) are as follows:

(A) $-0.01 > z'/L_o \geq -0.10$ (3)
(B) $-0.10 > z'/L_o \geq -0.40$ (15)
(C) $-0.40 > z'/L_o \geq -1.10$ (18)
(D) $-1.10 > z'/L_o \geq -1.80$ (4)
For humidity and moisture flux the number of runs in groups (B) and (C) are 9 and 17, respectively and, for momentum, the corresponding numbers for (A), (B) and (C) are 2, 14 and 8, respectively. This reduction was necessary because some of the time-series were found to be non-stationary.

A composite spectrum/cospectrum for each stability class was computed in a manner similar to that of section 5. Because the frictional velocity was not measured at \( z' = 11 \) m, the local scaling variables \( u_*, T_* \) and \( q_* \) and also \( z'/L_* \), needed for the presentation within the MOS framework, are not known. Extrapolation from the upper-level data would be a dubious procedure, since these variables show strong height dependence within the roughness sublayer (Rotach 1991 for \( u_* \) and also probably vary as a function of the structure and roughness of the underlying surface. Consequently no results from the lower level are shown in this and the following section.

In the following the non-dimensional dissipation and normalization functions which are used in the normalization of the spectra/cospectra are presented first.

\( (a) \) Dissipation and normalization rates

In practice the dissipation rates \( \varepsilon, N^* \) and \( \gamma^* \) were computed using Eqs. (1) to (3) with the spectral densities determined as the average over a small number of points at frequencies within the \(-2/3\) region of the corresponding spectra (determined subjectively; usually for \( 2.5 < f < 5 \)). The non-dimensional dissipation rates in Fig. 8 were then calculated using Eqs. (4) to (6). Similarly the normalization functions for momentum, heat and moisture flux (Eqs. (19) to (21)) were estimated from the cospectral densities within the \(-4/3\) region (usually for \( 4 < f < 7 \)).

A plot of \( \phi_e \) (derived from \( w \)-data) versus \( z'/L_\nu \) is shown in Fig. 8(a). The solid line represents the Kansas data from Wyngaard and Coté (1971) expressed in the form:

\[
\phi_e = \{1 + 0.5(-z'/L_\nu)^{2/3}\}^{3/2}.
\]

The dashed curve is obtained from the St. Louis study by Clarke et al. (1982) for their suburban site. The present results differ markedly from the Kansas data but agree well with the St. Louis observations for small negative values of \( z'/L_\nu \). As instability increases, a trend similar to the Wyngaard curve with values between the Kansas and St. Louis data is followed. Values of \( \phi_e \) obtained from the horizontal wind components (not shown) were generally slightly higher and were associated with more scatter. As a consequence the ratio of vertical to horizontal velocity components in the inertial subrange of \( 4:3 \), required from isotropy, was not obtained in the present study. The values reached were generally smaller and of the order of 1.1 to 1.2. The present results also agree well with those from the 1986 study (Roth 1990).

There are possibly two different mechanisms that could produce the differences observed in Fig. 8(a). Clarke et al. (1982) conclude that under near-neutral conditions, when buoyancy production can be neglected, lower dissipation values are associated with lower values of \( \phi_m \). Because of the absence of reliable wind-profile data, \( \phi_m \) could not be directly evaluated; however, the observations that are available indicate a reduced increase of wind speed with height compared to the logarithmic prediction (see subsection 5(a)). The same effect has been suggested as being responsible for lower values of \( \phi_m \) found in previous studies over rough surfaces (e.g. Garratt 1978a, b). Furthermore, an analysis of integral statistic results in Part II provides indirect support. For instance, the transfer efficiency of momentum is unusually high (Part II, Fig. 9(a)) and the normalized vertical-velocity standard deviations are lower than in the reference data (Part II, Fig. 4(a)).
Figure 8. Non-dimensional dissipation rates for (a) turbulent kinetic energy, (b) heat and (c) moisture vs. \( z'/L_v \). The solid line is the rural reference. The dashed line in (a) is from the St. Louis study for the suburban site (Clarke et al. 1982).
With unstable conditions Clarke et al. (1987) show that at their suburban site more energy was produced locally than was dissipated, and that $\phi_e$ was lower at the suburban site than at their rural reference site. This produced a large residual component of turbulent kinetic energy (TKE) at their suburban site, which, they concluded, was due to vertical transport, flux divergence, pressure transport and, possibly, horizontal advection. It seems reasonable to suppose that large and organized horizontal and vertical structures in the urban environment as well as transfer of energy through pressure-velocity interactions will result in increased transport of locally produced TKE away from the surface, which would lead to local production exceeding the local dissipation. The relatively efficient momentum transfer observed in the present study (Fig. 7; Part II, Fig. 9(a)) indirectly supports this hypothesis.

The non-dimensional dissipation rates for temperature and humidity are compared with the Kansas temperature data (solid line) expressed in Fig. 8(b) and (c) as $\Phi_N = 0.74 (1 - 9z'/L_o)^{-1/2}$ (no such reference data are available for humidity but Ohtaki 1985 shows that $\Phi_N$ is similar to $\Phi_N$ over wheat fields). The trend of the Kaimal data is generally followed; however, the values are usually below that of the reference. It is possible that processes similar to those responsible for the decreased values of $\Phi_e$ also result in lower dissipations for temperature and humidity.

The cospectral normalization functions are plotted in Fig. 9(a–c) and compared against the Kaimal prediction of unity (solid line). The considerable scatter in $G(z'/L_o)$ reflects the ragged appearance of the individual cospectra in the inertial subrange (Fig. 5(a)) and is a feature also observed by Wyngaard and Coté (1972). The variability is decreased for $H(z'/L_o)$ and $Q(z'/L_o)$; however, the values for these quantities are generally lower than for the reference, especially for moisture.

(b) Velocity spectra

The composite vertical-velocity spectra for the four stability groups are plotted in Fig. 10(a). The solid line represents a composite Kansas spectrum for neutral stratification (Kaimal et al. 1972) and the dashed line above is the Kaimal spectra for $z/L = -2$ (note that Kaimal et al. use $z/L$ not $z'/L_o$). All spectra coincide in the inertial subrange because of the normalization procedure. The behaviour at the high-frequency end of all variables has been discussed in reference to the spectra/cospectra normalized by their corresponding variances/covariances and will not be repeated here. Similar to the style observed by Kaimal et al. (1972) (their Fig. 3) an orderly progression of both the spectral peak and the spectral densities is observed in the direction of increasingly smaller $f$ as $-z'/L_o$ increases. This behaviour is most pronounced for values of $f$ between 0.03 and 0.3. The least unstable class is distinctly different from other curves. The peak frequencies in the present data are shifted slightly towards lower values compared to the reference. The correspondence with observations from both Högström et al. (1982), at their central city site (dotted line in Fig. 10(a); representing near-neutral conditions), and the suburban study by Clarke et al. (1982) (not shown) is good.

The horizontal-velocity spectra are presented in Fig. 10(b, c). The area between the solid and lower dashed lines is what Kaimal et al. (1972) describe as an excluded zone which separates the stable from the unstable regime (caused by a sudden shift in the dominant scales of motion as $-z'/L_o$ changes sign). In a similar way as in the urban study by Clarke et al. (1982) the present $u$-component composite spectra cluster around the lower end of the Kaimal unstable region, and are within this supposedly excluded zone. The present observations show fairly rapid roll-off at the low-frequency end, in agreement with Högström et al. (1982) (dotted line in Fig. 10(b)). The $v$-component spectra (Fig. 10(c)) correspond well with the Kaimal reference and with those of Högström.
Figure 9. Normalization functions for the fluxes of (a) momentum, (b) heat, and (c) moisture vs. $z'/L_v$. The solid line at unity is the rural reference (Kaimal et al. 1972).
Figure 10. Composite spectra for four stability classes of (a) \( w \), (b) \( u \), and (c) \( v \) vs. \( f \) on a log–log plot. The solid and upper dashed lines are the Kaimal et al. (1972) limits for neutral and \( z/L = -2 \), respectively. The lower dashed lines in (b) and (c) mark an excluded zone (see text). The dotted line represents near neutral data from the urban site in Högström et al. (1982).
et al. (1982), apart from the least unstable class, which again is found in the excluded zone. The low frequencies of the horizontal wind spectra are governed by mesoscale processes and depend on the height of the planetary boundary layer, $z_i$, rather than $z$ (Kaimal 1978). No values of $z_i$ were available; however, the present $u$ data cluster randomly in a small area at the low-frequency end whereas the $v$-component spectral densities show separation according to stability, in particular for the transition of the least unstable class to the next higher one.

(c) Scalar spectra

The present temperature observations crowd into a relatively narrow frequency band at the low-frequency end (Fig. 11(a)) unlike those of Kaimal et al. (1972), who observed a systematic shift of their spectra with increasing instability from the $z/L = -2$ curve (dashed line) towards the neutral curve (solid line). The results from the present

![Figure 11](image-url)  

Figure 11. Same as for Fig. 10 but for (a) $T$ and (b) $q$, and without the lower dashed and dotted lines.
study show higher energy content in the mid-frequency range with an associated shift of the peak frequencies towards higher values (as already observed in Fig. 4(a)). Apart from this mid-frequency range, agreement with other urban observations by Clarke et al. (1982) is good.

The humidity data (Fig. 11(b)) are compared to the Kaimal temperature curves (no such reference data exist for humidity). Similar to the present temperature observations the humidity spectra show no tendency towards separation according to the values of $z'/L_u$. Schmitt et al. (1979) conducted an analysis of the rural Minnesota data and observed that their data was within the unstable zone of the Kaimal temperature spectra. The only difference that they mention is a departure for $0.1 < f < 1.0$ marked by slightly larger spectral density values than the temperature reference. Whether humidity should be a function of $z'/L_u$ is questionable. McBean (1971) concludes that the humidity fluctuations are not governed by $z'/L_u$, but instead by $z'/L_q$ where $L_q$ is a stability parameter defined in terms of the moisture flux. McBean further points out that the similarity variables are based on local values; however, any observation is affected by large-scale space and time inhomogeneities of the atmospheric motions which, in particular, affect the low-frequency part of the spectra (see also discussion above in respect to the humidity spectra in Fig. 4(b)).

(d) Cospectra

The unstable $uw$ cospectra crowd into a narrow band at the low-frequency end (Fig. 12(a)). The same is observed in the reference (indicated in Fig. 12 as the area between the two dashed lines) but the energy densities are lower in the present study although the peak locations are at similar frequencies. The former might be a feature introduced by the relatively large normalization rates (Fig. 9(a)). The previously noted ‘dip’ at $f \approx 0.05$ is observed in the cospectra of all stability categories. Because of the large scatter and occasional negative values at the low-frequency and high-frequency ends, the results from the most unstable group have been omitted. In addition the $uw$ correlations for these cases were very low, which prevents a clear interpretation.

The $wT$ results are in good agreement with the Kaimal data (Fig. 12(b)). The least unstable group has the lowest energy densities and highest peak frequency, whereas the more unstable data crowd into a relatively narrow band. The same is observed for the moisture flux (Fig. 12(c)). Here the Kansas heat-flux data is used as a reference. The humidity data from the Minnesota experiment by Schmitt et al. (1979) presented within MOS (but using $H(z'/L_u)$ instead of $Q(z'/L_u)$ in the normalization) show good agreement with that reference.

8. Stability Dependence of Spectral Correlation Coefficients

The composite values of the $uw$ spectral correlation coefficients increase with decreasing instability at almost all frequencies, indicating that momentum transfer is most efficient for near-neutral conditions (Fig. 13(a)). The largest magnitudes are observed at frequencies slightly smaller than the peak frequencies in the corresponding $uw$ cospectra (Fig. 12(a)). All composite spectra correlation coefficients show a ‘dip’ for values of $f$ between about 0.05 and 0.06 which also corresponds to a region with slightly lower energy densities in the corresponding composite cospectra. The observed stability dependence is in agreement with measurements by McBean and Miyake (1972) from a flat rural site; however, they noted a slight roll-off below $f \approx 0.01$ and peak magnitudes which were slightly lower than those measured in the suburban data.
Figure 12. Same as for Fig. 10 but for cospectra of (a) $uw$, (b) $wT$, and (c) $wq$; the lower dashed lines approximate the lower unstable limit from Kaimal et al. (1972).
Figure 13. Composite spectral correlation coefficients for four stability classes of (a) $uw$, (b) $wT$, (c) $wq$, and (d) $Tq$ vs. log $f$. 
The heat-flux data show a tendency towards an increase in correlation coefficients with an increase in instability for $f$ greater than 0.02 (Fig. 13(b)). This effect is most pronounced for the transition from the stability class closest to neutral to the next unstable class. The low-frequency end is marked by small values and much scatter, which is similar to rural observations by McBean and Miyake (1972), who also observed the largest values to be associated with the most unstable group and measured a peak at $f = 0.01$ with slightly lower values of about 0.6 (compared to 0.75 in the present study) for unstable conditions.

The $wq$ observations are marked by considerable scatter at low frequencies but, in general, the least unstable group has the lowest magnitudes over the entire frequency range (Fig. 13(c)). For values of $f$ greater than 0.04 the correlation coefficients increase with increasing instability, a feature which was also observed by McBean and Miyake who reported measuring peak magnitudes which were slightly higher than those measured in the present study. The $Tq$ correlations exhibit a strong relation with stability. They are marked by a sharp roll-off at the low-frequency side but stay relatively high throughout most of the remaining frequency range before decreasing again at the highest frequencies (Fig. 13(d)).

It is generally supposed that the transfers of heat and moisture in the surface layer are similar, i.e. the correlation coefficients should equal each other. Specifically the transfer of these fluxes can only be regarded as similar if the spectral correlation coefficients also are similar over the entire frequency range measured. Comparing Fig. 13(b) and (c) shows that this is not the case over a suburban surface with $R_{wT}(f)$ being systematically larger than $R_{wq}(f)$ at all frequencies and for all stability categories.

In general no significant differences could be observed between the observations from the two levels (lower level not shown) in terms of shape, peak locations and magnitudes, the exception being the magnitudes of the least unstable group at the lower level which were slightly, but systematically, lower compared to the upper level.
9. DISCUSSION AND CONCLUSIONS

The data presented in this paper are the first detailed results of spectral/cospectral characteristics of all important atmospheric variables from the urban roughness sublayer. The spectra/cospectra show only minor differences of energy distribution in respect of frequency and the location of the peaks in comparison with data from the homogeneous surface layer. This is true for the spectra/cospectra normalized with the respective variances/covariances as well as being analysed within the MOS framework. The present observations are also in good agreement with those of the few available from other urban sites. Minor deviations, in comparison with reference data, which may be peculiar to the present site (and possibly to urban sites in general) include the following.

(1) The peak of the $w$-spectrum is slightly shifted towards lower frequencies.

(2) The peak of the $u$-spectrum is slightly shifted towards higher frequencies associated with a relatively fast roll-off at the low-frequency end, with spectral densities falling into Kaimal's supposedly excluded region.

(3) The $T$-spectrum is slightly shifted towards higher frequencies and the stability dependence of the low-frequency end is weak.

(4) The $q$-spectrum is characterized by a low-frequency end which does not show a roll-off for near neutral, to moderately unstable, conditions.

(5) The $uw$ cospectrum has two minor 'dips' at $f=0.05$ and $f=0.4$, and the $-4/3$ slope, which is only approached slowly at high frequencies.

Although the non-dimensional dissipation and normalization functions, in particular with increasing instability, follow the trend predicted by similarity theory (i.e. they display the same power-law dependence) the following differences, compared to the reference data, were observed.

(6) $\phi_w$ is systematically smaller, more so under slightly unstable conditions than with large instabilities.

(7) $\phi_u$ and $\phi_q$ are generally smaller.

(8) $H(z'/L_v)$ and $Q(z'/L_v)$ are generally smaller.

(9) $G(z'/L_v)$ is generally larger and marked by much scatter.

The similarity of the transfer mechanism for a particular $z'/L_v$ is given only if the spectral correlation coefficients agree scale by scale for all scales involved in the transfer. The present results show that

(10) under near-neutral conditions the momentum transfer is more efficient at low frequencies than the transfers of heat and moisture;

(11) under unstable conditions the heat transfer is most efficient at all scales, followed by water vapour and momentum;

(12) compared to the reference data $R_{uu}(f)$ is higher (especially at low frequencies) and $R_{wq}(f)$ smaller at most frequencies.

The spectra/cospectra normalized with the variances/covariances measured at the lower level agree very well with those at the upper level. This indicates that the turbulence structure is similar over the height interval used. The slight shift towards lower frequencies observed at the lower level may be due to the reduced increase in wind speed with height observed in the urban roughness sublayer when compared to the logarithmic prediction.

The differences from the reference data listed above suggest that $z'$ is less than $z^*$, i.e. the observations are from within the roughness sublayer. This possibility was also explored by Schmid (1988). He estimated the range of conditions when measurements
at $z' = 19\text{ m}$ at the Sunset site are expected to be fully representative and therefore when a homogeneous surface layer is present. His source-area calculations in relation to the standard deviations of stability and normalized lateral velocity define a criterion for the representativeness of a measurement which can be applied to the present observations. The current data applied to Schmid's (1988) Fig. 11.1 demonstrate that for most of the time the turbulence sensors were located below $z^*$. It should be noted, however, that the height of the roughness sublayer is not necessarily the same for momentum, heat and water vapour. A roughness sublayer (although this term strictly refers only to the spatial variability introduced by the bluff bodies) develops in relation to the distribution of the momentum sinks and heat and moisture sources, but these do not necessarily coincide.

In view of the inhomogeneity of the boundary conditions, the deviations observed in the normalized spectra/cospectra are of moderate magnitude. Close to the roughness elements we would expect that the characteristic length of the turbulent eddies and the turbulent statistics were influenced by the surface structure. Observational support for roughness effects on spectra/cospectra are sparse, the exception being velocity measurements within crop and forest canopies (e.g. Baldocchi and Meyers 1988). The spectral/cospectral wind and temperature results from within and just above an urban canyon from Rotach (1991) also show some anomalies, which, however, do not seem to correlate with surface roughness length scales.

The spatial scales which would most likely influence the turbulence measurements at the present site are, for wind and momentum, the spacing of the house rows ($50-70\text{ m}$, depending on wind direction); and, for temperature, the horizontal spacing between house roofs and street surfaces ($25\text{ m}$), both of which have relatively warm temperatures during daytime, as well as the house-row and street/alley spacing ($50\text{ m}$). These scales were identified by Schmid and Oke (1992) (using a roughness inventory and remotely-sensed surface temperatures of the general Sunset area) as the dominant contributions to a variance analysis of roughness and temperature features. Supposing that the individual wakes and plumes leaving the surface are related to the above scales (and also considering the height of the houses to be important) one would expect that this should result in subsidiary spectral features in the following cases, using $z' = 19\text{ m}$:

\[
0.3 < f < 0.4 \quad (50 < \lambda < 70) \\
\lambda \approx 0.8 \quad (\lambda \approx 25) \\
\lambda = 2.2 \quad (\lambda \approx 8.5)
\]

The suburban results, however, do not exhibit any such irregularities (not even at the lower level) with the possible exception of the lower energy densities observed in the $uw$ cospectrum for $0.3 < f < 0.6$ (Fig. 12(a)). It appears that surface features only produce clearly recognizable anomalies in the spectral/cospectral shapes if the sensors are located in the immediate vicinity of the roughness elements.

Three possibilities are suggested to explain some of the observed differences from reference data, and which may be used to understand the relative success of MOS applied to the present measurements:

1) Wakes which are shed from buildings produce turbulent energy at a smaller scale than that produced by the mean wind shear. The energy-containing eddies produced by wake and shear will result in an energy input at slightly higher frequencies compared to the undisturbed flow. This might explain the relatively fast roll-off and the shift in the peak frequency observed in the $u$-spectrum in the present study and also in those of
Jackson (1978), Clarke et al. (1982), Roth (1990) and Rotach (1991). The turbulent wakes created by the bluff bodies are also a zone of small eddies which develop in response to vortices shed from the edges of buildings. These eddies are efficient in transporting momentum across the mean stream lines, which could explain the relatively high correlation coefficients for momentum observed between medium and high frequencies. At the low-frequency end, organized local-scale or urban-scale horizontal or vertical flows could be responsible for the increased efficiency of momentum transfer.

(2) It should be recognized that the time averages in use are linked to a spatial average as the result of a time-varying source area which includes a variety of combinations of surface cover (Schmid et al. 1991). This would tend to blur any easily recognizable features in the results due to the inhomogeneous surface.

(3) In MOS the scaling parameters are defined as their surface values which, however, are likely to vary with height in the roughness sublayer. It is possible, however, that the turbulent statistics are in equilibrium with the local fluxes as they are in the homogeneous surface layer. Experimental support for the success of local scaling is given for near-neutral urban data by Högström et al. (1982).

It was mentioned in section 1 that the structure of turbulence in the roughness sublayer must be considered to be three-dimensional. It is therefore appropriate to discuss to what extent the present results are affected by this consideration. The present observations could be used to study some aspects of the vertical variability, for example it was shown that the vertical variability was small for normalized spectra/cospectra. No spatial information was obtained because the measurements refer to one point only. Very few observations are available to shed light on horizontal inhomogeneity close to a rough surface. For instance, considerable spatial inhomogeneity for momentum flux profiles (up to 20–30%) has been shown to exist by Mulhearn and Finnigan (1978) who made their measurements over an array of stones in a wind tunnel. Furthermore, although not directly related to the characteristics of turbulence, it is interesting to note that the sensible heat flux at the present site shows variations of up to 25–40% within horizontal scales of $10^2$–$10^3$ m (Schmid et al. 1991). These variations, however, reflect the preferential influences of the immediate surroundings and mean that the measurements may not be representative of larger-scale surface conditions. In the case of appropriately normalized turbulence parameters (using local scaling) the horizontal variability is expected to be smaller. Again it should be pointed out that the time averages used contain a spatial averaging component (see above).

Quantitative estimates of spatial inhomogeneity are not possible but, based on the small vertical variability measured and the arguments presented above, it is thought to be relatively small at the upper level. Consequently it is suggested that the present results are applicable to other urban areas of similar morphological structure. However, there is certainly a need for more research in this area and future measurement programs of similar nature should be designed to be able to deal with the subject of horizontal variability.

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