Surface influence upon vertical profiles in the atmospheric near-surface layer

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SUMMARY

Observations from two towers situated in flat, tree-covered terrain \( z_0 \) lying between 0·4 and 0·9 m have been used to investigate the flux-profile relations in the height range \( z/z_0 \) from 5 to 85, where \( z \) is the height above the zero-plane displacement. The analysis confirms a lower height limit (at \( z = z_a \)) to the validity of the Monin and Obukhov functions \( \Phi_{M, N}(z/L) \) in unstable conditions and, by implication, of the logarithmic wind law in neutral conditions. We find \( z_a/z_0 \approx 35 \) and 150 for wind at the denser and less dense \( (z_0) \) surfaces, whilst for temperature \( z_a/z_0 \approx 100 \).

The level \( z_a \) corresponds with the top of the transition layer, within which it is assumed the profiles depend additionally upon a length scale \( z_s \), related to surface wake generation. On the assumption that \( z_s \ll z_a \), modification of the profiles in the transition layer is then described through a function \( \delta(z/z_*) \) whose explicit form is derived from length-scale considerations in a region of wake-shear interaction. The observed non-dimensional profiles \( \Phi^* \) are well represented by

\[
\Phi^* \approx \Phi(z/L) \exp(0-7z/z_*)
\]

both for wind and temperature. For wind at both surfaces, the depth \( z_* \) is approximately constant in unstable conditions and equal to 30, \( \delta \) being the tree spacing. We tentatively conclude that \( \delta \) is the relevant surface length scale \( z_* \) characterizing the wake field and depth of penetration \( z_* \).

1. INTRODUCTION

According to boundary-layer theory, a well-developed boundary layer (of depth \( D \)) supports an ‘inner layer’ of depth approximately 0·1 \( D \) (e.g. Tennekes and Lumley 1972; p. 162) within which the turbulent fluxes decrease gradually with height. Further, if \( D/z_0 \) is sufficiently great \( (z_0 \) is the aerodynamic roughness length of the surface) an ‘inertial sub-layer’ exists within the ‘inner layer’ and here, taking the neutral case for example, a logarithmic wind law is predicted both from dimensional analysis (e.g. Sutton 1953; pp. 78–82) and more detailed boundary-layer theory (e.g. Sheppard 1969; Tennekes and Lumley 1972; pp. 151–156). Observations well above both smooth and rough surfaces (generally for \( z_0 > 100 \)) in the wind tunnel (e.g. Hinze 1959; pp. 465–487) and in the atmosphere (e.g. Sutton 1953, chapter 7) confirm the predicted ‘inertial sub-layer’ law.

At lower heights \( (z_0 < 100) \) within which the influence of the surface should be important, deviations from the log law have been observed in the wind tunnel (e.g. Raupach et al. 1979) and claimed by Thom et al. (1975) and Garratt (1978a) for atmospheric observations extrapolated from unstable conditions. The above considerations suggest defining a level \( z_* \) as the interface between the ‘inertial sub-layer’ above and a ‘transition layer’ below. The latter thus has a height range

\[
z_0 < z < z_*
\]

and together with the ‘inertial sub-layer’ constitutes the ‘inner layer’. Both Tennekes (1973) and Townsend (1976; pp. 139–143) imply \( z_*/z_0 \approx 50–100 \).

\[\dagger\] The basis of this is in classical fluid dynamics describing the transition region between the viscous sub-layer and the fully turbulent layer in flow over a smooth flat plate. Hinze (1959; p. 473) adopts the same terminology for flow over a rough wall for which Rotta attempted to take into account the effect of the roughness upon the velocity profile close to the surface. Raupach et al. (1979) refer to this transition layer as the roughness sub-layer.

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Garratt (1978a) based his analysis upon the observed non-dimensional vertical gradients of wind ($u$) and potential temperature ($\theta$)

$$
\Phi^0_M = (k_u z/\theta_*) \partial u/\partial z; \quad \Phi^0_H = -(k_\theta z/\theta_*) \partial \theta/\partial z
$$

(1)

where $z = Z - d$, $Z$ being the height above the ground and $d$ the zero plane displacement; $u_*, \theta_*$ are the turbulent scales for velocity and temperature respectively, defined uniquely in terms of the surface turbulent fluxes of momentum and sensible heat. Conditions were non-neutral, so that comparison was made with the universal formulation of the profiles predicted by Monin and Obukhov (1954) for non-neutral conditions in the 'inertial sub-layer'; viz. for $z > z_*$,

$$
\Phi^0_{M,H} = \Phi_{M,H}(z/L)
$$

(2)

where $L$ is the buoyancy length scale. Equation (2), containing two height/length scales, reduces to the gradient form of the logarithmic law in neutral conditions containing the height scale $z$ only ($L = \pm \infty$), where $\Phi(0) = 1$ and $k_u$ and $k_\theta$ are not necessarily equal. Both wind tunnel (e.g. Clauser, 1956; Schlichting 1968; Townsend 1976) and atmospheric (e.g. Businger et al., 1971; reviews by Dyer 1974 and Yaglom 1974) observations support values of $k_u$ and $k_\theta$ in the ranges 0.35 to 0.43 and 0.40 to 0.47 respectively. Monin and Yaglom (1971) have given a comprehensive review of 'interpolation' and semi-empirical treatments which yield explicit forms of $\Phi(z/L)$. In the present study we adopt those suggested by Dyer (1974), mainly based on observations over low-$z_0$ surfaces and applicable to the 'inertial sub-layer',

$$
\Phi_H(z/L) = \Phi^2_M(z/L) = (1 - 16 z/L)^{-\frac{1}{4}} \quad \text{for } z/L < 0,
$$

and

$$
\Phi_H(z/L) = \Phi_M(z/L) = (1 + 5 z/L) \quad \text{for } z/L > 0
$$

with $k_u = k_\theta = 0.41$. At present the analogous flux-profile relations within the 'transition layer' are unknown.

The observations considered in the present paper are from two towers situated in flat, aerodynamically very rough terrain and cover the height range $z/z_0$ of 5 to 85, the lowest heights being of the same order as the spacing ($\delta$) and height ($h$) of the roughness elements (trees).

2. Sites, experimental details and data selection

Measurements of the turbulent fluxes of momentum and sensible heat and related vertical profiles were made at two sites (designated M1 and M2), roughly 5 km apart, in flat, very rough, tree-covered terrain near Daly Waters, Northern Australia (16°16’S, 133°28’E) during July–August 1974. Results of a flux-profile analysis of M1 observations have been presented by Garratt (1978a,b), where details of instrumentation and techniques can be found (see also Clarke and Brook 1979). A summary of surface and observation details is shown in Table 1.

The specific density ($\lambda$) of trees, where $\lambda$ is the estimated average frontal area of tree divided by the average horizontal area occupied by each tree, has been determined from surface-based and aerial photographs (see Garratt 1979a); estimates have a probable uncertainty of ±50%.

The aerodynamic roughness length ($z_0$) has been derived from wind profile analysis as described in Clarke and Brook (1979) – see Appendix 1. Determination of the zero plane displacement ($d$) is described in section 3.
TABLE 1. Details of surface dimensions at M1 and M2 and observation levels on the two masts. Here, $\delta$ is the estimated average horizontal spacing between trees; $l_h$ is an estimated horizontal dimension (width) of the trees and $\lambda$ is the estimated specific density of trees.

<table>
<thead>
<tr>
<th>Site</th>
<th>Mean tree height, $h$ (m)</th>
<th>Zero plane $d$ (m)</th>
<th>Spacing $\delta$ (m)</th>
<th>Width $l_h$ (m)</th>
<th>Density $\lambda$</th>
<th>$z_0$ (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>M1</td>
<td>8.0</td>
<td>4.8</td>
<td>20</td>
<td>1.5</td>
<td>0.03</td>
<td>0.4</td>
</tr>
<tr>
<td>M2</td>
<td>9.5</td>
<td>7.1</td>
<td>10</td>
<td>2.0</td>
<td>0.2</td>
<td>0.9</td>
</tr>
</tbody>
</table>

Measurement heights, $Z$ (m)

$u$, $\theta$ Profiles

<table>
<thead>
<tr>
<th>Site</th>
<th>11.2</th>
<th>15.3</th>
<th>21.8</th>
<th>31.8</th>
<th>48.3</th>
<th>17.5</th>
</tr>
</thead>
<tbody>
<tr>
<td>M1</td>
<td>10.0</td>
<td>12.3</td>
<td>15.8</td>
<td>20.8</td>
<td>28.3</td>
<td>17.5</td>
</tr>
<tr>
<td>M2</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Layer geometric mean heights, $z = Z - d$ (m)

Layer

<table>
<thead>
<tr>
<th></th>
<th>1-2</th>
<th>2-3</th>
<th>3-4</th>
<th>4-5</th>
</tr>
</thead>
<tbody>
<tr>
<td>M1</td>
<td>8.2</td>
<td>13.4</td>
<td>21.5</td>
<td>34.3</td>
</tr>
<tr>
<td>M2</td>
<td>3.9</td>
<td>6.7</td>
<td>10.9</td>
<td>17.0</td>
</tr>
</tbody>
</table>

Temperature profiles were not measured at site M2.

$\Phi^0_M$ and $\Phi^0_H$ have been determined from wind ($\Delta u$) and temperature ($\Delta \theta$) differences between adjacent levels, and are appropriate to effective levels ($\bar{z}$) given by the geometric mean heights of the air layers 1–2, 2–3, 3–4 and 4–5 as shown in Table 1. The eddy fluxes described in Clarke and Brook (1979) were used to deduce the parameters $u_*$ and $\theta_*$; the constants $k_u$, $k_\theta$ were taken as 0.41.

Data were selected according to criteria based on known or assumed accuracy of mean $u$ and $\theta$ observations and the eddy fluxes; thus data were excluded when,

(i) $\Delta u < 0.1 \text{ m s}^{-1}$; $u_* < 0.2 \text{ m s}^{-1}$
(ii) $|\Delta \theta| < 0.05 \text{ degC}$; $|\theta_*| < 0.05 \text{ degC}.$

In addition data were excluded for the transition periods 0700–0900 inclusive and 1700–1800 inclusive, where generally fluxes were small, changed sign and $\theta$ exhibited a non-monotonic variation with height. Neutral and near-neutral conditions were absent under predominantly clear skies, and during the rapid transition at sunrise and sunset. At night a stably stratified boundary layer was rarely observed, turbulence generally decaying rapidly after sunset in the presence of strong surface cooling.

For the subsequent analysis, confined to unstable conditions, 157 and 142 one-hour runs were available at M1 for $\Phi^0_M$ and $\Phi^0_H$ and 111 runs at M2 for $\Phi^0_M$ only.

3. Zero plane displacements ($d$) and aerodynamic roughness lengths ($z_0$) at sites M1 and M2

The flux-gradient analysis requires determination of the quantity $d$, whilst $z_0$ is required for aerodynamic and scaling considerations. Use of inertial sub-layer relations (Eq. (2) in integrated form) allows both of these to be evaluated from the observations. However, previous experience (wind tunnel observations) and boundary-layer theory suggest the possibility of a deep transition layer above our surfaces (see Introduction). One can show that for observations within the transition layer, values of $d$ will be seriously underestimated whilst values of $z_0$ are little affected. The latter are dealt with in Appendix 1.

For $d$, Garratt (1979a) applied the inertial sub-layer relations to the wind and tempera-
ture observations and found

at M1, \( d/h = 0.06 \), \( d = 0.5 \text{ m} \)

at M2, \( d/h = 0.37 \), \( d = 3.5 \text{ m} \)

For inertial sub-layer observations these values should correspond to values determined by other means, independent of the atmospheric data.

One such method identifies \( d \) with the mean level of momentum absorption within the canopy (Thorn 1971), requiring in-canopy wind profiles not available in the present experiment. Another approach is to consider the relation between \( d/h \) and element density \( \lambda \), although the precise functional dependence has been difficult to determine even for well-specified rough surfaces and flow characterization (e.g. Marshall 1971; Wooding et al. 1973; Seginer 1974). Fortunately Kutschbach (1961) derived, from wind profiles measured above bushel basket arrays, the variation of \( d/h \) with \( \lambda \) (see his Fig. 12), and Seginer (1974), using a canopy mixing-length model and taking the canopy element drag coefficient as an additional variable, showed \( d/h \) as a function of \( \lambda \) (see his Fig. 5). Applying their results to our two surfaces, with \( \lambda(M1) = 0.03 \) and \( \lambda(M2) = 0.2 \), gives lower and upper limits to \( d/h \),

at M1, \( d/h = 0.45 - 0.6 \)
at M2, \( d/h = 0.65 - 0.75 \)

The former compares with a value of 0.64 used by Garratt (1978a) in an analysis of M1 observations. These values are significantly greater than those determined from the observations (see additional comments in Garratt 1979a) and are taken, together with indications from boundary-layer theory, as evidence for a deep transition layer – hence the requirement to treat the observations appropriately.

Rather than take the mean values determined by each of the above ranges of \( d/h \), we have used the upper values for our main analysis, giving

at M1, \( d = 4.8 \text{ m} \)
at M2, \( d = 7.1 \text{ m} \).

This then allowed the subsequent results to be substantiated by repeating the analysis with the lower values of \( d/h \) in each case.

4. Data analysis

The variations of observed gradients \( \Phi_0^g \) and \( \Phi_0^h \) for the four layers at site M1 have been described by Garratt (1978a) using \( d_{u,0} = 5.1 \text{ m} \). These data have been recalculated with the revised value of 4.8 m and, together with the M2 \( \Phi_0^g \) data using \( d_u = 7.1 \text{ m} \), have been grouped in narrow ranges of \( L \) to give mean values and standard deviations presented in Appendix 2. The ratio \( \Phi_0^g/\Phi(z/L) \) is shown in Figs. 1 to 3 as a function of \( z/L \), and in most cases is significantly less than unity. Strong variations of this ratio with \( z/L \) are not apparent in most cases. The small values could result from overestimates in the fluxes or to inadequate fetch requirements both of which seem unlikely (e.g. Garratt 1978a); overall, they could be increased by the choice of much smaller (but physically unrealistic) zero plane displacements (e.g. Garratt 1979a).

Observations of \( \Phi_0 < \Phi(z/L) \) are apparently related to levels within the transition layer where enhanced vertical mixing effectively reduces the vertical gradients below those given by Eq. (2). The latter are appropriate to the height range

\[ z_a < z \ll D \]

in which the fluxes are essentially constant with height and the relevant height and length
Figure 1. Variation of $\Phi_M/\Phi_M(\xi/L)$ with $\xi/L$ at M1 for four adjacent air layers – see data averages for each $L$ range in the Appendix. Mean data represented as follows: layer 1-2 ($\Delta$), 2-3 (○), 3-4 (●) and 4-5 (X). The pecked curves represent the locus of the standard deviations of the individual data about the mean for each narrow $L$ range.

Figure 2. As in Fig. 1, for M2.

Figure 3. Variation of $\Phi_M/\Phi_M(\xi/L)$ with $\xi/L$ at M2 for four adjacent air layers; other information as in Fig. 1.
scales upon which the vertical gradients depend are $z$ and $L$.

For heights $z < z_*$ we assume the vertical gradients depend upon an additional length scale $z_*$. This length scale is imposed upon the flow through the action of turbulent wake generated by flow around the roughness elements. Wind tunnel studies of turbulent wake structure above, (i) a single bluff body (e.g. Counihan et al. 1974) and (ii) an array of bluff bodies (e.g. Perry and Joubert 1963; Perry et al. 1969) on a rough surface show that the wake field penetrates to heights several times that of the roughness elements. Within this wake field vertical flux divergence and horizontal variations in mean velocity and friction velocity $u_*$ are likely to exist, but are probably significant only near the tops of the roughness elements (see wind tunnel observations of Mulhern and Finigan 1978; Raupach et al. 1979). Through most of the transition layer, we therefore assume the profile modification can be described in terms of $z_*$.

Now it seems reasonable to assume that the depth $z_*$, at least in neutral conditions and for $z_* < |L|$, will be uniquely determined by $z_*$, since the latter effectively determines the dominant length scale of the wake field comprising, in part, the transition layer. Hence we take $z_* \propto z_*$ where the constant of proportionality is of order unity. For a neutral layer, we write for velocity

$$\frac{\partial u}{\partial z} = \left(\frac{u_*}{kz}\right) \phi(z/z_*)$$

so that the observed wind gradient $\Phi^0_M$ should satisfy,

$$\Phi^0_M = \phi_M(z/z_*)$$  \hspace{1cm} (3)$$

In the non-neutral case we assume the observed wind and temperature gradients within the transition layer should satisfy a modified version of Eq. (2), viz.,

$$\Phi^0_{M,H} = \Phi^*_{M,H}(z/L, z/z_*)$$

the superstar indicating a general functional form for $z < z_*$. We write this as

$$\Phi^0_{M,H} = \Phi_{M,H}(z/L) \phi_{M,H}(z/z_*, z_*/L)$$  \hspace{1cm} (4)$$

in analogy with Eq. (3) which itself may be rewritten as

$$\Phi^0_M = \Phi_M(0) \phi_M(z/z_*)$$

Eq. (4) suggests that the observations presented in Figs. 1 to 3 can be represented in the transition layer by $\phi(z/z_*, z_*/L)$

5. THE GRADIENT FUNCTION $\phi$

For practical reasons we consider a possible functional form of $\phi(z/z_*)$ i.e. as applicable to the neutral case. Other forms based on the turbulent energy equation and a wake mixing length model have been described by Garratt (1979b). The latter for instance yielded $\phi^{-1}_M = 1 + b(z/z_*)^{-1}$, with $b$ a constant. Unfortunately none of the above is entirely satisfactory, being based on dubious physical assumptions. In the absence of a convincing physical theory of wake-shear interaction, we shall attempt therefore to describe the observations in Figs. 1 to 3 as a transition layer phenomenon on the basis of dimensional arguments. Now above $z = z_*, \phi_M$ is constant and equal to unity, so that the layer $z < z_*$ is one in which significant variations in $\phi_M$ occur. Dimensionally the depth $z_*$ is of order the length scale of changes in $\phi_M$, viz.,

$$z_* \sim \phi_M/(\partial \phi_M/\partial z)$$  \hspace{1cm} (5)$$

whence it follows from integration of Eq. (5), with $\phi_M = 1$ at $z = z_*$,
SURFACE INFLUENCE IN THE NEAR-SURFACE LAYER

\[ \phi_M = \exp\{-\alpha (1 - z/z_*)\} \]
\[ \equiv \alpha \exp\{\alpha (z/z_*)\} \]

where \( \alpha \) is an unknown constant, and \( \alpha \equiv \exp(-\alpha) \). Equation (6) serves as a simple interpolation formula and we make the assumption that it may be used to describe \( \phi_H \) also, so that it has been applied to all the data shown in Figs. 1 to 3 to determine \( \alpha \) and \( z_* \) for each \( L \) range.

6. MEAN PROPERTIES OF THE TRANSITION LAYER

(a) Transition layer depth

Values of \( z_* \) for each data set are shown in Fig. 4 as a function of \( L \). Several anomalously high values (circled) are associated with low correlation coefficients of the corresponding 'least square' fits. For wind and temperature in unstable conditions, there is no evidence that \( z_* \) is affected significantly by buoyancy, at least for our observed ranges of \( z_*/L \) of greater than \(-0.7 \) for M2 and greater than \(-3.0 \) for M1. The average values of \( z_* \), viz. 60 m and 30 m approximately for M1 and M2 wind respectively and 40 m for temperature, probably apply to neutral conditions also.

Such values of \( z_* \) imply that most observations were made within the transition layer (as can be seen in Figs. 1 to 3), so that extrapolation from the observation levels according to the variation of \( \Phi^0/\Phi(z/L) \) with height given by Eq. (6) was necessary to infer \( z_* \). This situation is unfortunate, since a greater number of observations at levels above \( z_* \) would have

![Image of graphs showing inferred values of \( z_* \) as a function of \( L \): (a) for M1, temperature profile. (b) for M1, wind profile. (c) for M2, wind profile. Circled values correspond to low correlation coefficients in the least square fit for that \( L \) range.](image-url)
ensured greater confidence in the values derived here. However, in the planning of the experiment, maximum levels of surface-based observations were determined mainly by available instrument masts and fetch considerations. Nevertheless, there is no obvious reason why the inertial sub-layer structure should differ from that over low-\(z_0\) surfaces. Certainly there is similarity in the gross boundary-layer structure (e.g. Garratt and Francey 1978; also unpublished analysis of the author), suggesting in addition an inner layer of depth 150–200 m.

In terms of relative depth \(z_*/z_0\) referred to in the Introduction we find values of 150 and 35 approximately for the less dense and denser surfaces respectively; for the M1 surface \(z_*\) for wind is somewhat greater than for temperature, as noted by Garratt (1978a).

For wind the above values (call these \(z_{*M}/z_0\)) are sufficiently different for the two surfaces to suggest that \(z_0\) is not the surface length scale \(z_*\) which ‘uniquely’ determines \(z_{*M}\). Let us therefore compare \(z_{*M}\) with several possible candidates for \(z_*\) – these might include, in addition to \(z_0\), \(h\) (height of trees), \(\delta\) (spacing between trees), \(l_h\) (horizontal length scale of trees either in the longitudinal or transverse wind direction – for our two surfaces these are similar), appropriate magnitudes being given in Table 1. We find

<table>
<thead>
<tr>
<th>(z_{*M}/z_0)</th>
<th>(z_{*M}/h)</th>
<th>(z_{*M}/l_h)</th>
<th>(z_{*M}/\delta)</th>
<th>(z_{*M}) (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>M1</td>
<td>150</td>
<td>7.5</td>
<td>40</td>
<td>30</td>
</tr>
<tr>
<td>M2</td>
<td>35</td>
<td>3.2</td>
<td>15</td>
<td>3.0</td>
</tr>
</tbody>
</table>

which allows the tentative suggestion that we should look to the spacing as an important length scale characterizing the wake field and hence the depth of penetration \(z_{*M}\). However, we believe this fact, associated as it is with the increase in \(z_{*M}\) with decrease in density \(\lambda\), is valid only where the underlying surface makes a negligible contribution to the total surface stress (as with our two surfaces). We must admit, however, that these results for the atmospheric case are not in agreement with the recent wind tunnel experiments of Raupach et al. (1979) who varied \(\delta\) several-fold, but found \(z_{*M}\) relatively insensitive to this variation.

\(z_{*M}\) was related to the lateral horizontal scale \(l_h\), kept constant during their experiment. This obvious difference with our results is as yet unexplained though one must be aware of the scaling problems inherent in their wind tunnel studies. For instance, the depth of their boundary layer was considerably smaller, relative to the height of the roughness elements and hence to the inferred transition layer depth and inner layer depth (typical ratios of 20, 8 and 4 respectively) than in the atmospheric case (typical ratios 150, 35 and 10 respectively).

(b) The parameter \(\alpha\)

We find that \(z_*\) is essentially independent of stability (in unstable conditions) so that any stability dependence of \(\Phi_{M,H}^0\) not accounted for by \(\Phi_{M,H}(z/L)\) should then be seen in \(\alpha\). Matching Eqs. (4) and (6) suggests,

\[
\phi(z/z_*) = \{\phi_1(z_*/L)\} \{\phi_2(z/z_*)\}
\]

with \(\phi_2 = \exp\{a_1(z/z_*)\} \) and \(\phi_1 \equiv \alpha\).

We therefore show in Fig. 5 values of \(\alpha\) as a function of \(z_*/L\) for each data set. For wind \(\alpha\) appears to be independent of surface though with a tendency to increase slowly when \(z_*/L < -1\); in the case of temperature there is no significant variation. In all three cases \(\alpha\) has a value close to 0.5, i.e. the gradient \(\phi\) attains a minimum value of 0.5 as the surface is approached (\(z/z_* \to 0\)).

(c) Observations within the transition layer

The data presented in Figs. 1 to 3 are now more appropriately shown in Figs. 6 to 8,
with $\Phi^0/\Phi(z/L)$ as a function of $z/z_*$ for each $L$ range (represented by a symbol given in the Appendix data tabulation); the value of $z_*$ found from the fitting to Eq. (6), and shown in Fig. 4, is used. Smooth curves represent Eq. (6) with average values of $\alpha$ for each data set obtained from individual values shown in Fig. 5; for $z > z_*$, $\phi = 1$ is appropriate.

The smooth curves for $z < z_*$ represent the data with an overall r.m.s. deviation of the individual data about 0-1-0-15. This is comparable with that generally experienced at $z \gg z_*$ when $\phi = 1$ represents the mean curve (e.g. see data of Dyer 1967; Dyer and Hicks 1970; Businger et al. 1971; Hicks 1976).

Although the use of Eq. (2) for $z > z_*$ and Eq. (6) for $z < z_*$ is appropriate, the ‘kink’ at $z = z_*$ is a deficiency which requires matching of the equations in the region of $z = z_*$ to ensure continuity in $\partial \phi / \partial z$ at $z = z_*$. 

(d) Excess eddy diffusivity and mixing lengths

The eddy diffusivity, defined through the flux-gradient relation, both above and within the transition layer, can be written as follows:

\[ z > z_*: \quad K_{M,H} = k u_* z / \Phi_{M,H}(z/L) \]  

and

\[ z < z_*: \quad K_{M,H}^* = k u_* z / \Phi_{M,H}^* \]  

The excess eddy diffusivity in the transition layer is then given by,

\[ K_{M,H}^*/K_{M,H} = (\Phi_{M,H})^{-1} \exp\{-\alpha_1(z/z_*)\} \]  

In mixing length terms, Eq. (8a) implies

\[ l^*/l = \phi^{-1} \]

with $K^* = u_* l^*$, $K = u_* l$, although this is contained in Eqs. (3) and (4) implicitly. In other words, Eqs. (3) and (4), through the surface parameter $z_*$, describe the effects of the associa-
Figure 6. M1 wind profiles showing $\Omega_{M}/\Omega_{\infty}(\delta/L)$ as a function of $z/z_*$ for all $L$ ranges shown in the Appendix (with appropriate codes); $z_*$ values are from Fig. 4. The pecked curves represent the locus of the standard deviations of the individual data about the mean for each narrow $L$ range. The continuous curve below $z/z_*=1$ is the function $\phi_M \exp \{-\alpha_1 (1-z/z_*)\} \equiv \exp \{\alpha_1 (z/z_*)\}$ with $\alpha_1 = 0.62$ ($\alpha = 0.54$). Above $z/z_*=1$, $\phi_M = 1$.

Figure 7. As in Fig. 6, for M2 wind profiles (appropriate codes in the Appendix); $z_*$ values from Fig. 4. The continuous curve below $z/z_*=1$ has $\alpha_1 = 0.65$ ($\alpha = 0.52$).

Figure 8. M1 temperature profiles showing $\Omega_{M}/\Omega_{\infty}(\delta/L)$ as a function of $z/z_*$ for all $L$ ranges shown in the Appendix (with appropriate codes); $z_*$ values are from Fig. 4. Otherwise as in Fig. 6, where the continuous curve below $z/z_*=1$ has $\alpha_1 = 0.82$ ($\alpha = 0.44$).
ted wake turbulence upon the vertical mixing in terms of modification of the shear diffusivity or mixing length by a multiplicative factor $\phi^{-1}$. This is precisely how several workers (see Hinze 1959, pp. 472–473) have viewed the effect of viscosity upon the velocity profile in the transition layer above a smooth surface. Such an approach ensures continuity in the gradient function through the level $z = z_a$, though $\partial \phi/\partial z$ is discontinuous at $z = z_a$. In contrast, the few previous attempts to describe the ‘influence’ of the surface upon the velocity profile above a rough surface have been in terms of an additive wake diffusivity or mixing length $l_w$ (e.g. Rotta – see Hinze 1959, p. 473, and even Prandtl – see Hinze 1959, p. 281), using relations of the kind $l^* = l + l_w$ and $l^*2 = l^2 + l_w^2$. Their major disadvantage is to give a discontinuity in the gradient function at $z = z_a$.

7. Dependence of Parameters upon $d$

The description of the profile forms through the concept of a transition layer, characterized by the parameters $z$ and $z_a$ through Eq. (6), has been based on the upper values of $d$ for each surface determined from the $f(\lambda)$ dependence. This analysis (mainly through the preliminary paper of Garratt, 1978a) has been criticized by Hicks et al. (1979) who argued that the systematic deviations of $\Phi_{M,H}^0$ from $\Phi_{M,H}(z/L)$ could be removed, thus requiring no transition layer concept, by assuming much smaller values of $d$. The latter, however, are not acceptable on physical grounds as argued by Garratt (1979a).

Nevertheless we felt it important to show that the results of this paper are essentially unaltered when the lower values of $d$ for each surface are used. All of the $\Phi_{M,H}^0$ data shown in the Appendix were recalculated for these lower values of $d$ (in fact, for several other values), and for each $L$ range values of $\alpha$ and $z_a$ determined by fitting to Eq. (6). The overall average values for comparison with those previously determined are shown in Table 2.

<table>
<thead>
<tr>
<th>TABLE 2. MEAN VALUES OF $\alpha$ AND $z_a$ FOR ALL DATA SETS BASED ON THE UPPER AND LOWER VALUES OF $d$</th>
</tr>
</thead>
<tbody>
<tr>
<td>$d$ (m)</td>
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<tr>
<td>--------</td>
</tr>
<tr>
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8. Discussion and Conclusions

For a given surface with aerodynamic roughness length $z_0$, current boundary-layer theory shows that, in the adiabatic case, the logarithmic wind law is valid in the lower part of the boundary layer only for $z/z_0 \gg 1$. The lower limit of validity has been particularly stressed by Monin and Yaglom (1971, p. 288); Tennekes (1973) and Townsend (1976, pp. 139–143) and estimated by these authors ‘arbitrarily’ at 50–100 approximately. In the diabatic situation there is no reason to believe that the same general comments do not apply to Eq. (2) and it is relevant that the empirical relations for $\Phi_{M,H}(z/L)$ were determined experimentally in the atmosphere for $z/z_0 > 10^2 - 10^3$.

On this basis our own observations obtained in the range $z/z_0$ from 5 to 19 (for M2) and from 20 to 85 (for M1) would not be expected to satisfy Eq. (2) (of which the logarithmic law is a special case for zero heat flux) over most, if not all, of the height range but to show deviations according to a modified law involving additional scale parameters.
Subsequent analysis of wind and temperature data at two sites confirmed this expectation and modified equations (Eqs. (3), (4), (5) and (6)) utilizing an imposed surface length scale $z_*$ and assuming $z_*$ uniquely determines $z_*$ were developed to describe adequately the observed profiles. The profile deviations (in the form of reduced vertical gradients) were consistent also with the physical picture of surface-generated wakes producing enhanced vertical mixing in the layer immediately above the canopy (e.g. Thom et al. 1975), this being termed the ‘transition layer’ above which wake effects are negligible and the ‘inertial sub-layer’ relations apply.

The analysis showed that the data, viz. the non-dimensional gradients $\Phi_{M,H}^0$, are well represented by the ‘inertial sub-layer’ relations (giving the $z/L$ dependence) modified by an exponential function of $z/z_*$, viz.,

$$\Phi_{M,H}^0 = \{\Phi_{M,H}(z/L)\}^{\{\phi_{M,H}\}}$$

with $\phi_{M,H} = \exp\{-a_1(1-z/z_*)\} \equiv a \exp\{z_1(z/z_*)\}$. This specific form of the function was based on simple dimensional arguments for the length scale associated with the rate of change of $\phi$, implying that the mixing length of an interacting wake field and shear flow is equal to the shear mixing length attenuated by the height functions $\exp\{-a_1(z/z_*)\}$.

The data show that, in unstable conditions, both $a$ and the transition layer depth $z_*$ do not vary significantly with stability. The parameter $a$, equal to the asymptotic value of $\phi$ at the surface, is about 0.5 for both wind and temperature and is independent of surface. In the case of wind, the ratio $z_*/z_0$ has values of 35 and 150 for the denser and less dense surfaces respectively, whilst for temperature $z_*/z_0$ is 100. Comparison of $z_*$ with several natural surface length parameters gives $z_*/\delta = 3$ for both surfaces, suggesting that the spacing may be an important length scale characterizing the wake field and hence the depth of penetration $z_*$. This is probably true only where the underlying surface makes no significant contribution to the total stress (i.e. for $\lambda$ greater than about 0.03).

The assumption of constancy of $z_*/\delta$ for large $\lambda$ means that, for a given spacing, changing $h$ for example has no significant effect upon $z_*$ (the latter being referred to the zero plane) so long as $\lambda > 0.03$. This assumption allows a relevant comment to be made on the results found for the Thetford pine forest ($h = 16.6$ m, $\delta = 3.2$ m, $\lambda = 1.5$) and first published by Thom et al. (1975). Their conclusions were modified subsequently as discussed by Raupach (1979). Briefly, the results of Raupach’s flux-profile analysis at an effective height $z$ of 7-9 m ($Z = 20.5$ m; $d = 12.6$ m) gave, in our nomenclature, for near-neutral conditions $\Phi_M^0/\Phi_M(z/L) \approx 0.9$ and $\Phi_H^0/\Phi_H(z/L) \approx 0.45$, whence he concluded that at such low heights the wind profile was not significantly influenced by the proximity of the surface whereas in contrast this was not the case for the temperature profile. In comparing this to the result of Garratt (1978a) he proposed that the different result for wind was due to major differences in the surface structure.

However, we show that their result is entirely consistent with our results. Thus, taking $\delta = 3.2$ m for Thetford, we infer a value $z_* \approx 9.4$ m (giving $z_*/z_0 \approx 10$) to be compared with the observing height of 7-9 m, i.e. the observations were made at $z/z_* \approx 0.88$, which reference to Figs. 6 and 7 predicts $\phi_M^0 = 0.9-0.95$, close to that measured over the Thetford forest! The Thetford value of $z_*$, together with our own suggestion, suggest that, whilst $z_*/z_0 \approx 100$ is a useful generalization, the ratio probably decreases to about 10 for high density vegetation.

In the case of temperature, their value of $\phi_H^0$ is consistent with a measurement in the transition layer for heat at $z/z_* \approx 0.1-0.2$ according to Fig. 8, suggesting for Thetford forest $z_* \approx 40-80$ m. From this it is apparent that $z_*$ may not depend directly upon surface parameters such as $\delta$, as in the case for $z_*$, although we note that for M1, $z_* \approx$
100 and implied for Thetford is $z_{*H}/z_0 \approx 43$–86. Since $z_0/z_T$ is approximately constant for these surfaces (e.g. Garratt 1978b; $z_T$ is the surface scaling length for temperature), the result suggests, at least tentatively, that $z_{*H}/z_T \approx 500$–750 for tree-covered surfaces.

**APPENDIX 1**

**DETERMINATION OF AERODYNAMIC ROUGHNESS LENGTHS ($z_0$)**

Values of $z_0$ (M1) = 0.4 m and $z_0$ (M2) = 0.9 m used in the paper have been discussed by Garratt (1978a, b) and in Clarke and Brook (1979).

In the inertial sub-layer the equivalent neutral wind $u^N_*$ in the presence of buoyancy is given by

$$ku^N_*/u_* = k(u_*/L) + \Psi_M(z/L) \quad . \quad .$$  \hspace{1cm} (A1)

where the real wind $u_*$ is reduced below $u^N_*$ by an amount $(u_*/k)\Psi_M$. Here

$$ku^N_*/u_* = \ln(z/z_0) \quad . \quad .$$  \hspace{1cm} (A2)

For level 5 winds at M1, M2, mean values of $ku^N_*/u_*$ determined from Eq. (A1) were 4.71 and 3.23 respectively. Then Eq. (A2), with $d/h = 0.06$ and 0.37 for M1, M2 respectively (section 3), gives $z_0$ (M1) = 0.44 m and $z_0$ (M2) = 0.99 m, whilst with $d/h = 0.6$ and 0.75, Eq. (A2) gives $z_0$ (M1) = 0.39 m and $z_0$ (M2) = 0.85 m i.e. $z_0$ is not too sensitive to the value of $d$.

Equations (A1, A2) are valid for the inertial sub-layer, whilst the flux gradient analysis suggests the observations are made within the transition layer where these relations are not valid. However, we now show that, in the case of $z_0$, use of Eqs. (A1, A2) in the transition layer involves only small errors. In any case, such errors can be minimized by a suitable correction.

We assume that within the transition layer the effects of buoyancy upon the real wind may be similarly removed as in Eq. (A1), so that $u_s^N$ is there given by

$$ku^N_*/u_* = \ln(z/z_0) + \Psi^*_M(z/z_0) \quad . \quad .$$  \hspace{1cm} (A3)

This is schematically represented in Fig. 2 of Garratt (1978b) and shows that the transition layer wind is greater by an amount $(u_*/k)\Psi^*_M$ than that given by extrapolation of the inertial sub-layer relation to the appropriate level (in practice, by only a few cm s$^{-1}$).

Equation (A3) may be derived directly from Eq. (3). Integrating Eq. (3) down from $z_*$ to $z$, gives

$$\frac{k}{u_*}(u^N_* - u_*) = \int_{z_*}^{z} z^{-1} \phi(z/z_*) dz$$

whence

$$ku^N_*/u_* = \ln(z/z_0) + \int_{z_*}^{z} z^{-1} \phi(z/z_*) dz$$

$$= \ln(z/z_0) + \int_{z_*}^{z} \{1 - \phi(z/z_*)\} z^{-1} dz.$$

Hence

$$\Psi^*_M(z/z_*) = \int_{z_*}^{z} \{1 - \phi(z/z_*)\} z^{-1} dz.$$

To evaluate $\Psi^*_M$ we note $\phi = x \exp\{x(z/z_*)\}$, and to integrate we expand the exponential function. In fact $\Psi^*_M$ is accurate to 1% for $z/z_* > 0.1$ by considering only the first three terms. After some rearrangement of terms, we have with $\xi = z/z_*$,
\[
\Psi^*_M(\xi) = (x-1) \ln \xi - \alpha x_1(1-\xi) - \alpha x_1^2(1-\xi^2)/6 - \alpha x_1^3(1-\xi^2)/24
\]

Using experimental values, \(x = 0.5\); \(x_1 = 0.7\), yields

\[
\Psi^*_M(\xi) = -0.5 \ln \xi - 0.35(1-\xi) - 0.04(1-\xi^2) - 0.007(1-\xi^3)
\]

(A4)

Equation (A4) both gives a means of determining the error in \(z_0\) if Eq. (A2) is used rather than (A3) and to better estimating \(z_0\) from data within the transition layer using Eq. (A3).

In the case of level 5 winds, \(\xi = 0.725\) in both cases \((z_\phi(M1) = 60 \text{ m}; z_\phi(M2) = 30 \text{ m})\), whence

\[
\Psi^*_M(0.725) \approx 0.045.
\]

Equation (A3) then gives \(z_0(M1) = 0.41 \text{ m}\) and \(z_0(M2) = 0.89 \text{ m}\), not significantly different from the other estimates.

Note that for \(u_\phi = 0.5 \text{ m s}^{-1}\), the exess wind \((u_\phi/k)\Psi^*_M \approx 0.06 \text{ m s}^{-1}\) at \(\xi = 0.725\), in a mean wind of about 6 m s\(^{-1}\). (At \(\xi = 0.25\), the exess wind is 0.49 m s\(^{-1}\) in a mean wind of about 4.5 m s\(^{-1}\)).

As a final comment we have shown that Eqs. (3, 4, 6) describe adequately the gradient observations; likewise Eqs. (A3, A4) describe the actual winds equally so.

**APPENDIX 2**

Data from sites M1, M2 in the form of mean \(\Phi_H^0\) and \(\Phi_H^0\) values for narrow \(|L|\) ranges over four air layers at each site are presented. Basic observational data, to be found in Clarke and Brook (1979), have been selected as described in the text. For M1, \(d_{u,0} = 4.8 \text{ m}\) and for M2, \(d_u = 7.1 \text{ m}\). Also shown are the standard deviations of individual data about the respective means (\(\sigma\)), the maximum number of runs used (\(n\)) for any one layer and the range of \(L\) in each case. Codes (numbers, lettering) refer to data presented in Figs. 6, 7, 8.

Close inspection of \(\Phi_M^0, \Phi_H^0\) values at each site shows a non-monotonic variation with height – this is by no means unexpected.

One’s expectancy upon gradient behaviour depends upon the situation, i.e. the physical variables upon which the gradient may depend. Thus in a study of vertical profiles near sunset or downwind of a surface transition, gradients will minimize or maximize at some level. In the former example the gradient depends not on one variable but at least two, i.e. height and time. In the latter example the gradient again depends upon at least two variables, height and fetch.

However, these examples are excluded if we require steady-state, homogeneous conditions. Thus in the most relevant example, that of the inertial sub-layer, the gradient will depend upon one variable only (the height) and we expect monotonically decreasing absolute gradients with height. (Note here that \(u_\phi\) and \(L\) are independent of height, so that \(\partial u/\partial z\) or \(\Phi\) depends only upon height on any one occasion.) On this basis then inertial sub-layer observations should have \(u\) increasing and \(\partial u/\partial z\) decreasing with height. Acceptance of data on any one occasion according to the first criterion is usual; not so application of the second. With a 1% accuracy in \(u\), \(\partial u/\partial z\) is typically 10% accurate, and random errors in an anemometer at any one level can give a non-monotonic variation in \(\partial u/\partial z\) with height. However, when averaged over the whole profile, and for a number of runs, such effects are minimized and one then does expect monotonically decreasing gradients with height.

In the transition layer the gradient still depends upon \(z\) only on any one occasion and so we expect, on average (e.g. for each \(L\) range as presented in this Appendix), monotonically decreasing values of \(\partial u/\partial z\) with height. Our data show this exclusively (e.g. see data tables in Clarke and Brook 1979).
The situation is somewhat different if we consider the dimensionless gradient $\Phi^0$. In the inertial sub-layer it depends upon one variable only, $z/L$, and either remains constant with height (in neutral conditions) or decreases (in unstable conditions). In the transition layer, by its very nature, $\Phi^0$ depends upon two variables at least, i.e. $z/L$ and $z/z_*$, and need not necessarily vary monotonically with height. Indeed, use of the relation described in the text implies a level ($z < z_*$) where

$$z = (5.7 z_* - L)/16$$

at which $\partial \Phi^0/\partial z = 0$; precisely the nature revealed in the tabulated $\Phi^0$ values. Monotonic behaviour should, however, be expected in the dimensionless quantity

$$\Phi^0/\Phi(z/L)$$

since this depends upon one variable, $z/z_*$ only. Reference to the data allows this quantity to be evaluated; they are actually plotted in Figs. 6–8.

Thus for M2 temperature we find, from 16 cases, 12 show a monotonic increase; for M2 wind, 12 again. For M1 wind, layer 3–4 shows a systematic low value probably related to a systematic error in anemometer at level 4; otherwise from 17 cases, 15 show an increase from layer 1–2, to 2–3, to 4–5, and from 3–4 to 4–5.

To summarize, in the inertial sub-layer or transition layer $\partial u/\partial z$ and $\partial \theta/\partial z$ depend upon one variable $z$ only. The data satisfy the requirement of monotonically decreasing values with height. In the case of $\Phi^0$, this depends upon two variables in the transition layer and monotonic behaviour is not expected. In the case of $\Phi^0/\Phi(z/L)$ which depends upon one variable, $z/z_*$, monotonic behaviour is predominantly observed (as evidenced in Figs. 6–8).

<table>
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<th>$L$ range (m)</th>
<th>Code</th>
<th>$n$</th>
<th>$\Phi^0_M$</th>
<th>$\sigma$</th>
<th>$\Phi^0_M$</th>
<th>$\sigma$</th>
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<th>$\sigma$</th>
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### M1 - Temperature Profile

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### M2 - Wind Profile

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### References


Yaglom, A. M. 1974