The structure of turbulence in a coastal atmospheric boundary layer

By YAPING SHAO, JÖRG M. HACKER and PETER SCHWERDTFEGER
Flinders Institute for Atmospheric and Marine Sciences G.P.O. Box 2100, Adelaide 5001, Australia

(Received 20 May 1990, revised 28 March 1991)

SUMMARY
This study, based on airborne observations obtained during the Upper Spencer Gulf experiments in South Australia, investigates the structure of turbulence in a coastal boundary layer, which represents an extreme situation of horizontal inhomogeneity. The development of the coastal boundary layer under mainly onshore wind conditions, which is characterized by the formation of a convective thermal internal boundary layer (TIBL), is illustrated using cross-sections of potential temperature, sensible-heat fluxes and turbulent kinetic energy. The similarity of turbulence within the TIBL and the effectiveness of local scaling for inhomogeneous conditions are examined. It is suggested that, even in advective conditions, turbulence can be considered as being in local equilibrium. By introducing a simple TIBL model, local similarity relationships can be linked to external parameters. Some turbulence properties, such as the variance of vertical velocity, are found to be determined by three external parameters: the surface heat flux over land, the onshore wind speed and the thermal stability of the approaching airflow.

1. INTRODUCTION

Atmospheric boundary layers in coastal areas, associated with intense contrasts between surface properties of water and land, are typical cases of horizontal inhomogeneity. Here, a step change in surface thermal properties is often the governing factor which causes marked transformations in the airflow across the coastline. The boundary-layer structure is characterized by the development of a thermal internal boundary layer (TIBL), the upper boundary of which distinguishes transformed from untransformed airmasses. With the growing interest in dispersion problems related to sources of air pollution in coastal areas, TIBLs have been subject to several experiments carried out in recent years. Representative studies are those of Raynor et al. (1979), Gamo et al. (1982) and Durand et al. (1989). Apart from these observational studies, theoretical investigations and numerical modelling have been complemented by, for instance, Venkatram (1977, 1986) and Durand et al. (1989).

The structure of turbulence is an important aspect of coastal boundary-layer studies. This is not only because turbulence adjustment in coastal areas is an interesting physical phenomenon, but also because the structure of turbulence needs to be specified in dispersion models, especially in Lagrangian diffusion models. Available statistics (e.g. Durand et al. 1989; Shao and Hacker 1990) have shown that turbulence properties in coastal boundary layers vary in a complex way with time and in space. As a consequence, studies carried out under different conditions are difficult to compare and results obtained from a particular experiment are difficult to generalize. It is therefore required to develop a framework, within which turbulent properties can be compared and the results obtained from particular experiments can be generalized. Attempts to test the possible validity of the existing similarity theories under inhomogeneous conditions have been made by, for instance, Smedman and Högström (1983) and Durand et al. (1989).

In coastal boundary layers, horizontal advection cannot be neglected, but the importance of advection must be considered distinctively for different moments of turbulence. For instance, first-moment budgets include no local production terms and are only balanced between advection and transport, while second-order and higher-order moment budgets are dominated by local production and destruction terms in most circumstances. Lenschow et al. (1980) showed that although advection plays an important
role in the budgets of heat and momentum, and its influence on the behaviour of first-order moments cannot be neglected, it has only a limited effect on the behaviour of second-order moments. Shao and Hacker (1990) assumed that even in coastal areas the direct advective influence on some second-order and higher-order moments is not important and turbulence is in local equilibrium. The results presented in Shao and Hacker confirmed that the local similarity theory can be applied to describe the behaviour of some second-order and higher-order moments. However, since the scaling parameters have to be determined locally, the applicability of the local similarity theory is limited. It is thus necessary to express the local scaling parameters and the corresponding similarity relationships in terms of external parameters which determine the basic physical properties of the coastal boundary layer.

The structure of a coastal boundary layer and the similarity of turbulence in the TIBL are investigated in this study on the basis of airborne observations. It is suggested that the behaviour of turbulence within the TIBL is similar, and local scaling, which is effective in describing the behaviour of turbulence in homogeneous CBLs (convective boundary layers), can also be applied to TIBLs. Following a brief review of local scaling, the structure of the coastal boundary layer and its development have been visualized using cross-sections of potential temperature, $\theta$, sensible-heat flux, $H$, and turbulent kinetic energy, $e$. The similarity relationships for some important turbulence quantities are determined from the airborne observations and compared with those obtained for homogeneous CBLs. It is shown that, by introducing a simple model for the TIBL, turbulence properties within the TIBL can be related to three external parameters: the surface sensible-heat flux over land, the onshore wind speed and the stability of the onshore airflow. It is believed that these similarity relationships are relevant for dispersion models, in which the characteristics of turbulence have to be specified.

2. LOCAL SCALING FOR CBLS OVER UNIFORM SURFACES

The local similarity hypothesis for CBLs over uniform surfaces is derived from the local free convection prediction (Wyngaard et al. 1971; Wyngaard 1973) and is in principle a reformulation of the mixed-layer similarity hypothesis (Sorbjan 1986). Local similarity theory assumes that turbulence properties are in equilibrium with local governing parameters, and predicts that the dimensionless groups formed with the scaling parameters in Set (1):

\[
\begin{align*}
\text{Velocity scaling:} & \quad w_{**} = (w'\theta' z g / T)^{1/3} \\
\text{Temperature scaling:} & \quad \Theta_{**} = w' \theta' / w_{**} \\
\text{Length scaling:} & \quad z
\end{align*}
\]

are constant with height, $z$; where $T$ is temperature, $g$ the acceleration due to gravity and $w'$ and $\theta'$ are deviations from mean vertical velocity and potential temperature, respectively. For $\sigma_w^2$ (variance of vertical velocity) and $\sigma_{\theta}^2$ (variance of potential temperature), for instance, the similarity relationships can be formulated as

\[
\begin{align*}
\sigma_w^2 / w_{**}^2 & = k_w \\
\sigma_{\theta}^2 / \Theta_{**}^2 & = k_{\theta}
\end{align*}
\]

where $k_w$, $k_{\theta}$ are constants. These predictions should be understood as the asymptotic case of $z/\Lambda \to -\infty$, where $\Lambda$ is the local Monin–Obukhov length (Nieuwstadt 1984). This is because the shear production of turbulence is excluded from the scaling procedure,
which is identical to supposing $\Lambda = 0$. Therefore, the local free-convection prediction should, in principle, only be effective for very unstable conditions.

Mixed-layer scaling has the advantage that the scaling parameters are constructed by a few external parameters, whereas for local scaling the scaling parameters need to be derived locally. However, under certain circumstances, local and mixed-layer scaling have simple relationships (Sorbjan 1986). Following Deardorff (1970) and Tennekes (1970), the scaling parameters for well-mixed CBLs are given by

Velocity scaling: $w_* = \left( w' \theta_0 z_i g / T \right)^{1/3}$

Temperature scaling: $\Theta_* = w' \theta_0' / w_*$

Length scaling: $z_i$

where $z_i$ is the inversion height. Mixed-layer similarity theory predicts that the dimensionless variables formed with these parameters are functions of $z / z_i$ only. It is well known that the profile of turbulent heat flux in CBLs can be approximated by

$$ H = H_o (1 - az / z_i) \quad (3) $$

with $a$ being a constant of about 1.2. Using (3), it is found that

$$ w_* = w_*(z / z_i)^{1/3} (1 - az / z_i)^{1/3} \quad (4) $$

$$ \Theta_* = \Theta_*(z / z_i)^{-1/3} (1 - az / z_i)^{2/3} \quad (5) $$

Thus, (1) and (2) are identical to

$$ \sigma_*/w_*^2 = k_w(z / z_i)^{2/3} (1 - az / z_i)^{2/3} \quad (6) $$

$$ \sigma_/\Theta_*^2 = k_\Theta(z / z_i)^{-2/3} (1 - az / z_i)^{4/3} \quad (7) $$

which are functions of $z / z_i$ only. Unfortunately, these predictions are useful only for the lower half of the CBL, since (1) and (2) are not valid for the upper part where turbulent heat flux is small or negative and the influence of entrainment can not be neglected.

Turbulence partitioning and the concept of ‘top-down’ and ‘bottom-up’ diffusion (Wyngaard and Brost 1984) lead to a new development in the local scaling of convective turbulence (Sorbjan 1988), which effectively remedies the failure of local scaling in the upper part of the CBL. Sorbjan suggested that both the local buoyancy flux and a physical quantity of interest can be partitioned into two parts associated with the ‘top-down’ and ‘bottom-up’ processes (to be denoted by subscripts $t$ and $b$, respectively) and the two parts of the physical quantity are constant when scaled with the corresponding component of the local scaling parameter. For $\sigma_*, for instance, it follows that

$$ \bar{w'} \theta' = \bar{w'} \theta'_t + \bar{w'} \theta'_b $$

$$ \sigma^2_w = \sigma^2_{wb} + \sigma^2_{wt} $$

and

$$ \sigma^2_{wb} / w_*^2 = k_{wb} \quad (8) $$

$$ \sigma^2_{wt} / w_*^2 = k_{wt} \quad (9) $$

where $k_{wb}$ and $k_{wt}$ are constants and $w_*^2$ and $w_*^2$ are the components of the local scaling velocity defined as

$$ w_* = \left( \frac{g}{T} zw' \theta'_b \right)^{1/3} \quad (10) $$
\[ w_{*t} = \left( \frac{g}{T} (1 + D)(z_i - z)w'r' \right)^{1/3} \]  

(11)

with \( D \) being the thickness of the interfacial layer above the CBL, normalized with \( z_i \).

Thus, mixed-layer scaling can be expressed in terms of local component scaling

\[ \frac{\sigma_{w}^2}{w_{*t}^2} = k_{wb} \left( \frac{w_{*b}}{w_{*t}} \right)^2 + k_{wi} \left( \frac{w_{*i}}{w_{*t}} \right)^2. \]  

(12)

For a linear partitioning of local buoyancy flux

\[ \bar{w}' \theta'_{b} = \bar{w}' \theta'_{0} (1 - z/z_i) \]  

(13)

\[ \bar{w}' \theta'_{i} = \bar{w}' \theta'_{0} z/z_i \]  

(14)

with \( r \) being a constant of about \(-0.2\) (Tennekes 1973), it can readily be shown that

\[ \frac{\sigma_{\theta}^2}{w_{*t}^2} = k_{wb} (z/z_i)^{2/3} (1 - z/z_i)^{2/3} + k_{wi} r^{2/3} (z/z_i)^{2/3} (1 - z/z_i + D)^{2/3}. \]  

(15)

Similar predictions can be made for other turbulence quantities. The validity of 'top-down' and 'bottom-up' scaling for CBLs over uniform surfaces was tested and confirmed by Hartmann (1990). It will be shown later in this study that local scaling is also effective in describing the turbulence behaviour within the TIBL, although its formation is closely related to horizontal advection; but first the structure of the coastal boundary layer and the development of the TIBL will be examined.

3. THE USG EXPERIMENTS

The data-set used in this study was obtained during the USG (Upper Spencer Gulf, South Australia) experiments carried out during the Australian summers of early 1987 and 1988 using the instrumented aircraft of the Flinders Institute for Atmospheric and Marine Sciences—a GROB G109B. A detailed description of this aircraft, its instrumentation and the procedure of data acquisition are given by Hacker and Schwerdtfeger (1988), while the design of the USG experiments is described in Shao and Hacker (1990).

Figure 1 illustrates the site for the experiment and the flight pattern. The measurements took place on the eastern side of the Gulf near Mambray Creek (32°52'S, 137°56'E). In this area, the Gulf is about 10 km wide with an average water depth of about 7 m. The beach area extends for about 2 km and is partly covered by very shallow water to an extent that depends on the tides. The adjacent land is flat for about 20 km before rising steeply towards the southern Flinders Ranges in the east. During the periods of observation, the land surface consisted of dry farmland (harvested wheat fields) and uncultivated areas covered with small bushes about 0.5 m high. The boundary layer here is essentially two-dimensional, since the surface conditions parallel to the coastline are uniform.

The airborne measurements were carried out mostly in onshore wind conditions. The flight pattern as shown schematically in Fig. 1 consisted of traverses across the coast and runs parallel to it at four levels (20, 60, 120 and 240 m) above water and terrain. The across-coast traverses (denoted by the letter T in Fig. 1), covering about 13 km, were flown to gain an overview of the boundary-layer structure. To obtain accurate statistics and turbulent fluxes, runs of about 10 to 12 km parallel to the coastline were flown progressively over water, beach and land along tracks P1, P2, P3 and P4 (Fig. 1). During the observations, the aircraft was flying at a true airspeed of about 40 m s\(^{-1}\), while the data sampling rate was 13 Hz. The whole pattern, including all traverses and runs, was
Figure 1. Site and flight pattern of the Upper Spencer Gulf experiment. For further explanations see text.
flown once or twice per day, as summarized in Table 1 of Shao and Hacker (1990). In addition, temperature and wind velocity at 1, 2, 5 and 10 m above ground were measured on a 10 m mast located 3 km onshore (Fig. 1).

4. STRUCTURE AND DEVELOPMENT OF THE COASTAL BOUNDARY LAYER

Marked transformations occur in an airflow moving across a coastal surface discontinuity. It is typical of coastal regions that a stably stratified airflow approaching the adjacent land is heated from below and becomes convective in the lower part, which is usually known as the TIBL. The development of turbulence within the TIBL is responsible for intensive vertical transport of heat, momentum and mass.

As a background for the development of the boundary-layer structure in the USG region, the thermal contrast between water and land encounters diurnal variations accompanied by the intensification and decay of the onshore wind speed associated with the development of a shallow sea-breeze system. An example is given in Fig. 2 where the time series of temperature, wind speed and direction observed on the mast are shown. From midnight to early morning, the air was stably stratified near the surface and the temperatures at all levels decreased steadily from about 21°C to a minimum of about 15°C at 0830 h. In this period, the wind speed decreased and its direction changed from SW to SE. During the following two hours, the temperatures rose by nearly 10 degC while small changes in wind speed and direction occurred. The passage of a sea-breeze front registered at 1030 h was marked by a sudden increase in the wind speed to 4 m s⁻¹ and a change in the wind direction to SW. At the same time, an unstable stratification was established which gradually intensified during the rest of the day with wind speed increasing until the late afternoon. Around 1900 h, temperature and then wind speed began to decrease rapidly until a nocturnal surface layer was established around 2100 h.

In Figs. 3, 4 and 5, cross-sections of θ, H and e are shown. These cross-sections are derived from the cross-coast traverses (pattern 'T' in Fig. 1, which was completed within a half-hour period) by interpolating the aircraft data onto a regular grid using smooth surface fitting. The series from which the cross-sections were derived are also shown for comparison.

(a) The thermal field

In Figs. 3 and 4, the θ and H cross-sections for USG 5 and three other successive observations (USG 8, USG 10 and USG 11, see also Fig. 2) are presented to illustrate the development of the thermal structure of the coastal boundary layer. Although USG 5 was performed on the day previous to the other three observations, they can be considered as comparable, since the synoptic situation remained unchanged. The four observations represent different phases of the development of the coastal boundary layer, corresponding to the thermal contrast of the surfaces (Table 1).

USG 5, carried out around sunrise (0600 LMSST), is representative for the boundary layer in nocturnal situations. Before sunrise, the water surface (22°C) was about 10 degC warmer than the land surface (12°C). Figure 3(a) shows that the air over land was warmer than the ground and stably stratified, while the air over water was cooler than the surface, thus unstable and more turbulent. Over land, H was nearly zero, but exceeded 50 W m⁻² over water (Fig. 4(a)). Dramatic changes in the boundary-layer thermal structure occurred during the two hours after sunrise. Corresponding to a rapid increase in the land surface temperature, a shallow convective layer formed over the land, with an upper
Figure 2. Time series of 10-minute averages of temperature and wind speed at heights of 1, 2, 5 and 10 m measured on the 10 m mast from 0000 LMST, 1 March to 0000 LMST, 2 March 1988. The wind direction was measured at the 7 m level. Airborne measurements were carried out during the shaded time-periods.

TABLE 1. CONTRAST IN WATER AND LAND SURFACE TEMPERATURES

<table>
<thead>
<tr>
<th>USG</th>
<th>T_{water}(°C)</th>
<th>T_{beach}(°C)</th>
<th>T_{land}(°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>5</td>
<td>22</td>
<td>18</td>
<td>12</td>
</tr>
<tr>
<td>8</td>
<td>22</td>
<td>23</td>
<td>33</td>
</tr>
<tr>
<td>10</td>
<td>22</td>
<td>30</td>
<td>50</td>
</tr>
<tr>
<td>11</td>
<td>22</td>
<td>28</td>
<td>43</td>
</tr>
</tbody>
</table>

bound being no less than 120 m high, as marked in Figs. 3(b) and 4(b). The sensible-heat flux was positive in the convective region with a maximum of 75 W m\(^{-2}\) measured for the lowest traverse. Since a very weak offshore wind or wind parallel to the coast prevailed during the observations, Figs. 3(b) and 4(b) represent a transient period when a stable layer over water and a convective layer over land coexisted.

The deformation in the thermal structure of the onshore airflow during the day is depicted in Figs. 3(c), 4(c) and Figs. 3(d), 4(d). Using \( H \) as a criterion, the coastal boundary layer can be divided into an outer region and an inner region (TIBL), as
Potential Temperature in °C

Figure 3. Cross-sections of potential temperature for USG 5 (0620 LMST, 29 February 1988), USG 8 (0820 LMST, 1 March), USG 10 (1320 LMST, 1 March) and USG 11 (1520 LMST, 1 March). The upper boundary of the TIBL is marked with the thick dashed line. The space series of potential temperature, from which the cross-section is derived, are also shown. The regions with rapid increase in turbulent kinetic energy are shaded.
Figure 4. As Fig. 3 but for sensible-heat flux.
Turbulent Kinetic Energy in $Jm^{-3}$

Figure 5. As Fig. 3 but for turbulent kinetic energy.
indicated by the thick dashed line defined by $H = 0$. The outer region was stably stratified with $\theta$ increasing with height and the isotherms running horizontally. Here, $H$ was small or slightly negative. The TIBL was nearly well mixed, with $\theta$ being constant with height and with its gradient pointing onshore. The sensible-heat flux, with a maximum of about 350 W m$^{-2}$ (for USG 10) and 175 W m$^{-2}$ (for USG 11), decreased with height within the TIBL. As indicated by the thick dashed line in Figs. 3(c), 4(c) and Figs. 3(d), 4(d), the depth of the transformed layer or TIBL increased gradually with fetch.

(b) The turbulence field

In conjunction with the development of the boundary-layer thermal structure, pronounced changes occurred in the turbulence field, as illustrated by the $e$ cross-sections (Fig. 5). Corresponding to the unstable stratification over the water at sunrise (Fig. 5(a)), relatively strong turbulence was observed there. Over land, light turbulence prevailed in low levels, while areas of stronger turbulence were found at higher levels (120 m to 240 m), possibly related to a local system. With the establishment of the shallow convective layer (Figs. 3(b), 4(b)), the intensity of the turbulence increased over land during the following two hours with $e$ reaching 1 J m$^{-3}$, but decreased over water. In the next observation (Fig. 5(c)), turbulent activity increased significantly and $e$ over land exceeded 2.5 J m$^{-3}$. Turbulence began to decay in the early afternoon (Fig. 5(d)) with the maximum of $e$ being less than 2 J m$^{-3}$.

Figures 5(c) and 5(d) are typical for well-established TIBLs. In Fig. 5(d), for example, while $e$ was less than 0.25 J m$^{-3}$ in the untransformed outer region, it increased rapidly in a narrow area close to the coastline, until reaching more than 1.5 J m$^{-3}$ over land. Within the TIBL, an equilibrium with the new environment was eventually achieved and the increasing rate of $e$ was reduced. In most observations, a narrow region with $e$ rapidly increasing can be identified (shaded area in Figs. 3, 4 and 5). Hence, using $e$ as a criterion, a turbulent IBL can be defined as the region of high-turbulence levels with an ‘$e$-front’ as its upper limit. Similar phenomena were observed by Raynor et al. (1979) and Durand et al. (1989).

Although turbulent fluctuations are closely related to the boundary-layer thermal structure, ‘$e$-fronts’ extended to levels higher than the upper boundary of the TIBL identified from the boundary-layer thermal structure (thick dashed line in Figs. 3 and 4). Gamo et al. (1982) also found that the turbulent IBL is about 1.4 times deeper than the TIBL. As can be seen from Figs. 4(c) and 4(d), a negative sensible-heat flux prevailed in the ‘$e$-front’ region. Although detailed budget studies of energy would be required for a better understanding of this interfacial area between the outer region and the well-mixed region, it is plausible to assume that the ‘$e$-front’ area is similar to the upper part of CBLs (where $H < 0$) in non-advective conditions, where turbulence is maintained by penetrative thermals when $H$ becomes negative. It is also possible that the rapid increase in $e$ is related to internal gravity waves excited at the top of the TIBL.

Pure waves were observed in the approaching stable airflow on several occasions. Entering the convective region over land, these waves were usually swamped by the convective turbulence. An example is presented in Fig. 6, where waves with wavelengths of from 500 m to 600 m and maximum amplitude of 0.3 m s$^{-1}$ in w-components (0.1 degC in $\theta$) were the dominating phenomenon in the onshore airstream. An estimation of the Brunt–Väisälä frequency $N$, which gradually decreased further inland, indicated that the stratification was becoming less stable in the onshore direction. At about 8 km onshore, the waves were eventually swamped by the penetrative convective turbulence.
Figure 6. An example of waves swamped by penetrative turbulence. High-pass filtered space series of $w$ and $\theta$ and Brunt-Väisälä frequency $N$ are plotted versus distance from the water line.

5. TURBULENCE CHARACTERISTICS

Turbulence statistics from the USG experiments have shown that turbulent fluctuations within the TIBL are ten times more intense than those in the outer region (Shao and Hacker 1990). Apart from the intensity, other distinguishing contrasts between turbulence in either part of the TIBL can be identified from the space series. Figure 7 shows sections of space series of $\ddot{w}$, $\ddot{\theta}$ and $\ddot{q}$ ($q$ is specific humidity), as well as instantaneous covariances $\dot{w}\dot{\theta}$ and $\dot{w}\dot{q}$, where $\dot{x}$ is defined as $x'/\sigma_x$ with $\sigma_x$ being the standard deviation of $x$. Figure 7(a) is typical for the outer region (P1(120 m) USG1), while Fig. 7(b) (P3(120 m) USG6) is representative for the TIBL.

Within the TIBL (Fig. 7(b)), turbulence is characterized by identifiable discrete thermals. Intense updraughts of $\ddot{w}$ (bursts) are associated with peaks of $\ddot{\theta}$. In the immediately adjacent larger areas, downdraughts (sweeps) of relatively cool elements prevail. As can be seen from $\dot{w}\dot{\theta}$, which is proportional to the instantaneous sensible-heat flux, the convective elements represent the most effective channel of energy transfer in the vertical. Humidity fluctuations show considerably different characteristics from those of potential temperature and vertical velocity. Figure 7(b) shows that the positive humidity fluctuations have a 'top-hat' appearance, while the negative fluctuations are dry bursts correlated with the downdraughts, causing intense instantaneous upward latent-heat fluxes. In the upper part of well-mixed layers, Crum et al. (1987) also reported the 'top-hat' appearance, but the dry bursts were not obvious in their observations.

In the outer region, turbulence and wave motions coexist. Figure 7(a) shows that the updraughts and downdraughts are randomly embedded in the oscillations on a larger scale. In the present example, waves with a wavelength of about 1 km can be identified in the traces of $\theta'$ and $q'$ although not in $w'$. Unlike in the TIBL, $\ddot{w}$ is hardly correlated with $\ddot{\theta}$ and $\ddot{q}$. Hence, the updraughts and downdraughts produce no effective transport of energy in the vertical.
Figure 7. Examples of series of $\dot{\theta}$, $\ddot{\theta}$, $\hat{w}$, $\hat{q}$ and $\ddot{\hat{w}}$: (a) for stable conditions (P$_s$(120 m) USG1); (b) for convective conditions (P$_s$(120 m) USG6).
The evolution of turbulence can be conveniently studied using probability density functions (PDF). Representative PDFs for $\dot{w}$ ($p(\dot{w})$) as presented in Fig. 8(a) show the adjustment of turbulence in the onshore direction. Outside of the TIBL (thin curves in Fig. 8(a)), $p(\dot{w})$ is approximately normal-distributed, indicating the symmetric fluctuations in the untransformed airflow. In contrast, within the TIBL, $p(\dot{w})$ is asymmetric with a mode of $-0.3$ to $-0.5$. This implies that the most frequent value of $w'$ is about $-0.3$ to $-0.5 \sigma_w$ in the convective region. From the continuity principle, it can be expected that the less frequent updraughts must be more intense than the downdraughts, as already discussed within Fig. 8(b). These results agree with those by Caughey et al. (1983) obtained in homogeneous convective conditions. On most occasions, $p(\dot{w})$ does not vary significantly within the TIBL in the onshore direction.

Temperature fluctuations show clear changes from stable to convective conditions as illustrated in Fig. 8(b), where the variation of $p(\dot{\theta})$ in onshore direction is presented.

Figure 8. (a) Probability density of vertical velocity $p(\dot{w})$ determined from the 16 runs parallel to the coast for USG1 plotted as four graphs according to the height of the runs. Thicker lines represent the runs inside the TIBL. (b) Same as (a), but for probability density of potential temperature $p(\dot{\theta})$. 
In general, in the outer region (thin curves) \( p(\hat{\theta}) \) is symmetrical and the mode is close to the linear average, while within the TIBL it is obviously asymmetric with a negative mode. In the TIBL, \( \theta' \) is rarely smaller than \(-1.5\sigma_\theta\) and its most probable value is around \(-0.8\sigma_\theta\).

The asymmetry of the PDFs is described by their skewness which is also an important parameter in dispersion models. Skewnesses of \( w, S_{k_w}, \) and \( \theta, S_{k_\theta} \), averaged over 8 observations show that \( S_{k_w} \) is a constant of value 0.2 in the untransformed airflow, while, within the TIBL, \( S_{k_w} \) falls between 0.3 and 0.7. The magnitude of \( S_{k_\theta} \) and its variation in the vertical are comparable with those observed in the convective boundary layers over uniform surfaces (Lenschow et al. 1980) and large-eddy simulation (Moeng and Wyngaard 1988). In the outer region, \( S_{k_\theta} \) falls between \(-0.3\) and 0.2 on average and increases dramatically to around 1.3 in the TIBL.

6. Similarity of Turbulence within the TIBL

In a recent study, Shao and Hacker (1990) tested the effectiveness of local scaling in the coastal boundary layer and found that some turbulence properties obey the local similarity prediction. They supposed that even in coastal areas the direct advective influence on some second-order and higher-order moments is not important and turbulence is in equilibrium with the local forcing represented by the scaling parameters. The advective influences on local forcing were however not neglected, since the scaling parameters were derived from local measurements. The observations from the USG experiments showed that the local free-convection prediction is effective for the runs within the TIBL, if the sensible-heat flux is sufficiently large. It has been shown in Shao and Hacker (1990) and Shao (1990) that \( \sigma_w^2 \) and \( \sigma_\theta^2 \) obey the local free-convection predictions (1) and (2) while the turbulent transport of \( \sigma_w^2 \) can be expressed as

\[
\overline{w'^3} = k_{w^3}/w_{w^3}
\]

with \( k_w = 1.8, k_{\theta} = 1.4 \) and \( k_{w^3} = 1.2 \), respectively. Although these results confirmed that the local similarity hypothesis is valid for some second-order and higher-order moments, its applicability is limited because the scaling parameters have to be determined locally. It is shown in the rest of this paper that the local scaling parameters and the corresponding local similarity relationship can be expressed with external parameters which determine the basic physical properties of the coastal boundary layer. This is realized for the convective part of the coastal boundary layer by introducing a simple model for the TIBL.

(a) A simple model for the TIBL and similarity predictions

Based on the observational results presented in section 4, the coastal boundary layer can be simplified to a conceptual model schematically depicted in Fig. 9 (see also Venkatram 1986). Since the surface conditions in the direction parallel to the coastline are uniform, the boundary layer is essentially two-dimensional in the onshore (\( x \)) and vertical (\( z \)) directions. It is supposed that the onshore airflow is stably stratified over water, with \( \gamma = \partial \hat{\theta}/\partial z \) being positive and \( \partial \hat{\theta}/\partial x \) zero. With the TIBL it is well mixed in the vertical with \( \gamma = 0 \) and \( \partial \hat{\theta}/\partial x > 0 \). The surface heat flux, \( H_o \), is assumed to have a step change, from zero over water to a constant positive value over land. Although Venkatram (1986) pointed out that it is difficult to believe that \( H_o \) is constant over land in the onshore direction, this assumption seems to be confirmed in several studies (e.g. Kerman et al. 1982). Within the TIBL, \( H \) is supposed to decrease linearly with height.
and vanishes at the upper boundary of the TIBL, i.e.

\[ H(x,z) = H_0 (1 - z/\zeta) \]  \hspace{1cm} (17)

where \( \zeta = \zeta(x) \) is the height of the TIBL. As indicated by the observations (Figs. 3, 4), there exists an entrainment layer with prevailing negative heat fluxes above the TIBL. Despite uncertainties about the entrainment rate and the depth of such a layer, its existence is widely accepted (Venkatram 1977, 1986; Smedman and Högström 1983; Durand et al. 1989).

Following the algorithm of mixed-layer scaling, the velocity and temperature scaling parameter can be introduced as

\[ w_\ast = \left( \frac{g}{T} \frac{w' \theta_0}{\zeta_1} \right)^{1/3} \]

\[ \Theta_\ast = \frac{w' \theta_0}{w_\ast} \]

where \( \zeta_1 = 1.2 \zeta \) is used to keep the formulation consistent with mixed-layer similarity theory. It follows that

\[ w_\ast' = w_\ast (z/\zeta_1)^{1/3} (1 - 1.2z/\zeta_1)^{1/3} \]  \hspace{1cm} (18)

\[ \Theta_\ast' = \Theta_\ast (z/\zeta_1)^{-1/3} (1 - 1.2z/\zeta_1)^{2/3}. \]  \hspace{1cm} (19)

If the entrainment process in the interfacial region is significant, the algorithm of mixed-layer partitioning can be followed with

\[ \frac{w_\ast}{w_\ast} = \frac{w_\ast}{w_\ast} (1 - z/\zeta_1) \]  \hspace{1cm} (20)

\[ \frac{w_\ast}{w_\ast} = \frac{w_\ast}{w_\ast} z/\zeta_1 \]  \hspace{1cm} (21)

and

\[ w_{\ast l} = w_\ast (z/\zeta_1)^{1/3} (1 - z/\zeta_1)^{1/3} \]  \hspace{1cm} (22)

\[ w_{\ast r} = w_\ast (rz/\zeta_1)^{1/3} (1 - z/\zeta_1 + D)^{1/3} \]  \hspace{1cm} (23)

\[ \Theta_{\ast l} = \Theta_\ast (1 - z/\zeta_1)^{1/3} (z/\zeta_1)^{-1/3} \]  \hspace{1cm} (24)

\[ \Theta_{\ast r} = \Theta_\ast (rz/\zeta_1)^{1/3} (1 - z/\zeta_1 + D)^{-1/3}. \]  \hspace{1cm} (25)
Since $\zeta$ varies in the onshore direction, $w^*_{\Theta}$ and $\Theta^*_{\Theta}$ and therefore also $w^*_{\Theta br}$, $w^*_{\Theta tr}$, $\Theta^*_{\Theta br}$ and $\Theta^*_{\Theta tr}$ are not independent of fetch. This also represents the major difference between the TIBL and a homogeneous CBL, for which the scaling length, $z_i$, is a constant. In fact, $\zeta$ is the only parameter which reflects the horizontal inhomogeneity in the scaling procedure. Clearly, the remaining task in describing turbulence properties within the TIBL in terms of external parameters is to derive an appropriate expression for $\zeta$ or $\zeta$ related to these parameters.

This can be achieved by using the heat-budget equation. In stationary conditions, the energy equation describing the basic feature in the coastal boundary layer can be simplified to

$$\frac{-a^3}{w} + \frac{\partial w^*}{\partial z} = 0$$

for a two-dimensional problem in the $x-z$ plane. Integrating the above equation from the ground to $\zeta$, supposing that $H(x, \zeta) = 0$ and that $\partial \theta / \partial x$ is independent of $z$, it follows that

$$U_m \frac{\partial \theta}{\partial x} = H_o / \rho c_p$$

where $U_m$ is the depth-mean wind velocity over $\zeta$. From Fig. 9 it is readily found that

$$\frac{\partial \theta}{\partial x} = \frac{\partial \zeta}{\partial x}$$

and that

$$\zeta \frac{\partial \zeta}{\partial x} = \frac{H_o}{\rho c_p U_m \gamma}$$

Since the right-hand side of the above equation is independent of $x$, an integration over $x$ leads to Weisman’s (1976) approximation for the depth of the TIBL, namely

$$\zeta = \left( \frac{2H_o x}{\rho c_p \gamma U_m} \right)^{1/2}$$

According to Stunder and Sethuraman (1985), (30) is the overall best simple approximation for the TIBL height.

Applying (18) and (19), the local free-convection prediction can now be reformulated as

$$\sigma^2_w / w^2_{*w} = 1.8(z / \zeta)^{2/3}\left(1 - 1.2z / \zeta\right)^{2/3}$$

$$\sigma^2_{\theta} / \Theta^2_{*\theta} = 1.5(z / \zeta)^{-2/3}\left(1 - 1.2z / \zeta\right)^{4/3}$$

$$\overline{w^3} / w^3_{*w} = 1.2(z / \zeta)(1 - 1.2z / \zeta)$$

These predictions are expected to be valid for the lower half of the TIBL. Applying (22) to (25), the similarity relationships for the whole TIBL become

$$\sigma^2_w / w^2_{*w} = 1.8(z / \zeta)^{2/3}\left(1 - z / \zeta\right)^{2/3} + k_{w3} r^{2/3}\left(z / \zeta\right)^{2/3}\left(1 - z / \zeta + D\right)^{2/3}$$

$$\sigma^2_{\theta} / \Theta^2_{*\theta} = 1.5(z / \zeta)^{-2/3}\left(1 - z / \zeta\right)^{4/3} + k_{\theta3} r^{4/3}\left(1 - z / \zeta + D\right)^{-2/3}\left(z / \zeta\right)^{4/3}$$

$$\overline{w^3} / w^3_{*w} = 1.2(z / \zeta)(1 - z / \zeta) + k_{w3} r(z / \zeta)(1 - z / \zeta + D)$$
where the parameters $k_{w1}$, $k_{w2}$ and $k_{w3}$ remain to be determined. Note that $\xi_1 = 1.2\xi$ and $\xi$ given by (30) depend only on external parameters; hence the purpose of describing the turbulence properties within the TIBL in terms of external parameters is realized. The similarity relationships given in (31) to (36) will predict $\sigma_w^2$, $\sigma_{\theta}^2$ and the vertical transport of $\sigma_w^2$, once the three external parameters, namely, the surface heat flux, the onshore wind velocity and the stability of the approaching airflow, are given.

(b) Similarity relationships within the TIBL

The data used to test the similarity relationships given in (31) to (36) are listed in Table 2. The surface heat flux, $H_o$, is estimated from the linear extrapolation of the heat-flux profile measured along $P_4$ with the additional supposition that in the beach areas $H_o$ is half of the values shown in Table 2. Since accurate measurements for the onshore wind velocity are not available, the TIBL height, $\zeta$, is determined from the field of sensible-heat fluxes, while the factor $(2H_o/(\rho c_p y U_m))^{1/2}$ is determined from (30).

<table>
<thead>
<tr>
<th>Name</th>
<th>$H_o$ (W m$^{-2}$)</th>
<th>$(2H_o/(\rho c_p y U_m))^{1/2}$ (m)</th>
<th>No. of runs</th>
</tr>
</thead>
<tbody>
<tr>
<td>USG 1</td>
<td>325</td>
<td>4.3</td>
<td>12</td>
</tr>
<tr>
<td>USG 2</td>
<td>345</td>
<td>3.5</td>
<td>10</td>
</tr>
<tr>
<td>USG 3</td>
<td>225</td>
<td>3.0</td>
<td>9</td>
</tr>
<tr>
<td>USG 6</td>
<td>365</td>
<td>2.7</td>
<td>11</td>
</tr>
<tr>
<td>USG 8</td>
<td>160</td>
<td>2.0</td>
<td>6</td>
</tr>
<tr>
<td>USG 12</td>
<td>110</td>
<td>1.7</td>
<td>6</td>
</tr>
</tbody>
</table>

Figure 10 shows the profile of sensible-heat flux normalized with $H_o$. Although there is some scatter, $H/H_o$ decreases linearly with $z/\xi_1$. Hence, the supposition (17) used in the simple TIBL model is justified.

The scaled vertical-velocity variance, $\sigma_w^2/w_{*}^2$, is plotted versus the dimensionless height $z/\zeta$ in Fig. 11(a), together with the observations of Smedman and Högström (1983). For comparison, results for CBL over uniform surfaces are shown with $z_i$ replaced by $\zeta$. As can be seen from Fig. 11(a), (31) (thin solid line) agrees well with most data points in the lower half of the TIBL, but the prediction that $\sigma_w^2$ approaches zero at the top of the TIBL ($z = \zeta$ or $z = \xi_1/1.2$) is unrealistic. The best fit to the measured data points can be obtained with (34). Since the exact value of $r$ is unknown, a robust fitting is used to determine $k_{w1} r^{2/3}$, rather than $k_{w1}$. The thick solid curve in Fig. 11(a) shows that (34) can best describe the observations with $k_{w1} r^{2/3} = 0.1$.

The behaviour of the scaled temperature variance, $\sigma_{\theta}^2/\Theta_{*}^2$, is depicted in Fig. 11(b). The prediction given by (32) (thin solid line in Fig. 11(b)) agrees with the observations on most occasions in the lower half of the TIBL, while a considerable departure can be observed at relatively higher levels. The prediction (35) is shown by the thick line in Fig. 11(b) with $k_{m} r^{4/3} = 0.8$, which is again determined using the robust fitting. Figure 11(c) shows that the local free-convection prediction given by (33) agrees well with the observations in the lower part but less so in the upper part of the TIBL. Again, the data can best be described using (36) with $k_{w3} r = 0.04$, which leads to an improved agreement with the observations.

The similarity relationships obtained for the TIBL ((31) to (36)) have expressions very similar to those obtained for homogeneous convective boundary layers, except that the scaling length, $\zeta$, is a function of $x$. This can be interpreted as follows: the TIBL can be regarded as consisting of columns of air, the depth of which increases in the onshore
Figure 10. The profile of the sensible-heat flux, $H$, normalized with its surface value, $H_0$, as a function of $z/\zeta_i$.

Figure 11. (a) $\sigma_u^2/w_{*z}^2$ versus $z/\zeta_i$. Thin line is (31), thick line (34). Circles are the present observations, diamonds those of Smedman and Högström (1983), triangles those of Caughey and Palmer (1979) and Lenschow et al. (1980), and dashed line is after Moeng and Wyngaard (1989).
Figure 11. (b) $\sigma_{\theta}^2/\Theta_{*z}^2$ versus $z/\zeta$. Thin line is (32), thick line (35). Symbols and dashed line as for (a).

Figure 11. (c) $\overline{w'^3}/w_{*z}^3$, versus $z/\zeta$. Thin line is (33), thick line (36). Symbols and dashed line as for (a) but triangles are after Lenschow et al. (1980) only.
direction. The columns can be treated as well mixed individually and the turbulence properties as being in equilibrium with the corresponding surface conditions. The results presented in this section reveal that mixed-layer scaling can be applied to these columns, and that their depth can be used as scaling length. However, these columns are not totally independent of each other and their mutual interaction depends on the advection of heat between them. It is in the scaling length, $\zeta$ or $\xi$, that the inhomogeneity effect is included in the scaling procedure, because the dependence of $\zeta_i$ on $x$ is in principle determined by horizontal advection and vertical turbulent transport of sensible heat.

It is interesting to compare the results obtained in the present study with the similarity relationships obtained in homogeneous conditions. For this purpose, the results presented by Caughey and Palmer (1979) and Lenschow et al. (1980) as well as the large-eddy simulation of Moeng and Wyngaard (1989) are shown in Fig. 11(a, b, c). The comparison indicates that the normalized $\sigma^2_w$ and $w^{+3}$ are smaller, while $a^*$ is bigger in CBLs over uniform surfaces than in the TBL. The results of Smedman and Högström (1983) obtained also in a coastal boundary layer coincide with the observations in the USG experiments. An underestimation of ground heat fluxes in the present study may cause this disagreement, but a satisfactory explanation for this phenomenon will require further investigation.

Because of practical difficulties, the variations of $H_o$ and $\xi$ with time are not included in the turbulence scaling. As a consequence, possible inaccuracies are introduced in calculating the scaling parameters, $w^*$ and $\Theta^*$, by using the values of $H_o$ and $\xi$ estimated for certain time points, when the measurements of the turbulence statistics extended over a period of time. However, this inaccuracy is less serious than it seems. Since only runs made within the TIBL are used in the discussion of turbulence similarity, the time period of the measurements is between about 30 min (for USG12) to 75 min (for USG1, 2); although the complete flight pattern took about 2.5 hours. Both $H_o$ and $\xi$ vary with time, but the variation of $\xi$ is not independent of $H_o$, as can be seen from (30). Supposing that $H_o$ varies by $30 \text{ W m}^{-2}$ per hour at midday, and at twice that rate in the morning, it can be shown that the relative error in $w^*$, $\Theta^*$, and $\xi$ is less than 2% (for the 33 runs of USG1, 2, 6), and less than 10% (for the remaining 21 runs), with an average for all runs of less than 5%. Obviously, this error would not significantly change the conclusions reached in this study.

(c) Structure parameter similarity within the TIBL

The structure parameters for vertical velocity, $C_w$, and for potential temperature, $C_\theta$, are two other important quantities that need to be considered, since they are closely related to $\varepsilon$, the dissipation rate for turbulent kinetic energy, and to $\varepsilon_\theta$, the destruction rate for the potential temperature variance:

\[
C_w = 4\alpha \varepsilon^{2/3} \\
C_\theta = 4\beta \varepsilon^{-1/3} \varepsilon_\theta
\]

where $\alpha$ and $\beta$ are the Kolmogorov and the Corsin constants, respectively. For an arbitrary variable $a$, $C_a$ can be computed from the space series of $a$ i.e.

\[
C_a = \langle (a(r) - a(r + R))^2 \rangle |R|^{-2/3}
\]

where $a(r)$ is the value of a physical parameter at position $r$ and $a(r + R)$ is the value at $(r + R)$ with $R$ being the distance between the two measurements; the operator $\langle \rangle$ denotes the ensemble average. If $C_a$ is independent of $r$ and $|R|$, the turbulence is called homogeneous and isotropic, respectively. It is known that turbulence is effectively
isotropic ('local isotropy') in the inertial subrange and that the structure parameters are constant. Figure 12 shows examples of the structure parameters of vertical velocity and potential temperature. As can be seen, both $C_w$ and $C_\theta$ are approximately constant for $|R| < 80$ m, which confirms the suggestion of local isotropy.

The structure parameters used to compute $\varepsilon$ and $\varepsilon_\theta$ are the averaged values of $C_w$ and $C_\theta$ over the local isotropic range. The dissipation (destruction) rates determined in this fashion are linearly averaged over eight observations and presented in Table 3.

Table 3 shows that from over water to over land $\varepsilon$ increases 10 to 50 times. At the 20 m level, for instance, $\varepsilon(P_1)$ is about $12 \times 10^{-4}$ m$^2$s$^{-3}$ while $\varepsilon(P_4)$ exceeds $112 \times 10^{-4}$ m$^2$s$^{-3}$. On average, $\varepsilon$ decreases faster with height in the stable part than in the convective part of the boundary layer. In some cases, $\varepsilon$ in the convective region is constant up to the 120 m level, in agreement with the results obtained in well-mixed layers (Kaimal et al. 1976). In stable conditions, the rapid decrease of $\varepsilon$ with height also matches the findings regarding homogeneous stable boundary layers, reported, for instance, by Caughey et al. (1979). The destruction rate of temperature variance, $\varepsilon_\theta$, also increases considerably in the onshore direction and decreases with height.

Based on the results of the previous section, it can be expected that the similarity functions $\psi_{C_w}$ and $\psi_{C_\theta}$ defined by

$$
\psi_{C_w} = C_w \varepsilon_{2/3} / w^2 \\
\psi_{C_\theta} = C_\theta \varepsilon_{2/3} / \Theta^2
$$

| Table 3. $\varepsilon$ and $\varepsilon_\theta$ Averaged over Eight Observations |
|-----------------|-----------------|-----------------|-----------------|-----------------|-----------------|-----------------|-----------------|-----------------|
| $z$(m)          | $\varepsilon$    | $\varepsilon_\theta$ |
|                | $P_1$  | $P_2$  | $P_3$  | $P_4$  | $P_1$  | $P_2$  | $P_3$  | $P_4$  |
| 240             | 1.00   | 2.34   | 28.64  | 53.74  | 0.04   | 0.14   | 1.27   | 1.32   |
| 120             | 3.55   | 16.01  | 60.14  | 72.78  | 0.16   | 0.45   | 1.83   | 1.90   |
| 60              | 3.53   | 19.91  | 89.76  | 89.38  | 0.62   | 1.81   | 5.40   | 4.78   |
| 20              | 12.28  | 26.60  | 117.19 | 112.73 | 0.85   | 6.99   | 15.80  | 14.69  |
are functions of $z/\zeta_i$ only. Figure 13(a) shows that $\psi_{c_w}$ varies little in the entire TIBL. This result is in accordance with the observation that $C_w$ is constant in well-mixed boundary layers, obtained for instance by Kaimal et al. (1976). It seems, however, that $\psi_{c_w}$ decreases slightly with height and an appropriate approximation can be given by

$$\psi_{c_w} = 1.3 + 0.1(z/\zeta_i)^{-1}. \quad (37)$$

Figure 13(b) shows that $\psi_{c_\theta}$ decreases with height, which is also in qualitative agreement with the results found in well-mixed layers. The approximation shown in Fig. 13(b), however, only fits the observations in the lower part of the TIBL, while relatively large scatter occurs in the upper levels. Kaimal et al. (1976) reported that

$$C_\theta z_i^{2/3}/\Theta_*^2 = 2.67(z/z_i)^{-4/3}$$

for the lower part of a well-mixed layer. Therefore, quantitative difference again exists between the present results for the TIBL and results obtained for well-mixed layers.

![Figure 13](image-url)

Figure 13. (a) $\psi_{c_w}$ versus $z/\zeta_i$. The curve is given by (37). (b) $\psi_{c_\theta}$ versus $z/\zeta_i$. The curve is given by (38).
7. Conclusions

Using the airborne measurements obtained in the USG experiments, this study investigated the turbulent structure of a coastal atmospheric boundary layer—an extreme case of horizontal inhomogeneity. The spatial evolution of the coastal boundary layer and its development with time were presented. The similarity of turbulence in the convective region of the coastal boundary layer was investigated and the similarity relationships for some important parameters were determined from the observations.

It was shown that under onshore wind conditions, the coastal boundary layer is constructed of a stable outer region and a convective inner region (the TIBL) with intense thermal contrast between them. While sensible-heat fluxes in the outer region were small or negative, those within the TIBL reached 300 to 350 W m⁻². Advective influence caused sensible-heat fluxes to converge rapidly with height and at the upper boundary of the TIBL sensible-heat fluxes were close to zero. Turbulent kinetic energy increased rapidly in a narrow region near the top of the TIBL. This narrow zone, similar to a front, separates the region of intense turbulence from that of weak turbulence. The turbulence intensity within the TIBL was usually 5 to 10 times larger than in the outer region. The observations showed that the region of intense turbulence penetrated to higher levels than the well-mixed layer, defined from the thermal fields, and the 'e-fronts' were in accordance with regions of measurable negative heat flux.

Turbulence in the outer and inner regions of the coastal boundary layer differed not only in intensity. Whereas turbulence in the outer region was sporadic, uncorrelated and embedded in waves on larger scales, within the TIBL it was organized into identifiable thermals. There was no obvious significant difference between the turbulence characteristics in TIBLs and CBLs over uniform surfaces. The PDFs of vertical velocity and potential temperature were both skewed inside the TIBL, but had a predominantly Gaussian distribution in the outer region. It appears that the transformation from stable to convective turbulence was a sudden process, although further detailed studies are necessary.

This study extended the pseudo-homogeneity assumption used by Shao and Hacker (1990) and tested the effectiveness of local free-convection scaling and the 'top-down' and 'bottom-up' scaling for the TIBL. By using a simple model for the TIBL, it was possible to express the local scaling parameters in terms of external parameters. The local similarity relationships (1), (2) and (16) determined in previous studies were converted to (31), (32) and (33), which are effective for the lower half of the TIBL. These relationships depend only on three external parameters: the surface heat flux, the onshore wind speed and the stability of the outer region. Near the top of the TIBL, where sensible-heat fluxes became small, (31), (32) and (33) are not valid. Similar to the case of CBLs over uniform surfaces, this failure of local scaling was attributed to the entrainment at the top of the TIBL. The 'top-down' and 'bottom-up' scaling was used to obtain a best fit with the observations. It was shown that the similarity relationships (34) to (36) described the observations well. Furthermore, the similarity relationships for the structure parameters $C_w$ and $C_d$ were also established. Comparing the similarity functions obtained in this study with those obtained for homogeneous CBLs, some quantitative differences were detected. In general, normalized quantities related to velocity fluctuations were greater than those in homogeneous CBLs, while normalized quantities related to temperature fluctuations were smaller than those observed in homogeneous CBLs, in agreement with the observations of Smedman and Högström; but an explanation for this phenomenon must await further studies.
ACKNOWLEDGEMENTS

We thank Dr Michael R. Raupach and Dr Jörg Hartmann for helpful discussions. Part of the study was carried out under a grant from the Australia Research Council. The aircraft and its instrumentation were funded by donations of the late Dr Don Schultz of Glen Osmond, South Australia.

REFERENCES


Deardorff, J. W. 1970 Convective velocity and temperature scales for the unstable planetary boundary layer and for Rayleigh convection. J. Atmos. Sci., 27, 1211-1213


Hartmann, J. 1990 ‘Airborne turbulence measurements in the maritime convective boundary layer’. Ph.D thesis, Flinders University of South Australia


Shao, Y. 1990 ‘Turbulence and turbulent diffusion in a coastal atmospheric boundary layer’. Ph.D thesis, Flinders University of South Australia


<table>
<thead>
<tr>
<th>Author(s)</th>
<th>Year</th>
<th>Title and Details</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>1986</td>
<td>An examination of methods to estimate the height of the coastal internal boundary layer. <em>Boundary-Layer Meteorol.</em>, 36, 149-156</td>
</tr>
</tbody>
</table>