The marine atmospheric boundary layer during SEMAPHORE. I: Mean vertical structure and non-axisymmetry of turbulence

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Summary
This paper describes the vertical structure of the marine atmospheric boundary layer, analysed using aircraft measurements in the Azores region during SEMAPHORE (Structure des Echanges Mer–Atmosphère, Propriétés des Hétérogénéités Océaniques: Recherche Expérimentale). The flights were performed in the middle of the solar day, during near-anticyclonic conditions, with weak-to-moderate winds and weak surface sensible-heat flux, over a homogeneous oceanic area. The boundary layer was characterized by a mixed layer, driven by surface fluxes, which was not coupled to the overlying, broken stratocumulus, layer. This decoupling is demonstrated by the shape of the continuous profile of the dissipation rate of turbulent kinetic energy, computed from the continuous slant profiles performed by the aircraft. This method makes it possible to determine the thickness of the mixed layer, which is a relevant scale for the parametrization of the profiles. The turbulence field is analysed through the wavelength of the spectrum peak of the vertical velocity, computed on 392 runs more or less parallel or perpendicular to the mean wind. It demonstrates a stretching of the most energetic eddies along the mean wind. The consequences of this non-axisymmetric behaviour of the turbulent field on the flux estimates from aircraft data are discussed.

Keywords: Aircraft observation Boundary layer SEMAPHORE

1. Introduction

Mass and energy, transferred towards the atmosphere at the open ocean surface, are involved in the marine atmospheric boundary layer (MABL). Given the considerable area covered by the oceans on the earth, it is of fundamental importance to adequately monitor the exchanges within and outside the MABL. This knowledge requires data, and mainly turbulence data, in order to parametrize the flux at the ocean surface and in the MABL. Up to now, instrumented aircraft remain the best tools for in situ measurement of the mean and turbulent quantities throughout the whole MABL. Furthermore, the boundary layer over the open ocean is generally much more homogeneous (in horizontal directions) than continental boundary layers, at least for scales up to some hundred kilometres. In this way, the MABL constitutes a convenient laboratory for studying turbulence processes and, therefore, improving their parametrization.

Several major campaigns were devoted to the study of the marine atmosphere in the intertropical zone, because in these areas a large amount of energy is transferred from the ocean towards the atmosphere: among others, GATE (the Atlantic Tropical Experiment of the Global Atmospheric Research Program) and, more recently, TOGA-COARE (the Tropical Ocean/Global atmosphere Coupled Ocean–Atmosphere Response Experiment) were the most important. However, equatorial meteorological conditions (namely, weak winds and strong convection) make it uncertain that the results from such experiments can be applied in the extratropics. In the latter, several airborne campaigns were performed during cold-air outbreaks, during which strong surface fluxes of momentum, heat and evaporation occur. The turbulence structure was analysed and the resulting parametrizations did not differ significantly from those obtained over continental surfaces (see for instance Lenschow et al. (1980) for the results of the Air Mass Transformation Experiment, AMTEX, campaign). More recently, the ASTEX (Atlantic Stratocumulus Transition Experiment; Albrecht et al. 1995) was conducted to analyse the structure of the stratocumulus layer, at the top of the extratropical MABL.

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In June 1992, during the ASTEX field program, the SOFIA campaign (Surface of the Ocean, Flux and Interaction with the Atmosphere; Weill et al. 1995) was performed to improve the parametrization of the air–sea fluxes and of the MABL turbulence. This experiment was conducted in the Azores region, with an instrumented ship and two instrumented aircraft. Anticyclonic conditions prevailed during the whole campaign, leading to weak-to-moderate winds, and to the presence of broken stratocumulus (Sc) at the top of the boundary layer. The turbulence structure was described by Réchou et al. (1995) and Réchou and Durand (1997). SOFIA was, in fact, the first realization of a programme, devoted to the study of the ocean–atmosphere interaction, whose most important phase was the SEMAPHORE (Structure des Echanges Mer–Atmosphère, Propriétés des Hétérogénéités Océaniques: Recherche Expérimentale) field campaign, performed in the autumn of 1993 in the Azores region. During SEMAPHORE, two instrumented aircraft were simultaneously used for in-situ turbulence measurements in the MABL (Lambert and Durand 1998), generally in the middle part of the day. At this time, the boundary layer was often covered by broken stratocumulus, topped by a sharp temperature inversion. The meteorological conditions varied during the campaign, ranging from anticyclonic, low-wind speed regimes to low pressures associated with moderate-to-strong winds. Nevertheless, the surface sensible-heat flux did not exceed 30 W m\(^{-2}\) in most cases, whereas the surface latent-heat flux was one order of magnitude greater (Durand et al. 1998). These conditions can be considered as representative of vast extratropical, oceanic boundary layers, except for specific conditions like atmospheric fronts, very strong winds or strong surface variability (close to the Gulf Stream for example).

We present in this paper a description of the mean and turbulent structure of the atmospheric mixed layer driven by surface flux. In the first section, a general overview of the SEMAPHORE campaign and of the aircraft data is given. The following section describes the vertical structure of the mixed layer through the averaged profiles of the mean thermodynamic parameters. In particular, we present the method used to discriminate between the coupled and decoupled Sc layers, and to determine the mixed-layer thickness. We then analyse the turbulent field and, more precisely, the non-axisymmetry of the energetic scales and its consequences on the estimations of the various turbulence moments.

In a companion paper (Lambert et al. 1999), we analyse the turbulence structure of the MABL through the normalized profiles of the principal moments.

2. The data

A general description of the strategy and equipment used during the SEMAPHORE campaign is given by Eymard et al. (1996). The two aircraft used were the Fokker 27 instrumented by the French Institut National des Sciences de l’Univers, and the Merlin IV from ‘Météo-France’. The turbulence measurements were acquired at a rate of 64 s\(^{-1}\) (Fokker 27) or 50 s\(^{-1}\) (Merlin IV) after anti-aliasing filtering. The data processing enabled calculation of the three wind components, the temperature and the moisture at a rate of 16 s\(^{-1}\) for the Fokker 27 and 25 s\(^{-1}\) for the Merlin. Details concerning the instrumentation and data processing can be found in Lambert and Durand (1998) and in Durand et al. (1998). A thorough intercomparison between the two aircraft, including turbulence moments, was performed by Lambert and Durand (1998) and demonstrated that the two platforms could be used simultaneously.

The experimental area was situated to the south of the Azores archipelago. The aircraft were based on the island of Santa-Maria (37°N, 25°W), which is the southernmost island. As described by Eymard et al. (1996), the Azores current was oriented NW–SE in the
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SEMAPHORE domain, and lies around 34°N. Across this current, SST gradients of about 1 to 2 deg C 100 km\(^{-1}\) were observed. The aircraft performed two kinds of missions during the experiment: the first was devoted to the description of the boundary layer structure above the area of the Azores current (Kwon et al. 1998); and the second to the analysis of the mean and turbulence structure of a homogeneous area, situated north of the current. The data used in this paper relate to these latter flights, mainly performed during anticyclonic conditions. In consequence, we will assume horizontal homogeneity (at a scale of several tens of kilometres) and will only analyse the vertical structure of the MABL.

3. VERTICAL STRUCTURE OF THE MABL

(a) Mixed-layer thickness

As shown by Réchou et al. (1995), the MABL, topped by broken Sc, often presents in its lower part a mixed layer driven by surface fluxes and whose thickness, \(h\), is much smaller than the altitude of the inversion (\(Z_i\)) which caps the Sc layer. There is therefore a decoupling between the surface and the Sc layer. They demonstrated that \(h\) is a relevant scale for mixed layer parametrization, that the normalized altitude is \(z/\tilde{h}\) (where \(z\) is the altitude) instead of \(z/Z_i\), and that the convective-scale (\(w_c\); Deardorff 1970) must include \(h\) rather than \(Z_i\). The problem is: first, to determine whether the situation examined is coupled or decoupled; and, in the latter case, second, to accurately determine \(h\), because the top of the mixed layer is not associated with a strong vertical gradient of potential temperature and/or moisture. This is the reason why we have developed a method based on determination of the profile of the dissipation rate \(\varepsilon\) of turbulent kinetic energy. In the inertial subrange, \(\varepsilon\) is related to the wave number \(\kappa\), and to the spectral energy \(S_i\) of the velocity component \(U_i\) by the well-known Kolmogorov formula:

\[
S_i(\kappa) = \alpha_i \varepsilon^{2/3} \kappa^{-5/3} \tag{1}
\]

where \(\alpha_i\) is the Kolmogorov constant, for which we used a value of 0.52 in the sampling direction and 0.52 times 4/3 in the transverse directions. These values lie within the range of commonly used values (Kaimal and Finnigan 1994). The most widely used technique to calculate \(\varepsilon\) from aircraft measurements is the following: on a straight and level run (generally of several tens of kilometres), one computes the time series of a turbulent velocity component \(U_i\), and the corresponding power spectrum \(S_i\). Fitting \(S_i\) with a \(-5/3\) power law in a wave number range included in the inertial subrange, therefore provides the value of \(\varepsilon\) using (1). Another technique, already used by Shaw and Businger (1985), consists in high-pass (or band-pass) filtering the time series of the velocity component in order to keep only signal frequencies involved in the inertial subrange. Let us call \(\kappa_1\) and \(\kappa_2\) (\(\kappa_2 > \kappa_1\)) the two wave numbers corresponding to the cut-off of the filter (\(\kappa_1\) is the high-pass filter cut-off frequency, \(\kappa_2\) is the Nyquist wave number). Integrating (1) between \(\kappa_1\) and \(\kappa_2\), and using the Parseval’s relation, we obtain:

\[
\varepsilon = \sigma_{U_i}^2 \times (3/2) \alpha_i (\kappa_1^{-2/3} - \kappa_2^{-2/3})^{-3/2} \tag{2}
\]

where \(\sigma_{U_i}^2\) is the variance of the filtered velocity \(U_i\). This technique was applied, not on straight and level runs (as is generally done), but on the time series of the turbulent vertical velocity of the air, computed from the continuous slant profiles performed by the aircraft (Druilhet et al. 1985). An example of a slant descent (from 2600 m to 100 m altitude) is given in Fig. 1: the vertical velocity was filtered in the wavelength range 6 m to 100 m. \(\varepsilon\) can be computed from this signal, on overlapping segments short enough to obtain a quasi-continuous profile. In the example presented in Fig. 1, \(\varepsilon\) was computed on overlapping
segments corresponding to altitude bins of about 150 m, the vertical distance between two consecutive values being about 6 m. The greatest values of $\varepsilon$ are obtained when the aircraft leaves the upper stable layer and penetrates into the cloud layer. A second sequence of high values is apparent in the last third of the sample, i.e. in the mixed layer.

The quasi-continuous profiles of $\varepsilon$ were then used to characterize the vertical structure of the MABL: Fig. 2 presents the results obtained from two aircraft soundings, performed over the same area, at around 1600 UTC and 1800 UTC. The boundary layer is characterized by a broken Sc layer, identified on the profiles of liquid water content, extending from 1400 to 1800 m, and topped by a sharp temperature inversion accompanied by considerable variation in moisture. The thermodynamic profiles appear very similar for the two soundings. However, the profiles of $\varepsilon$ present very different behaviour: on the 1600 UTC profile, there is a layer of low values of $\varepsilon$, between 900 and 1350 m, which therefore corresponds to a zone of low turbulent kinetic energy, and translates as a decoupling between the cloud layer and the mixed layer driven by surface fluxes. Two hours later, the profile of $\varepsilon$, computed over the same area, indicates continuous turbulence from the surface up to the top of the Sc layer. The difference between these two profiles result from the diurnal cycle of the Sc layer. In the middle of the day, the absorption of short-wave radiation at the top of the cloud counteracts the long-wave radiative cooling. The downward buoyancy source at the top of the cloud therefore decreases, decoupling appears and the cloud cover tends
to decrease. But, in the late afternoon, short-wave radiation vanishes and the long-wave radiative cooling at the Sc top acts as a source for downward buoyancy, which extends the in-cloud turbulence towards lower altitudes. The consequence is the coupling between surface and cloud layer, and a deepening of the cloud layer. These processes are relatively well known. They were described by numerical models (see for instance Bougeault 1985) and by observation (Betts 1990; Betts et al. 1995).

The vertical structure of the MABL, and particularly the coupling or decoupling behaviour, is of fundamental importance for the definition of the characteristic length-scales used for profile parametrization: in the 1600 UTC case of Fig. 2, the mixed layer, driven by surface fluxes, extends up to $h = 900$ m. This value must be used as a characteristic length-scale. In the 1800 UTC case, the mixed layer extends from the surface up to $Z_i$, which attains 1800 m. In the latter case, there is no difference between the mixed layer and the MABL, and $Z_i$ will be used as a characteristic length-scale. As can be seen on Fig. 2, we could not discriminate between the two situations without the continuous profile of $\varepsilon$. Moreover, the values of $\varepsilon$, computed on the straight and level runs between 1600
and 1800 UTC over the same area, are indicated on the two profiles (solid dots); given the altitudes of these runs, these values cannot indicate the coupling or decoupling behaviour of the MABL.

The continuous profiles of ε were therefore used to analyse systematically the vertical structure of the boundary layers measured by the aircraft in the homogeneous area during the SEMAPHORE experiment. Given the time of the flights (around the middle of the day), most of the boundary layers were decoupled. Table 1 presents the mean parameters of the situations analysed in this paper. The meteorological conditions were generally anticyclonic, with surface pressure ranging from 1015 to 1029 hPa. The mean wind in the boundary layer was low to moderate, blowing from 018° to 198°, in relation to the southern position of the experimental area in the anticyclonic region. The mixed-layer thickness, h, is much lower than the altitude of the trade inversion Z_i. The average SST over the period was 20.4 °C. Between October 15th and November 16th, it decreased by a little more than 1 deg C, mainly just after the storm which crossed the area on October 29th (Caniaux and Planton 1998; Giordani et al. 1998). The potential-temperature variations shown in Table 1 are mainly related to the variations of surface atmospheric pressure rather than to those of surface temperature. By contrast, specific humidity presented considerable variation, ranging from 5.8 g kg\(^{-1}\) on October 16th to 9.0 g kg\(^{-1}\) on November 16th, which, at low altitudes (say, at the top of the surface layer) corresponds to relative humidities of 45% and 69%, respectively. The lifting condensation level (LCL), computed from the average potential temperature and specific humidity in the mixed layer, presented values generally considerably greater than the altitude of the mixed layer top, h. Only five experiments presented values of (LCL - h) lower than, or of the order of one hundred metres. These situations were favourable to the development of some scattered Cumulus (Cu) below the Sc layer.

(b) Mean profiles

In order to define profiles averaged over the whole campaign, the normalized altitude \(\hat{Z}\) was defined in the following way:

\[
\hat{Z} = \frac{z}{h} \quad \text{for } z \leq h \\
\hat{Z} = 1 + \frac{(z - h)}{(Z_i - h)} \quad \text{for } z > h.
\]  

(3)

The averaged profiles of potential temperature and specific humidity are presented in Fig. 3. For each experiment, the mean value in the mixed layer (indicated in Table 1) was subtracted from the profile. The potential temperature is well mixed from the surface up to the top of the mixed layer (\(\hat{Z} = 1\)). Between \(h\) and \(Z_i\), the profile is stably stratified. The transition between the two profiles, at \(z = h\), which cannot easily be seen on the individual profiles because of the fluctuations in potential temperature and moisture, is clearly demonstrated here, and validates the notion of decoupling. In the temperature inversion layer, at \(z = Z_i\) (or \(\hat{Z} = 2\)) the average jump in temperature reaches 4 deg C. The profile of the humidity mixing ratio decreases slightly with altitude in the mixed layer. The important drying which accompanies the temperature inversion at \(z = Z_i\) (\(\hat{Z} = 2\)) attains on average -3 g kg\(^{-1}\). Two average profiles of the liquid water content, also presented in Fig. 3, were computed from the 1 s averaged data. Given the airspeed of the aircraft, 1 s of flight represents an 80 to 100 m path. The first profile represents the average of the paths within which liquid water was detected, and was therefore computed after removing cloud-free paths. We will call it the cloud-average profile (CAP). The second represents the average of the whole data set, including cloud-free and cloudy paths. We will call it the ensemble-average profile (EAP). The CAP presents a maximum of 0.23 g m\(^{-3}\) at \(Z_i\).
<table>
<thead>
<tr>
<th>Day</th>
<th>Flight number</th>
<th>Time (h:min UTC)</th>
<th>Surface pressure (hPa)</th>
<th>Potential temperature (°C)</th>
<th>Specific humidity (g kg⁻¹)</th>
<th>Wind speed (m s⁻¹)</th>
<th>Wind direction (Degrees)</th>
<th>LCL (m)</th>
<th>SST (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>15 Oct.</td>
<td>31</td>
<td>15:50 to 17:36</td>
<td>912 1735</td>
<td>1015</td>
<td>17.35</td>
<td>7.75</td>
<td>9.75</td>
<td>23</td>
<td>1040</td>
</tr>
<tr>
<td>16 Oct.</td>
<td>32</td>
<td>12:00 to 13:53</td>
<td>1261 1887</td>
<td>1020</td>
<td>16.85</td>
<td>5.8</td>
<td>6</td>
<td>41</td>
<td>1540</td>
</tr>
<tr>
<td>17 Oct.</td>
<td>33</td>
<td>11:11 to 13:06</td>
<td>1295 2006</td>
<td>1024</td>
<td>17.5</td>
<td>7.1</td>
<td>9.2</td>
<td>117</td>
<td>1290</td>
</tr>
<tr>
<td>17 Oct.</td>
<td>34</td>
<td>15:59 to 17:40</td>
<td>1017 1848</td>
<td>1023</td>
<td>17.8</td>
<td>7.33</td>
<td>10.2</td>
<td>128</td>
<td>1260</td>
</tr>
<tr>
<td>25 Oct.</td>
<td>40</td>
<td>14:56 to 17:31</td>
<td>697 2135</td>
<td>1017</td>
<td>18.13</td>
<td>8.53</td>
<td>7.3</td>
<td>100</td>
<td>980</td>
</tr>
<tr>
<td>26 Oct.</td>
<td>41</td>
<td>14:44 to 17:40</td>
<td>931 2216</td>
<td>1018</td>
<td>17.85</td>
<td>8.1</td>
<td>7.1</td>
<td>87</td>
<td>1050</td>
</tr>
<tr>
<td>5 Nov.</td>
<td>49</td>
<td>15:54 to 18:26</td>
<td>861 1766</td>
<td>1028</td>
<td>15.2</td>
<td>6.1</td>
<td>7.75</td>
<td>28</td>
<td>1310</td>
</tr>
<tr>
<td>6 Nov.</td>
<td>50</td>
<td>10:33 to 12:39</td>
<td>670 1775</td>
<td>1029</td>
<td>14.35</td>
<td>7.75</td>
<td>9.5</td>
<td>32</td>
<td>780</td>
</tr>
<tr>
<td>6 Nov.</td>
<td>50</td>
<td>16:12 to 18:13</td>
<td>965 1633</td>
<td>1027</td>
<td>15.4</td>
<td>6.73</td>
<td>8</td>
<td>38</td>
<td>1150</td>
</tr>
<tr>
<td>9 Nov.</td>
<td>52</td>
<td>14:56 to 18:03</td>
<td>940 1286</td>
<td>1023</td>
<td>14.8</td>
<td>6.93</td>
<td>5.9</td>
<td>18</td>
<td>990</td>
</tr>
<tr>
<td>15 Nov.</td>
<td>57</td>
<td>15:07 to 17:21</td>
<td>500 1140</td>
<td>1026</td>
<td>16.35</td>
<td>8.73</td>
<td>1</td>
<td>110</td>
<td>790</td>
</tr>
<tr>
<td>16 Nov.</td>
<td>58</td>
<td>13:34 to 16:29</td>
<td>491 1179</td>
<td>1024</td>
<td>16.8</td>
<td>9</td>
<td>6.2</td>
<td>198</td>
<td>770</td>
</tr>
</tbody>
</table>

\( h \) represents the thickness of the mixed layer, and \( Z_i \) the altitude of the trade inversion. The potential temperature, specific humidity and wind values represent averages over the whole mixed layer. LCL is the mixed-layer lifting condensation level and SST the sea surface temperature in the experimental area.
Figure 3. Average profiles, from top to bottom, of: potential temperature, $\theta$; specific humidity, $q$; (for these two parameters, the mean value in the whole mixed layer was subtracted from the profile for each experiment); liquid water content of the cloud-average profile (CAP); and liquid water content of the ensemble-average profile (EAP). The ordinate is the normalized height; horizontal bars give root mean square values. See text for precise definitions and further explanation.
The profile decreases above $Z_i$, but not as fast as expected for a homogeneous Sc layer. Two reasons can be invoked to explain this behaviour: first, $Z_i$ is an average value for an experiment whereas the top of the Sc could present horizontal variability, causing some cloudy regions to be higher than the average level of the top; and second, some Cu could penetrate into the Sc layer and locally overreach its top (Martin et al. 1995). Below $\bar{z} = 1.5$, a second weak maximum of 0.06 g m$^{-3}$ appears, which is probably related to the presence of scattered Cu fed by the underlying mixed layer. The EAP presents values much lower than those for the CAP, which result from the fractional cloudiness. This parameter, defined as the fractional area covered by the cloud at a specific level (Bechtold and Cuijpers 1995) can be estimated from the ratio of the two profiles (i.e. EAP/CAP) provided that we can assume that the aircraft randomly sampled both cloud and cloud-free air parcels. At $Z_i$, this ratio reaches about 0.06/0.23 = 0.26, whereas in the underlying Cu layer it drops to 0.006/0.10 = 0.06. These values can be considered as representative of the average cloudiness for the whole campaign, in spite of the variability encountered from one day to another.

The horizontal wind components presented in Fig. 4 are computed in a coordinate system linked to the mean wind in the mixed layer: $U$ and $V$ are the longitudinal and transverse components in this frame, respectively; therefore $V$, when averaged through the whole boundary layer equals zero. The profiles in Fig. 4 show that momentum is well mixed in the mixed layer. In the intermediate layer between $h$ and $Z_i$, $V$ remains weak on average, which suggests that the wind does not rotate in any preferred direction. The scatter on the two components is greater in the intermediate layer than in the mixed layer. $U$ significantly decreases (by 1.5 to 3 m s$^{-1}$ on average) in the intermediate layer and above. In the same area, the wind speed $||\bar{U}||$ also decreases but by a lesser amount (about 1 to 2 m s$^{-1}$ in the intermediate layer). Given that the transverse component is weak on average, this suggests that the wind vector turns in this area either clockwise or counter-clockwise with about equal probability. Similar decreases in wind speed above the marine atmospheric mixed layer have already been reported by, for example: LeMone and Pennel (1976) in the Puerto Rico cumulus field; Grossman and Betts (1990) in the cold air outbreak of GALE (Genesis of Atlantic Low Experiments); Grossman and Friehe (1986) during the Summer Monsoon Experiment; and Khalsa and Greenhut (1988) during FASINEX (Frontal Air-Sea Interaction Experiment). Several reasons have been invoked to explain this feature, for example, a low-level jet linked to a mesoscale meteorological situation or mesoscale circulation induced by the SST gradient. This problem has been exhaustively studied by Giordani (1997), Giordani et al. (1998) and Kwon et al. (1998) for the SEMAPHORE situations from both aircraft data and numerical simulations. These authors argue that such circulations are more likely to appear in anticyclonic conditions, which exclude atmospheric large-scale forcings as a possible cause. They conclude that the wind reinforcement in the mixed layer is provoked by mesoscale ageostrophic circulations resulting from the surface forcings (mainly horizontal gradients of sensible-heat flux and momentum flux).

4. **Non-axisymmetry of the turbulence field**

(a) **Characteristic length-scale**

In the case of incompressible, turbulent flow, in homogeneous and stationary conditions, the turbulence would be axisymmetric, i.e. the characteristic-scale (for example, the integral-scale) would be identical for each velocity component. However, the integral-scale, computed from a wind component measured in a given direction, varies according to
Figure 4. As Fig. 3 but average wind profiles: upper frame, wind speed, $\|\bar{U}\|$, from which the mean wind in the whole mixed layer, $\|\bar{U}\|$, was subtracted; middle and lower frames, horizontal and vertical wind components, $U - \bar{U}$ and $V$, respectively, in a coordinate system aligned with the mean wind in the mixed layer. See text for further details.

The orientation of this component with respect to the sample direction (Lumley and Panofsky 1964): if $\lambda_x^u$ is the integral-scale of the $u$ component, sampled along the $x$ direction, and $\lambda_x^v$ the integral scale of the $v$ component, sampled along the $x$ direction ($x$ being parallel to $u$, and $v$ perpendicular to $u$), then $\lambda_x^v = 2\lambda_x^u$; if $y$ is parallel to $v$, then $\lambda_y^v = \lambda_y^u = 2\lambda_y^u = 2\lambda_y^u$. Furthermore, the three-dimensional spectrum of the turbulent velocity field vanishes at low wave numbers, whereas the one-dimensional spectrum of a component remains at a non-zero, constant value (Kaimal and Finnigan 1994). In this model, the mean wind orientation does not intervene. However, several authors have demonstrated that the turbulence field
in the atmospheric boundary layer is far from being axisymmetric: Nicholls and Readings (1981) showed that the energetic eddies are elongated along the mean wind direction. Consequently, the cospectra of heat and momentum flux, when sampled parallel to the wind, peak at wavelengths greater than cospectra sampled across the wind. Lenschow and Stankov (1986) also stated that the integral-scales of the wind components are greater for along-wind than for crosswind samples. These two papers refer to data collected over the sea.

In order to check this behaviour in the mixed layer during SEMAPHORE, we have divided the set of data into two classes: the first one contains the straight and level runs performed with the aircraft track differing by less than 30 degrees from the mean wind direction; the second contains the runs for which the angle between the aircraft track and the wind direction is 90 ± 30 degrees. Hereafter, the first class will be called ‘A’ (‘along-wind’) and the second ‘C’ (‘crosswind’). All the runs involved in this study are about 25 to 30 km long. We have chosen to use the wavelength of the spectrum peak of the vertical velocity \( L_m \) as the characteristic length-scale. If the turbulence had been axisymmetric, it would have been independent of the sampling direction. The vertical velocity is perpendicular to the sampling direction (if we assume that the aircraft flies horizontally), as well as to the horizontal wind.

\( L_m \) was determined from the energy spectrum of the vertical velocity, computed on each run. The following relation provides a good fit to the data:

\[
nS_w(n) = \frac{nS_0}{1 + 1.5(n/n_m)^{5/3}}
\]

where \( n \) is the frequency in the aircraft time reference, \( S_w \) is the spectral energy, \( n_m \) is the frequency corresponding to the maximum energy and \( S_0 \) is the asymptotic value of the energy at \( n = 0 \). For high frequencies, \( S_w(n) \) follows a \(-5/3\) power law, in accordance with the Kolmogorov formulation in the inertial subrange. \( L_m \) is thus deduced from \( n_m \) using the Taylor hypothesis:

\[
L_m = \frac{\overline{T \Delta S}}{n_m}
\]

where \( \overline{T \Delta S} \) is the true airspeed of the aircraft on the run. \( n_m \) and \( S_0 \) were determined in order: first, to ensure that the variance of the time series was identical to the frequency-integral of the energy spectrum (Parseval’s relation); and second, to minimize the quadratic distance between (4) and the spectrum computed from the time series. An example of the computation of \( n_m \) and \( L_m \) on a run of each class is presented in Fig. 5. These two runs were performed consecutively, over the same area and at the same altitude (90 m). The first one was oriented at 84° with respect to the mean wind (C-class), and the second at 6° (A-class). The \( L_m \) values found are 395 m and 743 m, for the C and A samples, respectively. This example illustrates the ability of the model given by (4) to determine \( L_m \). However, given the fluctuations of the individual spectra, as can be seen on this example, we must work on a high number of samples in order to validate our conclusions.

It is now well known that, in the lower part of the atmospheric boundary layer, \( L_m \) is proportional to the altitude (Kaimal et al. 1976; Caughey and Palmer 1979; Druilhet and Durand 1997). In the middle and upper part of the boundary layer, \( L_m \) is close to the mixed layer thickness, \( h \). The profile of \( L_m \) computed for the SEMAPHORE data is presented in Fig. 6, for the two classes defined above. We have restricted the profile in the layer 0 to 300 m, in order to avoid a normalization with \( h \) which would have complicated the interpretation. Given the mixed-layer thickness encountered during SEMAPHORE (878 m on average; see Table 1), the 0 to 300 m layer represents approximately the lower third of the mixed layer. So, in this layer, we can consider that \( L_m \) increases more or less linearly
with altitude. The profiles of Fig. 6 were obtained from 198 runs from class A and 194 from class C. The result is unambiguous: the length-scales are considerably greater on the runs sampled along the wind than on crosswind samples. Knowing that the whole data set was included in these profiles, this translates as a frequent or systematic organization of the turbulence field with energetic eddies elongated in the mean wind direction.

As regards the smaller scales (inertial subrange), Lambert et al. (1999) show, in a companion paper, that the dissipation rate ε of the turbulent kinetic energy, computed from the energy spectrum of the vertical velocity using the Kolmogorov relation in the
(b) Consequences of the non-axisymmetry on the estimates of the turbulence moments

If we do not take into account instrument error (including sensor accuracy and calibration procedures), there are two sources of errors in computing turbulence moments, as analysed by Lenschow et al. (1994): a ‘random’ error translates the scatter of the estimations of a moment, and is mainly related to the ratio of the integral-scale of the process to the sample length; and a ‘systematic error’ which results from the wavelength cut-off of the high-pass filtering applied to the time series before computing the moment. To choose the length of the sample and the wavelength of the cut-off therefore, consists in making a compromise between these two errors. For the SEMAPHORE data, the cut-off wavelength of 5 km was demonstrated to be a good compromise (Lambert and Durand 1998). However, the non-axisymmetry of the turbulence-scale in the horizontal direction results in the spectra and cospectra shifting towards lower frequencies when sampled along wind. The consequences of the high-pass filtering, and the resulting systematic error, would therefore differ for the two classes of samples.

These consequences are illustrated in Figs. 7 and 8 which present, for the vertical velocity variance and latent-heat flux, respectively, profiles of the ratio of the non-filtered values to the values computed after high-pass filtering of the time series with a cut-off wavelength of 5 km. We consider the 0 to 300 m layer, for the reasons explained above. These figures illustrate the behaviour of the systematic error for the two classes of runs. For the C-class runs, filtering reduces both $w$ variance and latent-heat flux by about 10%. The ratio increases with altitude, which is explained by the corresponding increase in the characteristic length-scale, whereas the cut-off wavelength remains constant. For the A-class runs the reduction in both parameters due to filtering is considerably greater; it ranges

![Graph](image)

Figure 7. As Fig. 6, but for the ratio of the non-filtered to the filtered variance of the vertical velocity.
from 10 to 20% for the vertical velocity variance, and from 15 to 30% for the latent-heat flux. Two main conclusions can be drawn from these figures: first, there is a systematic error due to high-pass filtering, even for crosswind runs, which implies that wavelengths greater than 5 km contribute to the transfers; second, this error becomes considerable for along-wind runs. Such an error reflects the fact that not all the frequencies contributing to the transfer are captured in the sample. This can be illustrated by the technique called ‘ogive cospectra’, which consists of a representation of the cumulative cospectrum, starting from the highest frequencies (Desjardins et al. 1989). The ogive curve presents an asymptotic shape on the lower-frequencies side provided that the scales contributing to the transfers have been captured. If not, the ogive curve continues to increase, or decrease according to the sign of the flux. The advantage of this technique lies in the smoothing operation performed by the integration with respect to the cospectra, which are generally noisy at low frequencies and therefore difficult to analyse. Various techniques of flux computation, including the ogive method, were used by Friche et al. (1991) on the FASINEX data. In the companion paper (Lambert et al. 1999) we will give some illustrations of this method.

Mann and Lenschow (1994) proposed the following parametrization of the systematic error on the estimation of a flux, $\varphi$, in the convective boundary layer:

$$\delta \varphi = (\varphi - \varphi_0)/\varphi = bh(z/h)^{1/2}(L_c^{-1} - L^{-1})$$

(6)

where $\varphi_0$ is the flux value computed from high-pass filtered signals at the wavelength cut-off $L_c$, $L$ is the sample length and $b$ is a constant determined from data. Mann and Lenschow (1994) proposed $b = 1.2$. We can see from (6) that $\delta \varphi$ is always positive, which signifies that filtering removes a certain fraction of the flux. We computed $\delta \varphi$ from (6) with $h = 870$ m (an average value deduced for the campaign; see Table 1), $L_c = 5$ km and $L = 30$ km. The resulting values of $1/(1 + \delta \varphi)$ are plotted in Fig. 8: they compare remarkably well to the values computed from the C-class runs.

Figure 9 presents the profile of the ratio of the non-filtered moisture variance to the filtered moisture variance. Several differences appear with respect to the preceding
Figure 9. As Fig. 6, but for the ratio of the non-filtered to the filtered moisture variance.

parameters: the values of this ratio are of the order of 1.4, even for the lowest altitude runs; furthermore, they do not depend on the sample orientation with respect to the wind. This is related to the behaviour of the spectra of the scalars; unlike the vertical velocity, they are not bounded at the low-frequency side. This behaviour is also valid for the horizontal wind components and temperature. In fact, it is only vertical velocity which follows the 'classical' turbulence model, with a spectral gap between turbulence and mesoscale eddies. In the other parameters, considerable variance is induced by these mesoscale eddies. This mesoscale contribution masks the consequences of the along-wind stretching of the eddies.

This study enables us to determine the best approach for computing turbulence moments from aircraft measurements. The best estimations are obtained by averaging a large number of estimations computed from non-filtered time series. The averaging operation reduces the random error, whereas the systematic error is reduced by not using high-pass filtering. On the other hand, if we want to compute turbulence moments from a single run, we need to filter the signals, and we have to use a crosswind data sample.

5. CONCLUSION

The vertical structure of the MABL was analysed from aircraft data gathered during the SEMAPHORE experiment, conducted in the Azores region in autumn 1993. The data used in this paper were collected over a homogeneous area during anticyclonic conditions. The MABL was characterized by a mixed layer, driven by surface fluxes, which did not extend up to the broken Sc layer. The MABL can therefore be divided into three layers: the mixed layer, an intermediate layer and the cloud layer, the latter being topped by a sharp temperature inversion accompanied by significant drying. The mixed layer is thus decoupled from the cloud layer. The mixed-layer thickness, $h$, is an appropriate length-scale for parametrization, but it is often difficult to determine from the data, because of the lack of strong variations in temperature and moisture. We have developed a method, based on the determination of the continuous profile of the dissipation rate of turbulent kinetic energy. This profile, computed from the slant soundings performed by the aircraft, makes
it possible to determine if the mixed layer is decoupled from the cloud layer above and, if this is the case, to determine the value of $h$. The mean parameters (potential temperature, moisture and horizontal wind) were thus brought together into single profiles using $z/h$ as a normalized altitude. These average profiles clearly bring out the different structures of the mixed layer and of the intermediate layer, even though this difference is not evident in the individual profiles.

The non-axisymmetry of the turbulence field was demonstrated through the behaviour of the characteristic length-scale deduced from the spectrum peak of the vertical velocity. This parameter, when computed from samples performed along the mean horizontal wind, presents values much greater than those computed from crosswind samples. This difference is evident in the whole set of data and reveals, if not a systematic, at least a frequent organization of the energetic eddies which were elongated along the mean wind direction.

The consequence of this non-axisymmetry is that spectra and cross-spectra appear shifted towards lower frequencies when sampled along wind. If the turbulent time series are high-pass filtered, the turbulent fluxes when computed from along-wind samples show a systematic error, which attains 15 to 30% for a wavelength cut-off of 5 km. For crosswind samples, this error is reduced to about 10% and corresponds accurately with the formulation of Mann and Lenschow (1994). If the time series are not high-pass filtered, there is no significant difference in the turbulence moments computed from the two kinds of samples.

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