Experimental Study of the Vertical Structure of the Lower Troposphere over a Small Greek Island in the Aegean Sea

C. G. Helmis, C. Jacobides, D. N. Asimakopoulos, and H. A. Flocas

Division of Applied Physics, Department of Physics, University of Athens, Greece

(Manuscript received 20 June 2001, in final form 27 November 2001)

ABSTRACT

An experimental campaign was carried out on a small Greek island that is characterized by complex terrain; its aim was to study the local characteristics of the vertical structure of the atmospheric boundary layer (ABL). The instrumentation was installed close to the shoreline and consisted of a 13-m-high meteorological mast instrumented at three levels, and a high-range vertical monostatic sodar. Tethered balloon flights were carried out for 3 days under different atmospheric conditions. The analysis of the available data revealed interesting features of the vertical structure of the atmosphere over the island, with the development of a convective internal boundary layer (IBL) within the first 150 m above the ground, while the marine boundary layer (MBL) formed at higher altitudes, up to 450 m. Buoyant oscillations appear within the MBL in the form of gravity waves with frequencies of 7 min. Theoretical calculations of the IBL height verified the experimental results. During the night, a complex wind flow forms in the lower 250–300 m, resulting from the development of katabatic flows and topographic channeling.

1. Introduction

Scopelos is a small island in the northwestern Aegean Sea, which is situated in the eastern Mediterranean Basin between Greece and Turkey, extending a few hundred kilometers. An experimental campaign was carried out in summer 1996 near the shoreline at the northeastern part of Scopelos; the goal was the study of the airflow regime at a location where the new local airport was proposed to be constructed. During summer, a weak synoptic wind prevails over the Aegean Sea under the frequent presence of anticyclonic circulation, allowing the generation of local flows, such as sea and land breezes and katabatic winds. Furthermore, the development of an internal boundary layer (IBL) is evident over the examined area. Close to the shoreline an IBL is mechanically induced, which is in response to an abrupt change in roughness between the sea (upwind) and the land (downwind). In addition, a thermal (convective) IBL develops during the daytime as the air moves from the cool sea toward the much warmer land, producing changes in the surface heat fluxes. Thus, the developed flow over the coastal region combines features of the local marine boundary layer (MBL) and the thermally induced local flows.

Observations of the marine atmosphere structure, performed by Mandics and Owens (1975) and Gaynor and Mandics (1978), revealed the existence of a rich variety in the vertical structure associated with temperature inversions that were often perturbed by gravity waves or wind shear. The characteristics of the IBL were extensively studied on a theoretical and experimental basis by Stunder and Sethuraman (1985), Venkatram (1986), Hsu (1986), Bergstrom et al. (1988), and Garratt (1990). According to these studies, the characteristics of the IBL are greatly determined by the roughness of the shoreline, the maritime air stability, the temperature difference between the sea and the land, and the prevailing background flow. The depth of the convective IBL increases with distance downwind of the shoreline, while the influence of the sea is expected to minimize at a sufficiently large distance from the shoreline, where the IBL has achieved an equilibrium with the ABL.

Based on the above, the atmospheric environment in Mediterranean coastal regions is influenced by local circulation flows due to sea–land contrast, which is more pronounced in the islands. Although extensive research concerning the IBL is published, only a few studies examine the IBL over islands. During nighttime the winds over the island are low, and the development of strong katabatic flows combined with land-breeze flow...
are substantially evident. During the daytime, different sea-breeze systems, which develop over the island, interact and produce convergence zones accompanied by upward motion (Melas et al. 2000). Also, the strong influence on the marine atmospheric boundary layer (ABL) structure of isolated islands was demonstrated by Petenko et al. (1996). According to their study, this influence reveals the occurrence of inversion layers at a distance of several tens of kilometers from the shore, which modifies the stratification prevailing in the MBL. Thus, experimental or theoretical studies focusing on the modification of flow in coastal zones as a result of stability and roughness changes are still of major importance. This subject becomes more interesting in a small-scale sea, such as the Aegean, which is characterized by numerous scattered small islands and is surrounded by the Greek (in the north and west) and Turkish mainland (in the east).

In this respect, the objective of the present paper is to study the development of the IBL forming over the coast of a small Aegean island and the marine influence. Furthermore, the vertical structure of the lower troposphere is investigated during the daytime and nighttime, under weak synoptic conditions, allowing the development of local flows, or when onshore moderate synoptic winds blow from the northern sector of the open sea. Finally, an effort was made to test basic empirical relationships for the estimation of IBL height, taking into account the roughness and scalar flux changes.

2. Experimental site and instrumentation

Scopelos is positioned at latitude 39°N and longitude 24°E, with a total area of 96 km², and is characterized by complex topography (Fig. 1). The instrumentation was installed at a location in the northeastern part of the island, at 40 m ASL, close to the shoreline (150 m downwind). This site is also surrounded by a mountainous complex with maximum height 500–600 m at the western and eastern sides, forming a small valley (canal) oriented from north to south with an elevation of 200 m at its southern region (see Fig. 1).

The experiment covered a 10-day period (20–29 July 1996) where low or moderate background wind prevails. The instrumentation consisted of 1) a meteorological mast equipped with thermometers at three levels (2, 7, and 13 m), cup anemometers at two levels (7 and 13 m), and a wind vane at 7 m, allowing measurements of wind and air temperature; and 2) a 1600-Hz high-range vertical monostatic sodar, capable of measuring in real time the vertical component of the wind and the thermal turbulent structure of the atmosphere up to a height of 800 m, with a minimum discernible height of 50 m. It should be mentioned that the sodar was developed at the Department of Applied Physics of the University of Athens (DOAP-UOA) and provided variable operating parameters to best suit the experimental conditions (Asimakopoulos and Helmis 1994).

Also, tethered balloon flights were carried out at the
same location for three days (22, 24, and 27 July) under different meteorological conditions, using a specially designed meteorological package by DOAP-UOA, providing measurements of wind speed and direction, temperature, and humidity up to the height of 900 m with a sampling rate of 1 s (Soilemes et al. 1993). The classical psychrometric method is used to measure dry- and wet-bulb temperatures using two identical negative temperature coefficient (NTC) thermistors, while a cup anemometer and an electrically controlled magnetic compass were used to measure wind speed and direction, respectively. Data acquisition was obtained by a microcomputer placed in the airborne package.

Information on the synoptic-scale atmospheric circulation during the experimental period was derived from the “European Meteorological Bulletin” archive. The surface synoptic circulation during the first 2 days of the examined period results in moderate northerly winds, while on 22 and 23 July strong winds—“etesians”—prevail, blowing from the northeasterly sector. In the following days, from 24 to 28 July, anticyclonic circulation predominates over the whole Greek area (see Fig. 2), producing weak synoptic flow.

3. Results and discussion
a. The vertical structure of the atmosphere

Figure 3 shows the vertical profiles of wind direction and speed, temperature, and mixing ratio during three flights of the tethered balloon on 27 July 1996 at 0817–0910, 1027–1119, and 1221–1319 LST. The prevailing wind blows from the northerly/northeasterly sector up to a height of 900 m during the whole morning (Fig. 3a), indicating the maritime air character. The potential temperature profiles show (Fig. 3b) that the air is relatively stable, maintaining the same characteristics above the height of 450 m during all flights. It should be noted that similar characteristics were observed in other studies over coastal areas (Luhar et al. 1998), where the combined superadiabatic and neutral layer, typically 250 m deep, was capped by a fully stable stratification at higher levels. However, different characteristics are evident at the lower levels. While a ground-based inversion is initially present (not shown), after 1030 LST an unstable surface layer develops that intensifies in depth during the next 2 h, associated with the developed IBL. The formation of the IBL is evident from the potential temperature profiles, where the lower part (below 100 and 200 m, respectively, of the two last flights) corresponds to the IBL and the upper part (up to 450 m with intense stability factor) of the MBL flow (Fig. 3b). The depth of the MBL is more evident in the mixing ratio profiles (Fig. 3d), where there is an abrupt change (from 14 to 10 g kg$^{-1}$) of the mixing ratio values for heights below and above 450 m.

According to the wind speed profiles (Fig. 3c), a speed-up effect is evident in the lower levels (less than 200 m) that is probably attributable to the sloping terrain (about...
15\degree) close to the shoreline. At higher levels (more than 500 m) the wind speed is reduced to the corresponding background flow. It is noteworthy that during the fourth flight, after 1221 LST, the air is significantly cooler, as compared with the previous flight (see Fig. 3b) in the lowest 400 m. This, in relation to the substantial increase of the wind speed in this layer (Fig. 3c)—and the mixing ratio (Fig. 3d)—implies the strong advection of cold air masses from the sea over the inland.

Figure 4 shows the acoustic sounder fascimile record, which displays the backscattered signal echo intensity (convective turbulence) as a function of height and time, for the period 0600–1424 LST 27 July. It should be noted that strong echo (black) regions correspond to intense thermal turbulence in contrast to weak echo (white) regions. During the morning (Fig. 4a) a stable layer exists with an inversion layer above at a height that varies between 200 and 400 m. This is due to the influence of the island on the marine ABL, where the development of elevated inversion layers modified the usually unstable or neutral stratification that prevails over the ocean (Petenko et al. 1996). Later (after 0830 LST) an organized thermal plume activity is present at lower levels, from the ground up to 200 m, due to the development of the IBL. This is associated with the local thermal heating of the land. The height of the inversion layering rises continuously up to a height of 450 m. After 1030 LST (see Fig. 4b) the developed sea-breeze cell intensifies the flow from the sea, giving an intense inversion with strong echo returns at heights up to 400
m. The lower part of the layer (the IBL) that extends up to 150 m and the stable layer above (higher than 400 m) retain their characteristics.

Wavelike activity combined with double layering becomes evident on the fascimile record in the 200–400-m layer between 1000 and 1400 LST (see Fig. 4b). It should be noted that the flights of the tethered balloon between 1027–1119 and 1221–1317 LST are evident on the fascimile record as straight lines up to the height of 800 m. The characteristics of these disturbances are investigated by performing a Fourier analysis of the vertical velocity field at different levels within the height range of the sodar. More specifically, fast Fourier transform spectral analysis was employed using 512 points with segment and frequency averaging. Figure 5 shows the vertical velocity power spectrum at the height of 270 m, as it was estimated during the period from 1000 to 1400 LST. It can be seen that there is a spectral density maximum at the frequency of $2.5 \times 10^{-3}$ Hz that corresponds to time periods of about 7 min. It should be noted that the second spectral density maximum corresponds to time periods of 45 min and is attributed to large-scale variations.

To further investigate the nature of these disturbances, the Richardson number ($R_i$) and Brunt–Väisälä frequency ($N$) are estimated for the last two flights of the tethered balloon and for the periods 1027–1119 and 1221–1319 LST on the basis of tethered balloon data for different layers (200–400 m), using the relations:

\[ R_i = \frac{\alpha \theta}{\partial^2 \theta / \partial z^2} \]

\[ N^2 = -\frac{g}{\rho} \frac{\partial \theta}{\partial z} \]

where $\alpha$ is the thermal diffusivity, $\theta$ is the potential temperature, $g$ is the acceleration due to gravity, and $\rho$ is the density.

Fig. 3. (Continued)
FIG. 4. The sodar fascimile record on 27 Jul 1996 from (a) 0600–1000 and (b) 1000–1424 LST.
Figure 5. Vertical velocity power spectrum at 270-m height for the time period 1000–1400 LST 27 Jul 1996 as derived from fast Fourier transform spectral analysis.

Table 1. Estimated values of the Richardson number (Ri), Brunt–Väisälä frequency (N), and period (T) of gravity waves, as derived from tethered balloon data.

<table>
<thead>
<tr>
<th>Date</th>
<th>Time (LST)</th>
<th>Layer (m)</th>
<th>Ri</th>
<th>N (rad s⁻¹)</th>
<th>T (min)</th>
</tr>
</thead>
<tbody>
<tr>
<td>24 Jul 1996</td>
<td>1220–1247</td>
<td>200-400</td>
<td>12</td>
<td>8.64 × 10⁻⁴</td>
<td>12.1</td>
</tr>
<tr>
<td></td>
<td>300–400</td>
<td>13.2</td>
<td>9.11 × 10⁻³</td>
<td>11.5</td>
<td></td>
</tr>
<tr>
<td>27 Jul 1996</td>
<td>1027–1119</td>
<td>200–400</td>
<td>13.4</td>
<td>1.46 × 10⁻²</td>
<td>7.2</td>
</tr>
<tr>
<td></td>
<td>1221–1319</td>
<td>300–400</td>
<td>3.8</td>
<td>1.57 × 10⁻²</td>
<td>6.6</td>
</tr>
<tr>
<td></td>
<td></td>
<td>3.2</td>
<td>1.81 × 10⁻²</td>
<td>5.7</td>
<td></td>
</tr>
</tbody>
</table>

where \( \theta \) is the mean potential temperature, \( g \) is the gravitational acceleration, and \( \frac{\partial \theta}{\partial z} \) and \( \frac{\partial u}{\partial z} \) are the potential temperature gradient and wind shear, respectively. Table 1 presents the estimated values of the Richardson number (Ri), the corresponding Brunt–Väisälä frequency \( (N) \), and the period \( (T) \) of the gravity waves, as derived from the tethered balloon data. From this table, it becomes evident that Ri is characterized by positive values, much larger than unity, suggesting that buoyant oscillations could appear in the form of gravity waves with frequencies equal to or less than the Brunt–Väisälä frequency \( N \) and periods \( 2\pi/N \) (Arya 1988). This is in accordance with the wave activity of the thermal structure of the ABL observed on the facsimile record, and the estimated periods are consistent with the period derived from the Fourier analysis of the vertical velocity field (see Fig. 5). It is worth mentioning that similar disturbances are observed on the facsimile record of the radar on 24 July. The estimation of the same parameters for the period 1220–1247 LST, with the aid of the available balloon data for the same layers, results in relatively longer wave periods, of the order of 10 min, as in Table 1.

Figure 6 displays the profiles of the variance of the vertical velocity component \( \sigma_z^2 \) and the structure parameter of the temperature \( c_T^2 \) obtained by the sodar on 27 July for the period 1230–1310 LST. According to this figure, the variance \( \sigma_z^2 \) increases, as it is expected, up to 120 m and then decreases significantly up to the height of 200 m. At 200 and 300 m, the variance appears relatively constant, due to the presence of the elevated inversion in this layer (see Fig. 3b). On the other hand, the substantial decrease of the temperature structure parameter \( c_T^2 \) in the first 150 m above the ground supports the surface thermal heating with strong mechanical mixing, as expected. The subsequent increase up to the height of 250 m is associated with the above-mentioned inversion layer. Similar observations and characteristics have been reported from other experimental studies close to the shoreline (Helmis et al. 1987, 1995). It should also be mentioned that the observed minimum of \( \sigma_z^2 \) profile at a height of 200 m (Fig. 6) is associated with high wind speeds at heights up to 200 m (see Fig. 3c, the 1221–1319 LST flight). This behavior is in general agreement with Yoshiki (1997), who found that for a small island the mechanical turbulence generally dominates for high wind speeds, and the developed IBL is identified by a low level jet, when combined with a minimum of \( \sigma_v \).

b. The development of the IBL

The top of the mechanically induced IBL may be identified with changes in the velocity shear \( \partial u/\partial z \) or
with a sharp discontinuity in slope when plotted in log-linear form. The depth $h_i$ of the IBL that is developed due to the roughness change—from a smooth to a rough surface—under neutral conditions can be estimated using the following empirical formula (Kaimal and Finnigan 1994):

$$h_i = A_1 \left( \frac{x}{z_{02}} \right)^n,$$

where $n = 0.8$, $z_{02}$ is the roughness length downwind, and $x$ is the downwind distance (fetch). The constant of proportionality $A_1$ depends on the roughness change and varies between 0.35 and 0.75. It can be approximated as

$$A_1 = 0.75 + 0.03 M,$$

where $M = \ln(z_{01}/z_{02})$, and $z_{01}$ is the upwind roughness length.

Also, the vertical profile of the wind within the IBL under neutral conditions is given by the well-known logarithmic law

$$\ln(z) = \frac{u_{*2}}{k} \ln \left( \frac{z}{z_{02}} \right),$$

where $k$ is the von Kármán constant (0.4), and $u_{*2}$ is the downwind friction velocity.

Using Eqs. (3)–(5), the depth $h_i$ of the mechanically produced IBL is estimated during different experimental dates. In our case, the fetch $x$ is about 150 m. Incorporating the meteorological mast data and Eq. (5), the parameters $u_{*2}$ and $z_{02}$ are graphically calculated using a log-linear ($\log(z), \overline{u}$) plot (see Table 2). From this table, it can be seen that the estimated values of $h_i$ are less than 40 m, while their deviations are relatively small. This is expected, since the depth of the mechanically produced IBL actually depends mainly on the roughness value $z_{02}$ and the fetch, which in our case is constant.

It should be noted that the power-law formula (3) with $n = 0.8$ has been found by Bradley (1968) and Antonia and Luxton (1971) to be well fitted with their data. Besides, Arya (1988) and Stull (1988) state that the same formula can be applied to unstable conditions with a slightly larger value of $A_i$ than for the neutral case, thus resulting in relatively larger values of $h_i$. The height $h_i$ of the IBL can also be estimated on the basis of the following widely applied formula (Jegede and Foken 1999):

$$h_i = ax^b,$$

where $x$ is the fetch, and $a$, $b$ are empirical constants.

For fetch distances between 0.01 and 160 m, Walmsley (1989) had obtained values $a = 0.75$ m and $b = 0.8$ m. In our case, the use of (6) led to $x = 150$ m in an $h_i$ value of 41.3 m, which is slightly higher than the graphically estimated values given in Table 2. This is in a close agreement with the results of Walmsley (1989).

In order to include changes in scalar fluxes (for temperature and moisture) the development of the convective internal boundary layer (CIBL) after a cold–hot transition has to be taken into account. There are several formulas to determine the depth of the CIBL, taking into account the mechanically (shear) produced turbulence.

Table 2. Estimated mechanically induced IBL parameters.

<table>
<thead>
<tr>
<th>Date</th>
<th>Time (LST)</th>
<th>$u_{*}$ (m s$^{-1}$)</th>
<th>$z_{02}$ (m)</th>
<th>$h_i$ (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>22 Jul 1996</td>
<td>1000</td>
<td>0.89</td>
<td>1.2</td>
<td>34.2</td>
</tr>
<tr>
<td>24 Jul 1996</td>
<td>1005</td>
<td>0.34</td>
<td>0.8</td>
<td>30.8</td>
</tr>
<tr>
<td>27 Jul 1996</td>
<td>1030</td>
<td>0.67</td>
<td>1.05</td>
<td>32.8</td>
</tr>
</tbody>
</table>
Figure 7. Vertical profiles of potential temperature during the flights of the tethered balloon: 22 Jul 1996 (1001–1043 LST), 24 Jul (1005–1053 and 1202–1247 LST), and 27 Jul 1996 (1027–1119 and 1223–1319 LST).

The downwind friction velocity is estimated graphically with the aid of a log-linear plot of log z against T. Using the meteorological mast data and the potential temperature gradient above the thermal IBL, the depth of the mechanically induced IBL is estimated from the balloon data flights, taking into account that the top of the CIBL can be identified by the point where the potential temperature gradient shows a sudden change from an unstable to a stable layer (Venkatram 1986). Figure 7 gives the vertical profiles of the potential temperature for all experimental days, using the daytime balloon flights. It is worth noting that the temperature gradient, though it has different values above the IBL layer for the different days, remains almost constant during all the daytime flights of the same day.

The estimated parameters that lead to the calculation of the CIBL depth, employing meteorological mast data and the potential temperature gradient above the thermal IBL from selected flights, are presented in Table 3. From this table, it can be seen that the depth of the mechanically induced IBL, as it was expected, varies approximately from 60 to 110 m. In order to compare these results with the experimental data, the vertical profiles of the potential temperature for all experimental days are examined (Fig. 7).

### Table 3. Estimated CIBL parameters for selected tethered balloon flights.

<table>
<thead>
<tr>
<th>Date</th>
<th>Time (LST)</th>
<th>( \sigma_D ) ( \times 10^{-2} )</th>
<th>( T_{02} - T_{01} ) (°C)</th>
<th>( \frac{\partial T}{\partial z} ) ( \times 10^{-3} ) (°C m(^{-1}))</th>
<th>( h_u ) (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>22 Jul 1996</td>
<td>1000–1020</td>
<td>4.41</td>
<td>3.8</td>
<td>2</td>
<td>112.1</td>
</tr>
<tr>
<td>24 Jul 1996</td>
<td>1005–1025</td>
<td>1.69</td>
<td>3.1</td>
<td>1.8</td>
<td>59.8</td>
</tr>
<tr>
<td>24 Jul 1996</td>
<td>1202–1220</td>
<td>1.84</td>
<td>5.7</td>
<td>2</td>
<td>88.7</td>
</tr>
<tr>
<td>27 Jul 1996</td>
<td>1030–1050</td>
<td>3.61</td>
<td>6.0</td>
<td>6.7</td>
<td>69.6</td>
</tr>
<tr>
<td>27 Jul 1996</td>
<td>1200–1240</td>
<td>3.49</td>
<td>7.2</td>
<td>4.7</td>
<td>89.5</td>
</tr>
</tbody>
</table>
layer. In the next flight, at 1202–1247 LST, the top of the IBL rises to 200 m, as is indicated by the height of the inversion. It is characteristic that the gradient of the potential temperature above 150–200 m remains constant during the two flights, indicating the maintenance of the maritime boundary layer characteristics, despite the depth change of the IBL. On 22 July 1996, relatively stronger northerly background flow prevails as compared to the previous days, which is also reflected in the vertical profile of the wind speed (not shown). The potential temperature gradient above the IBL is almost the same, but the $\sigma_p$ value is more than double that of 24 July, when the IBL also develops. In this case, the estimated depth of the IBL is larger than on the previous days, which is also found in the temperature profiles (Fig. 7). However, it should be noted that the increased value of $h_u$ (112.1 m) in Table 3 is due to the large value of $\sigma_p$ rather than to the enhanced thermal inhomogeneity.

On 27 July, the wind blows again constantly from the

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**Fig. 8.** Vertical profiles of (a) wind direction and (b) potential temperature, during the flights of the tethered balloon at 2134–2222 and 2313–0007 LST 24 Jul 1996.
northeastern sector, due to the developed sea-breeze cell (see Fig. 3), but with greater stability of the MBL ($\theta^* / \partial \zeta$ takes values 2 or 3 times more than the previous cases). Larger values of the $T_{02} - T_{01}$ differences and higher $\sigma_p$ values form a value of $h_u$ of the order of 100 m, which is in agreement with the slope of the temperature profile in Fig. 7. It is worth noting that the observed profiles result from the IBL being combined with a speed-up effect, thus inducing greater depth of the IBL. Therefore, the comparison between the experimental data (potential temperature profiles of Fig. 7) and the estimations of $h_u$ shown in Table 3 for the three experimental days reveals that the experimental values of $h_u$ are in fairly good agreement with the estimated values, using Eq. (7).

c. Nocturnal airflow

Figure 8 shows the vertical profiles of the wind direction and temperature on 24 July 1996 during the nighttime for the periods 2134–2222 and 2313–0007 LST. In this case, southerly/southwesterly moderate wind prevails within the first 250–300 m above the ground (Fig. 8a), a result of the combined katabatic flows forming along the surrounding mountain ranges and the slope flow channeling close to the experimental site (see Fig. 1). At higher altitudes the wind is substantially reduced and turns to the northeasterly sector (Fig. 8a), following the background flow. In the temperature profile, a stable layer that is evident above 350 m (Fig. 8b) is clearly related to the airflow from the open sea. At the lower levels, a ground-based inversion is developed, progressively reaching different heights from 100 to 350 m. This layer seems to be associated with the development of the local katabatic and channeling flows. This thermal structure is also evident on the sounder fascimile record (see Fig. 9). In accordance with these records, the inversion layering developed between 2030 and 2330 LST with variable depth at different heights, and it was followed by a combined ground-based and elevated inversion, which eventually formed a strong surface-based activity. This inversion reached the height of 500 m after 2330 LST, maintaining a different stability factor at the various heights.

4. Concluding remarks

The study of the vertical structure of the ABL under different meteorological conditions over the inland revealed the strong influence of the isolated island on the marine ABL structure. Thus, interesting characteristics of the local flows were observed with the modified MBL alongside the developed IBL on the shoreline, in relation to the depth, the stability factor, and the thermal structure for all studied cases.

- During the nighttime, when the background flow is weak, a complex wind flow is formed, resulting from the development and combination of katabatic flows and topographic channeling flows, with strong stability and depth of the order of 300 m.
- During the daytime, the formation of an IBL up to a height of 100–150 m is evident, depending mainly on the stability of the incoming airflow, and the temperature difference between sea and land.
- When the background flow permits the development
of a sea-breeze circulation, the MBL close to the shoreline has a depth of the order of 400 m, with intense structure and strong layering. In many cases, wavelike activity is developed at elevated layers, with periods of 7 min.

- Well-known empirical relationships for the estimation of the height of the IBL, including changes in roughness and scalar fluxes, provided satisfactory results in comparison with the experimental measurements.

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