

Evolution of the Thorpe and Ozmidov scales at the lower atmospheric and planetary boundary layer

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Abstract. Turbulence affects the dynamics of atmospheric processes by enhancing the transport of mass, heat, humidity and pollutants. The global objective of our work is to analyze some turbulent descriptors which reflect the mixing processes in the atmospheric boundary layer (ABL). In this paper we present results related to the Thorpe displacements d_T , the maximum Thorpe displacement $(d_T)_{max}$, the Thorpe scale L_T and the Ozmidov scale, L_O , and their time evolution in the ABL during a day cycle. A tethered balloon was used to obtain vertical profiles of the atmospheric physical magnitudes up to 1000 m. We discuss their vertical and time variability, and also their relationships.

1. Introduction

The atmospheric boundary layer (ABL) is the lower part of the atmosphere characterized by a strong interaction with the underlying surface. The ABL turbulence affects its dynamics by enhancing the physical diffusion and changing the evolution of different processes. Atmospheric turbulence is an important subject in the atmospheric sciences because most of the fluxes in atmosphere depend on turbulent diffusion. Estimation of the dissipation of energy by turbulence and turbulent diffusion are important for a full understanding of ABL energetic and dynamic processes. For example, turbulent mixing and diffusion of natural and anthropogenic gases control ABL composition under stable and unstable conditions.

In the absence of turbulence, atmospheric temperature profiles become increasingly monotonic, due to the smoothing effect of molecular diffusion that occurs at Bachelor/Kolmogorov scales when the 3D turbulence cascade produces molecular mixing. Turbulence at larger scales as well as other causes such as fluid instabilities or internal wave breaking makes vertical overturns that appear as inversions in measured temperature profiles. These overturns produce small-scale turbulent mixing in ABL which is of great relevance for many processes ranging from medium to a local scale. Unfortunately, measuring at those small scales is very difficult. To overcome this disadvantage it is interesting to use theories and parameterizations which are based on larger scales because they are more easily accessible by conventional instruments.

The vertical extent of turbulent overturns can be obtained from the temperature profiles, and their study is based mainly on length scale analysis. Vertical overturns, produced by turbulence in density stratified fluids as lakes or the ABL, can often be quantified by the Thorpe displacements d_T , the maximum displacement length $(d_T)_{max}$ and the Thorpe scale L_T .

As mentioned, there is a great interest to use theories and parameterizations for small-scale dynamics which are based on larger scales –as L_T or $(d_T)_{max}$ -. But there are also several more reasons. The correlation between L_T and the Ozmidov scale L_O , defined below, can be used to estimate rates of dissipation of turbulent kinetic energy and vertical turbulent diffusivities which describe the efficiency of turbulent mixing at small scales. So a deeper insight on this relation is helpful to estimate mixing, at least that associated with patches of high turbulent activity. Another reason is related to the theories of turbulent stirring which often depend on hypotheses about the length scales of turbulent eddies (for example, mixing length theories).

One of the aims of this work is to analyze the behaviour of overturning length scales, which can be used as a tool to infer the small-scale dynamics of turbulence from the largest overturns present in profiles. The other interest of this paper is to calculate the Thorpe displacements, the maximum Thorpe displacement, the Thorpe and the Ozmidov scales at the ABL because we want to study the properties of the ABL turbulent patches which represent one of the dominant processes for mixing. And, finally, we also want to analyze their time evolution during a day cycle in order to relate it with the ABL stratification conditions.

Next we present the atmospheric data used for the analysis, in sect. 3 we present the Thorpe method and the definitions of the scale descriptors used and, finally, the results are presented and discussed.

2. Atmospheric data sets and meteorological instrumentation

The results presented in this paper are based on ABL data from 98 balloon soundings made in Almaraz (Cáceres, Spain) with a tethered sonde system. The data were selected for this analysis because they cover different stratified conditions –stable, unstable and neutral- and mixing conditions – from shear-driven turbulence to convective regions-. A total of 98 successful soundings approximately ranging from 150 m to 1000 m were carried out. We used ABL profiles obtained during balloon flights launched from 25TH to 29TH September 1995 in the time intervals 6:00-12:00 a.m. and 3:00-12:00 p.m. And from 5TH to 10TH June 1994 in the time intervals 5:00-12:00 a.m. and 5:00-12:00 p.m.

The meteorological conditions vary from clear to slightly cloudy for the 1995 campaigns whereas for the 1994 campaigns the situation consisted of slightly cloudy skies or clean skies with cloudy intervals.

The instrumented balloon for the ABL measurements was launched near the nuclear power station of Almaraz (CNA, nuclear central of Almaraz). Almaraz is located around 110 km away from Cáceres city, on the west of the Spanish plateau. This area is topographically influenced by the plain on the Tajo riversides and the Almaraz mountain range. The weather in this region is continental. The surroundings of the power station are constituted by pastures and meadows [1].

The ABL data were registered by a tether balloon sensing system. The main instrumentation used is a sounding system which includes a zeppelin-shaped tethered balloon from which a meteorological

probe was hanging. The equipment was completed by an atmospheric data acquisition system called ADAS [1]. Table 1 shows some characteristics of the sounding system.

Table 1. Experimental characteristics of the sounding system and the meteorological probes.

| Magnitude | Value |
|------------------------------------------|---------------------|
| Sampling frequency | 403 MHz |
| Temperature precision | (0.6, 1) °C |
| Temperature resolution | 0.01 °C |
| Atmospheric pressure precision | 3 hPa |
| Atmospheric pressure resolution | 0.1 hPa |
| Vertical data resolution | $\Delta z = 8$ m |
| Sampling interval | $\Delta t = 25$ min |
| Sampling interval of sequential profiles | 1 h |

3. Thorpe method and overturn length scales

Turbulence produces vertical overturns that appear as inversions in measured temperature profiles, as mentioned before [2]. Thorpe devised an objective technique for evaluating a vertical length scale associated with overturns in a stratified flow [2]. Thorpe’s method is commonly used in oceanic and lake measurements, also in laboratory experiments (for example, study of the vertical overturns occurring behind a biplane grid in a continuously stratified water channel [3]), and in the tropo-stratosphere to analyze turbulent parameters [4], but has never been applied to ABL measurements, probably because of the large horizontal deviations that free flying balloons have. The technique is on the other hand quite used in oceanography [2-5].

For horizontally homogeneous flows, Thorpe’s technique evaluates a vertical length scale associated with overturns in a stratified flow as follows [2]. If one considers an instantaneous vertical density profile $\rho(z)$, as the left graph of figure 1 shows, part of this profile appears to contain inversions which are gravitationally unstable. Thorpe’s technique consists of rearranging this density profile $\rho(z)$ so that each fluid particle is statically stable. The result of the method is a stable monotonic profile which contains no inversions. Anywhere in the initial profile where a gravitational instability exists, the displacement from the original position to the re-ordered stable position, known as the Thorpe displacement d_T , is non-zero.

Imagine a density profile as consisting of i samples of density ρ_i , each of which was observed at depth z_i . If the sample at depth z_n must be moved to depth z_m to generate the stable profile, the Thorpe displacement d_T is $z_m - z_n$ (figure 1). The distances $d_T(z)$ each fluid particles have been displaced are shown in figure 1, right graph, and are called Thorpe displacements.

Each displacement d_T represents the vertical distance that each fluid particle –characterized by a measured quantity- has to move up- or downward to its position in the stable monotonic profile. As turbulent eddies are not one-dimensional, the displacement d_T is not necessarily the distance sample actually traveled [5].

The Thorpe displacements let us to define overturns as a profile section for which $\sum_i d_{T_i} = 0$ while

$d_{T_i} \neq 0$ for most i [5], [6].

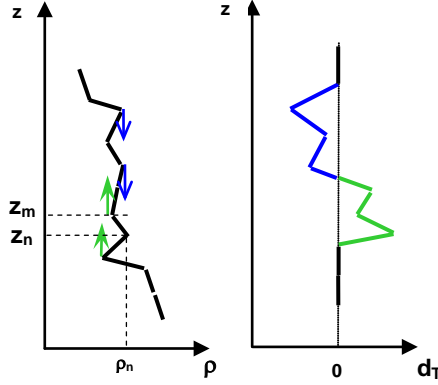


Figure 1. Scheme of a distinguishable overturn as well as the corresponding Thorpe displacements with their characteristic Z pattern.

Two Thorpe scales can be defined from the root mean square (*rms*) and the maximum of the Thorpe displacements, respectively, as [2], [3], [5]:

$$(L_T)_{rms} = \langle d_T^2(z) \rangle^{1/2} \quad (1)$$

and

$$(L_T)_{max} = \max[d_T(z)] \quad (2)$$

where $\langle \rangle$ signifies an appropriate averaging process - average over a single overturn -.

The scale $(L_T)_{rms} = L_T$ characterizes the turbulent motion at the time of the measurements and it is a statistical measure of the vertical size of overturning eddies [5], [7]. Although overturns are not one-dimensional, the Thorpe scale gives a good estimate of an overturn size as long as the mean horizontal gradient is much smaller than the vertical gradient. The scale $(L_T)_{max}$ represents the larger overturns which might have occurred at an earlier time when buoyancy effects were negligible [3] and, in some cases, is considered as an appropriate measure of the overturning scale.

Different researchers have found a linear relationship between L_T and $(d_T)_{max}$ for profiles from the equatorial undercurrent [6], [8]. For microstructure profiles from lakes under very different conditions of mixing and stratification, a power law - as $(d_T)_{max} \sim (L_T)^{0.85}$ - is found [9], [10].

Because of the expensive nature of collecting data at microscale resolution, many theories have been proposed that allow small-scale mixing rates to be estimated from coarser-resolution data. A basic goal is to determine the kinetic energy dissipation rate ε , which is related to the stratification and a turbulent mixing length, the Ozmidov scale L_O (Ozmidov 1965).

Ozmidov [11] considered that buoyancy limits the maximum vertical scale for overturns in turbulent stratified fluids. By setting the buoyancy forces equal to the inertial forces, Ozmidov derived a length scale L_O which would describe the largest possible overturning turbulent scale allowed by buoyancy as

$$L_O = \sqrt{\frac{\varepsilon}{N^3}} \Rightarrow \varepsilon = L_O^2 N^3 \quad (3)$$

where ε is the dissipation rate of turbulent kinetic energy and N the Brunt-Väisälä frequency or stability frequency, $N^2 = -g\rho_0^{-1}(d\rho/dz)$. The scale L_O has the following physical explanation: it corresponds to the vertical distance l that a particle of fluid moves if all its kinetic energy is converted to potential energy (assuming that $\varepsilon \sim w^3/l$, where w is a vertical velocity scale). Therefore L_O is a measure of the maximum size of an overturn in a stratified fluid. It is difficult to obtain an estimate for L_O independently since ε can only be resolved with microstructure measurements.

This relation is helpful to estimate mixing, at least that associated with patches of high turbulent activity [7]. A typical range for the Ozmidov scale is 10^{-2} to 1 m.

Since L_O and L_T are two different estimates of overturning lengths, it is reasonable to expect that a linear relationship exists between the two, although spatial and temporal variability in the turbulent field means that any relation is only valid in a statistical sense. Dillon studies the relationship between L_O and L_T and finds that $L_O = (0.79 \pm 0.4)L_T$ [5]. A number of other studies followed and their measurements showed that the Thorpe scale is nearly equal to the energy containing length scale or Ozmidov scale, L_O . For example, far from the surface in wind-forced mixing layers in the seasonal thermocline the relation $\langle L_T/L_O \rangle = 1.25$ has been reported [5]. Another results present $\langle L_T/L_O \rangle = [0.9, 1.4]$ for measurements of turbulence during conditions of Katabatic winds of mountain overflow [5-7, 8]. This then implies that the turbulent dissipation rate ε of an individual overturn can be calculated directly from the buoyancy frequency N^2 and the Thorpe scale L_T . Therefore, the Thorpe scales can be used to estimate rates of dissipation of turbulent kinetic energy and this is an essential result.

Moreover, the length scale ratio L_T/L_O can be interpreted as a *clock*, which increases monotonically as the turbulent event evolves [7], [12]. The ratio changes from values less than 0.5 for young, preturbulent overturns to about 1 after the transition to turbulence and increases beyond 1 as turbulence finally decays.

We can introduce also the Ellison scale that is another dynamical quantity used to estimate the overturning eddy size. The Ellison scale L_E is based on density ρ instead of temperature T , and is the first length scale based on statistics from empirical data. Its definition is

$$L_E = \frac{\langle \rho^2 \rangle}{\partial \rho / \partial z} \quad (4)$$

or

$$L_E = \frac{\rho'}{d\bar{\rho}/dz} \quad (5)$$

This length scale descriptor is the typical vertical displacement traveled by fluid particles before either returning towards their equilibrium level or mixing [13].

The Ellison scale deviates slightly from the procedure for the Thorpe scale L_T estimation. It is often assumed that there is also a linear relationship between L_T and L_E ($L_T \approx 1.2L_E$ [3]), but this is not often the case and then other parameters such as the mixedness are needed [14]. Both scales, L_E and L_T , are considered adequate measures for the overturning eddy size and, generally, agree well with the Ozmidov scale.

4. Quantitative results

Our methodology uses potential temperature instead of density and is based on reordering 98 measured potential temperature profiles, which may contain inversions, to the corresponding stable monotonic profiles. Then, the vertical profiles of the displacement length scales $d_T(z)$ can be calculated. To get the Thorpe displacements profiles, a bubble sort algorithm with ordering beginning at the shallowest depth was used in the analysis [2], [3], [5]. Bubble sort algorithm is a simple sorting algorithm. It works by repeatedly stepping through the data list to be sorted, comparing each pair of adjacent items and swapping them if they are in the wrong order. The pass through the list is repeated until no swaps are needed, which indicates that the list is sorted. The algorithm gets its name from the way smaller elements *bubble* to the top of the list. There are other sorting algorithms that have better performance than bubble sort; but one significant advantage that bubble sort has over most other implementations is that the ability to detect that the list is sorted is efficiently built into the algorithm. It performs better on a list that is substantially sorted having a small number of inversions, as in temperature vertical profiles.

Thorpe displacements are useful as a visual aid to define the vertical extent of some mixing events. The Thorpe scale is proportional to the mean eddy size as long as the mean horizontal potential temperature gradient is much smaller than the vertical gradient as happens at ABL.

4.1. Thorpe displacements at ABL

Figure 2(a), (b) and (c) shows the time evolution of the measured raw temperature profiles during a day cycle (from 6 GTM to 24 GTM). These profiles show the boundary-layer evolution during a diurnal cycle. From sunrise profiles it is clear the stable boundary layer and the residual layer –from 6:00 to 7:00 a.m.-. Later, the stable boundary layer is broken slowly and it appears a small mixed layer –at 8:00 a.m.- which evolves during the day and it changes the whole profile. From afternoon to night, the profiles look adiabatic and at midnight a small stable boundary layer grows.

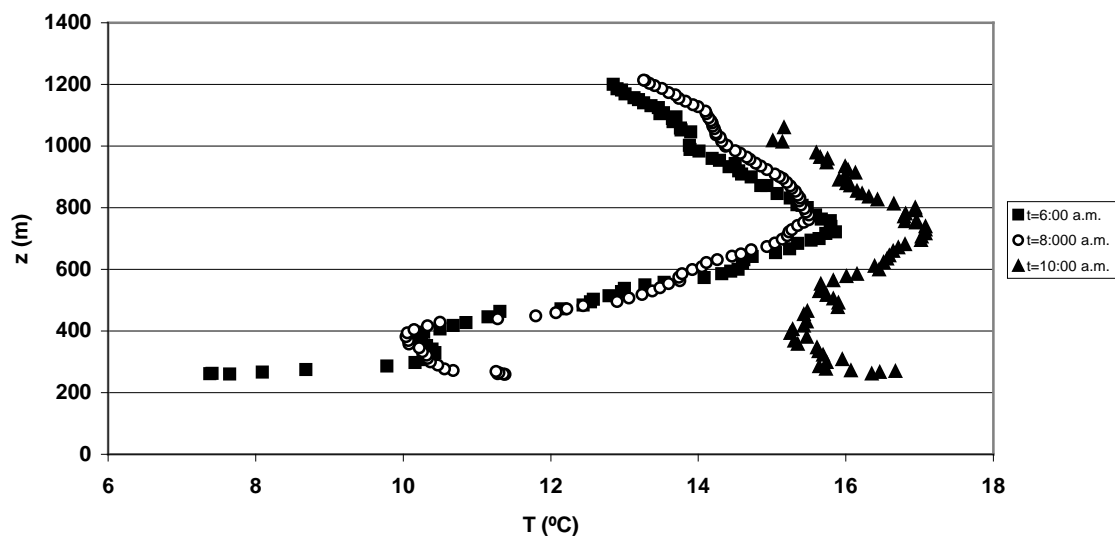


Figure 2(a). Time evolution of the raw temperature vertical profiles on 27TH September of 1995. The times of the profiles are shown at the right legend.

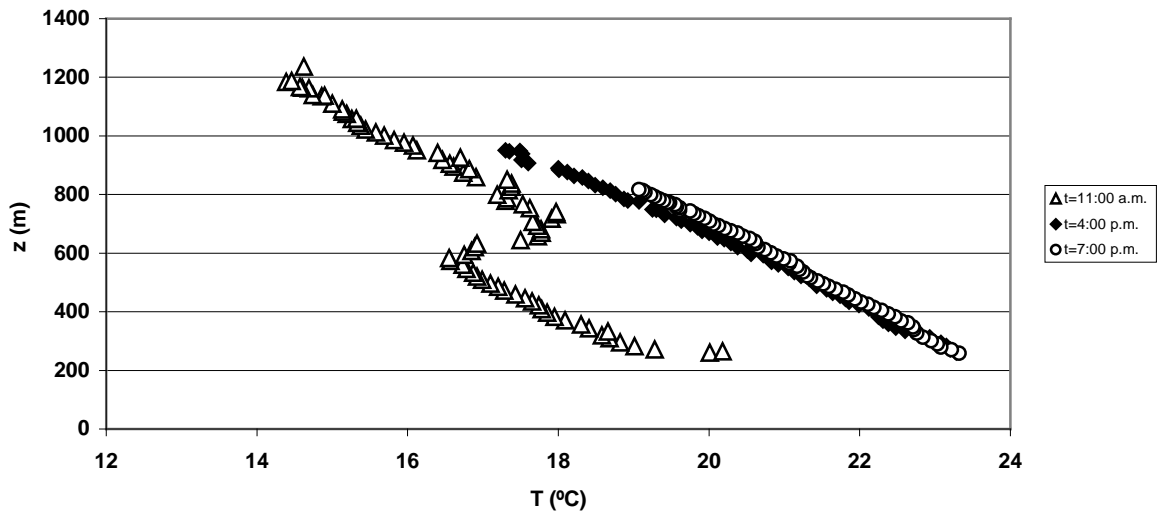


Figure 2(b). Time evolution of the raw temperature vertical profiles on 27TH September of 1995. The times of the profiles are shown at the right legend.

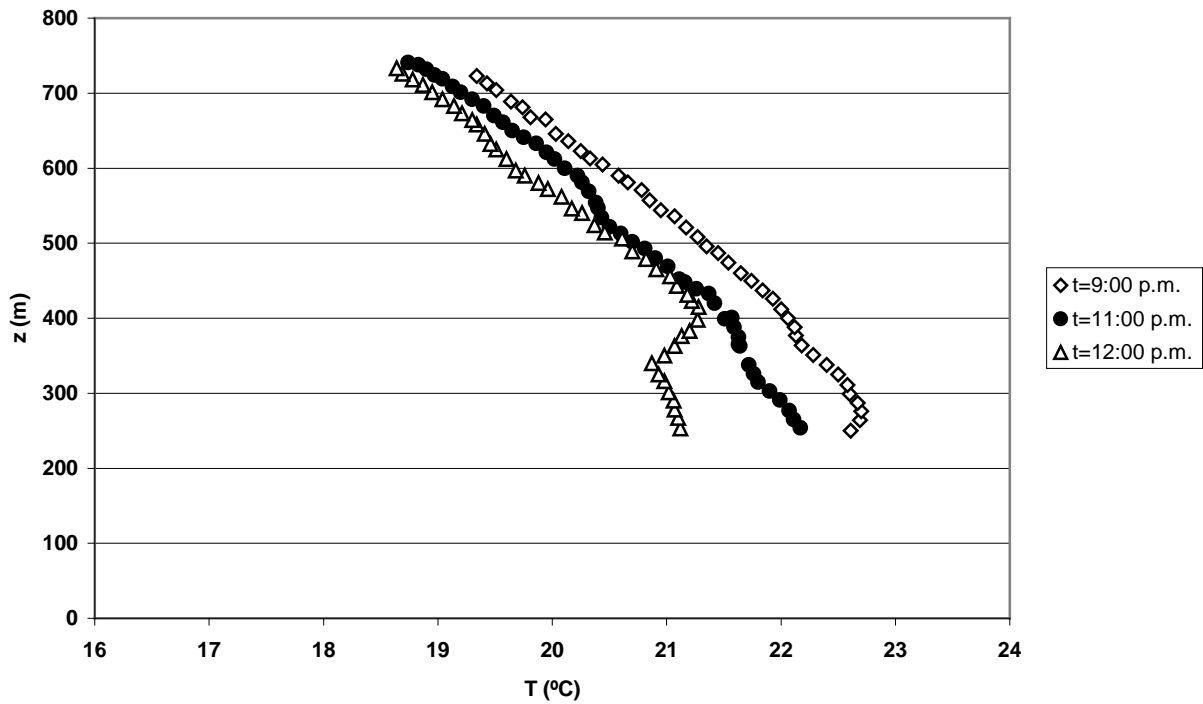


Figure 2(c). Time evolution of the raw temperature vertical profiles on 27TH September of 1995. The times of the profiles are shown at the right legend.

Figures 3 to 6 show the behaviour of the real potential temperature, θ , (left curve) and the corresponding calculated stable monotonic profiles, θ_s , (right curve) that were obtained from 07 GMT to 24 GMT. From real potential temperature profiles we reach the stable monotonic ones by means of the Thorpe displacements (central curve). As mentioned before, the stable monotonic profiles are gotten when the Thorpe displacements are applied to the real temperature profile.

These examples of calculated stable monotonic profiles, shown in figures 3 to 6, correspond to the campaign made the 25TH September of 1995. The stable profiles at 07:00 a.m. and 19:00 p.m. are very similar to the real potential temperature profiles. This happens under neutral stratification conditions. But the monotonic stable profiles at 11:00 a.m. and 17:00 p.m. are smooth and they are very different from the corresponding real profiles that are irregular with a lot of sharp features in the stratification. And all this occurs under convective conditions.

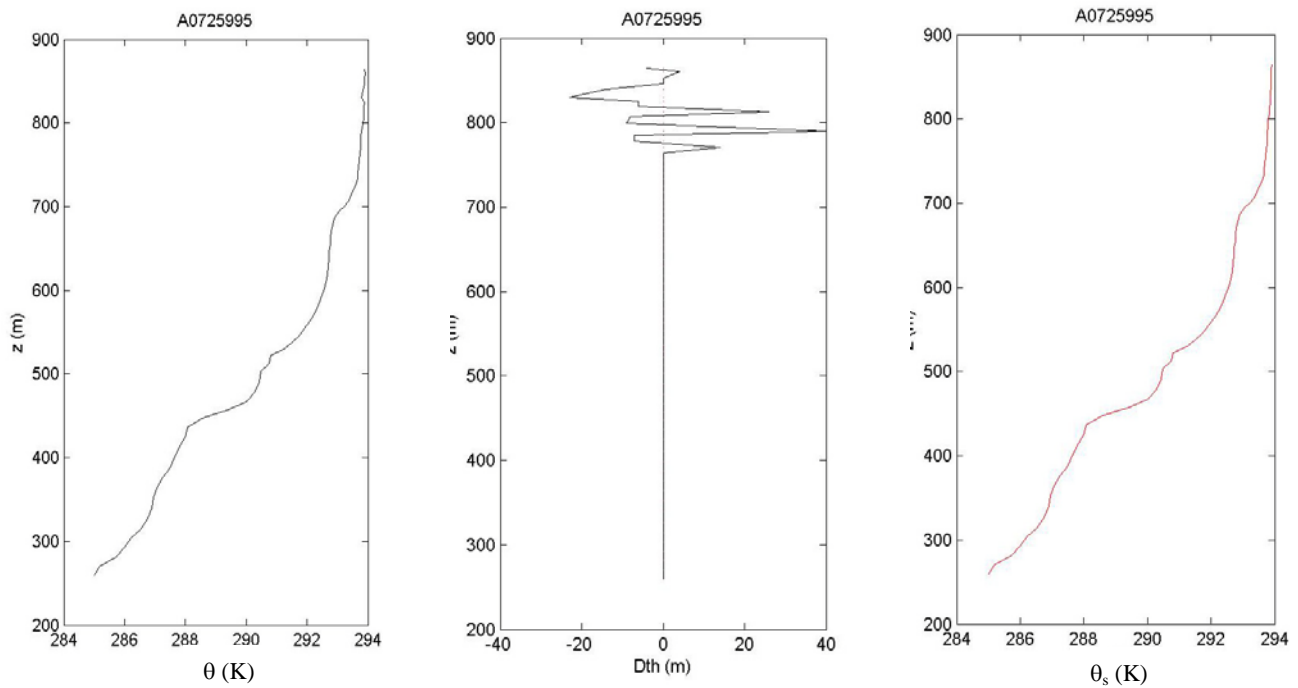


Figure 3. The real potential temperature profile (left curve) and the corresponding calculated stable monotonic profile (right curve) by means of the Thorpe displacements profile (central curve) corresponding to 07:00 GMT (25TH September of 1995).

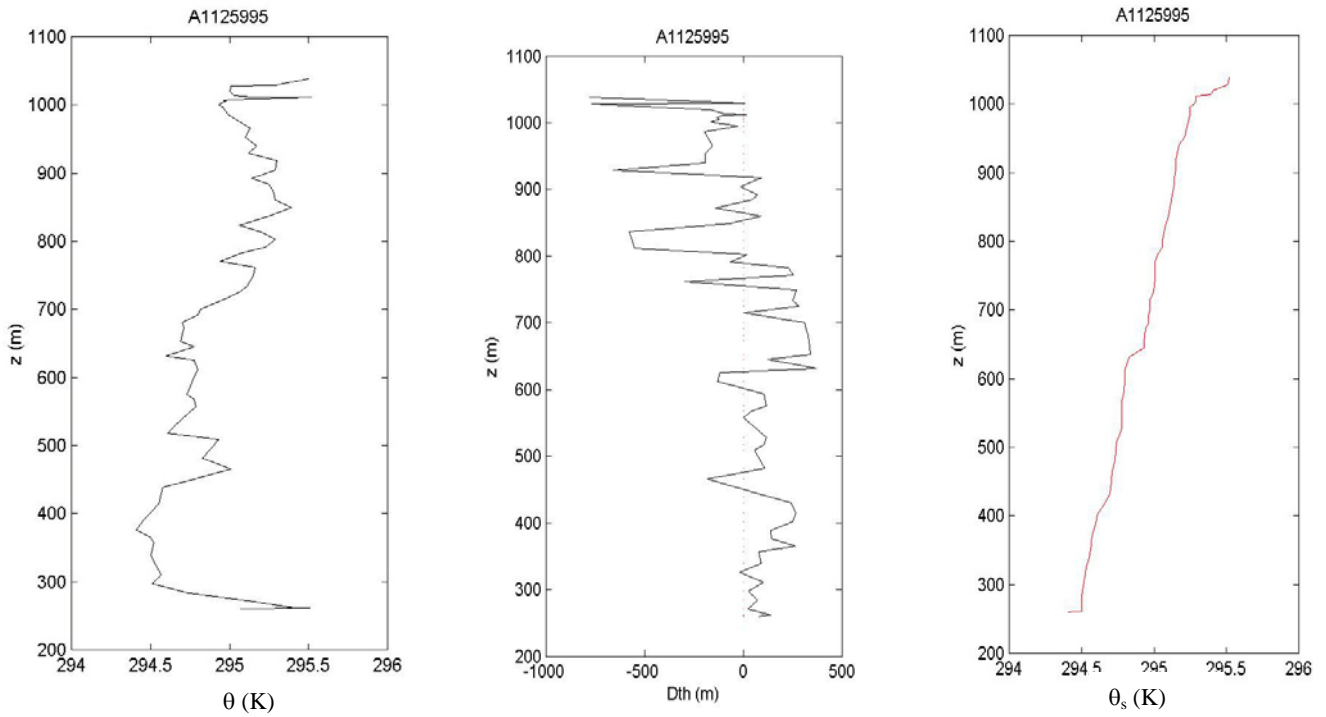


Figure 4. The real potential temperature profile (left curve) and the corresponding calculated stable monotonic profile (right curve) by means of the Thorpe displacements profile (central curve) corresponding to 11:00 GMT (25TH September of 1995).

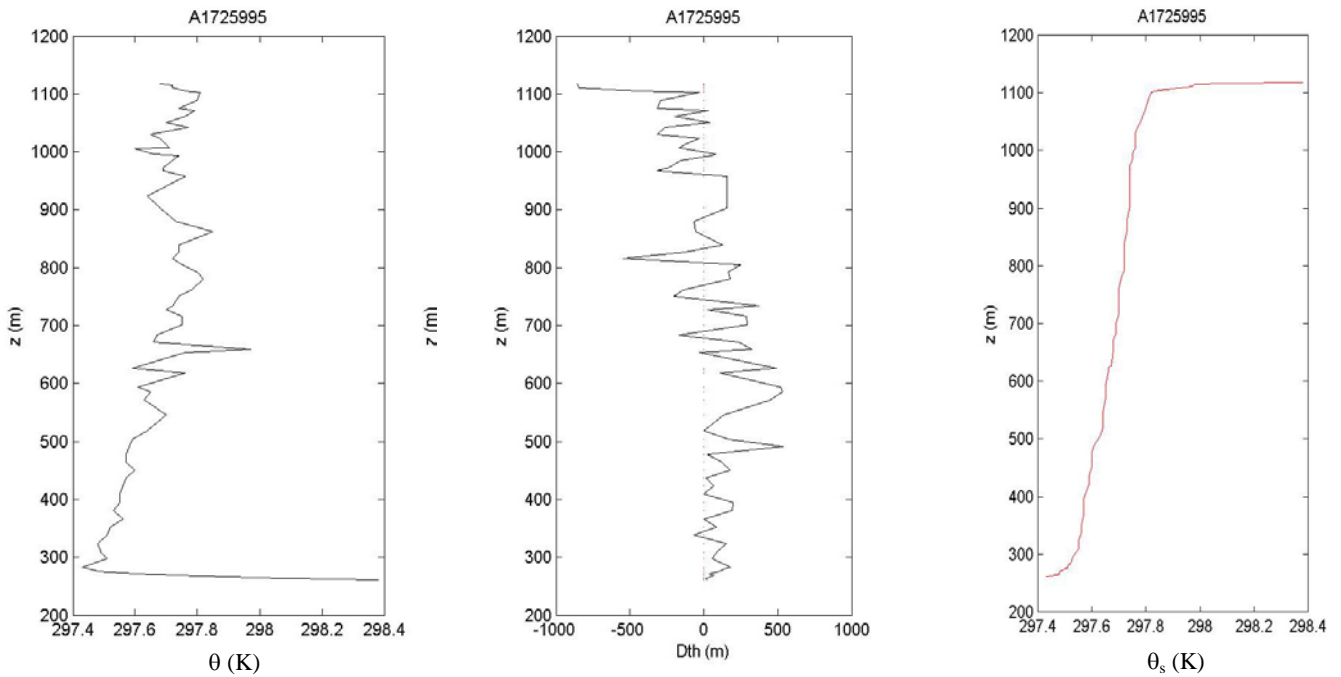


Figure 5. The real potential temperature profile (left curve) and the corresponding calculated stable monotonic profile (right curve) by means of the Thorpe displacements profile (central curve) corresponding to 17:00 GMT (25TH September of 1995).

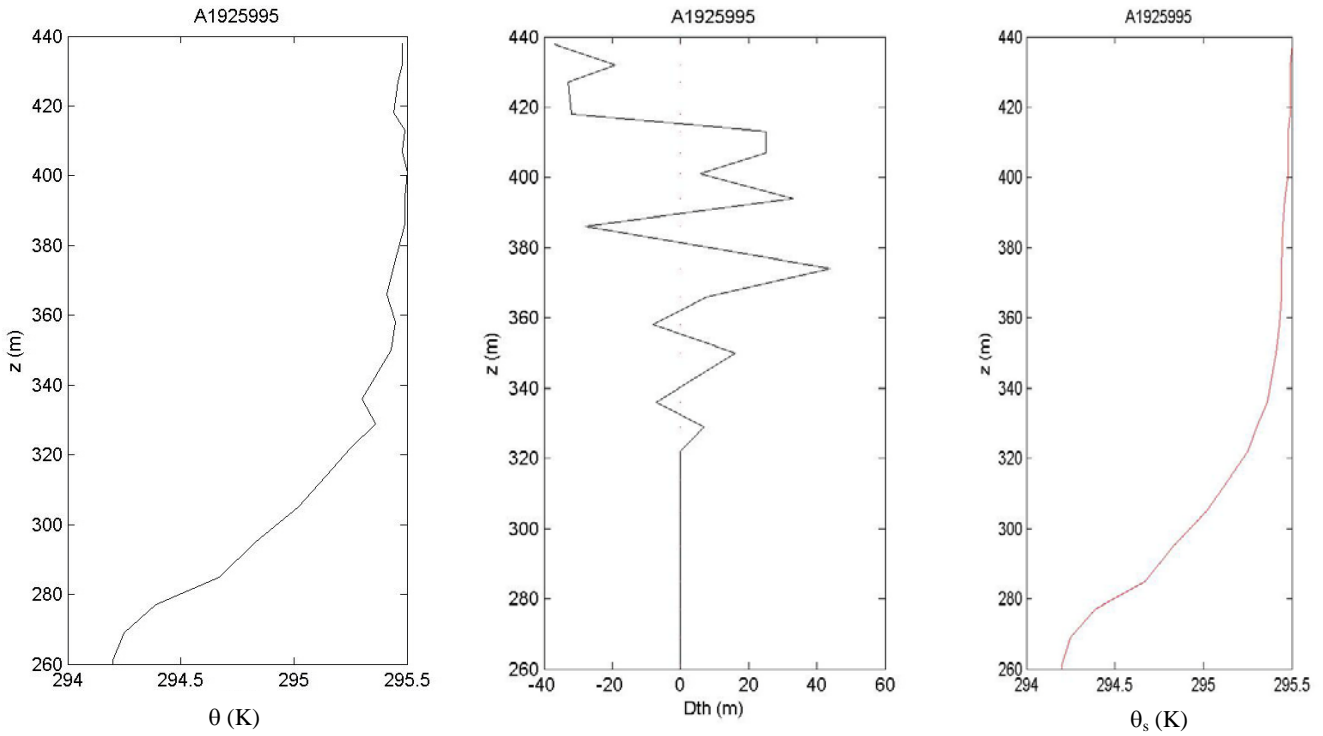


Figure 6. The real potential temperature profile (left curve) and the corresponding calculated stable monotonic profile (right curve) by means of the Thorpe displacements profile (central curve) corresponding to 19:00 GMT (25TH September of 1995).

The Thorpe displacements profiles at 07:00 a.m. and 19:00 p.m. - neutral stratification conditions - are clearly different from the profiles calculated at 11:00 a.m. and 17:00 p.m. - convective conditions. For the first case, Thorpe displacements are always zero except in a top region with isolated Z patterns which would correspond to discrete patches. For the second case, Thorpe displacements never are zero for the whole profile.

Figures 7 to 10 show the time evolution of the real potential temperature, θ , the potential temperature fluctuations, θ' , the vertical potential temperature gradient, $d(\theta)/dz$, and the Thorpe displacements, d_T , profiles from 07 GMT (at sunrise) to 24 GMT. The data correspond to the campaign made the 28TH September of 1995. We calculate the vertical potential temperature gradient because we need to use it to determine a region where is approximately constant. This region will be used to evaluate the vertical average which appears in Thorpe scale's definition.

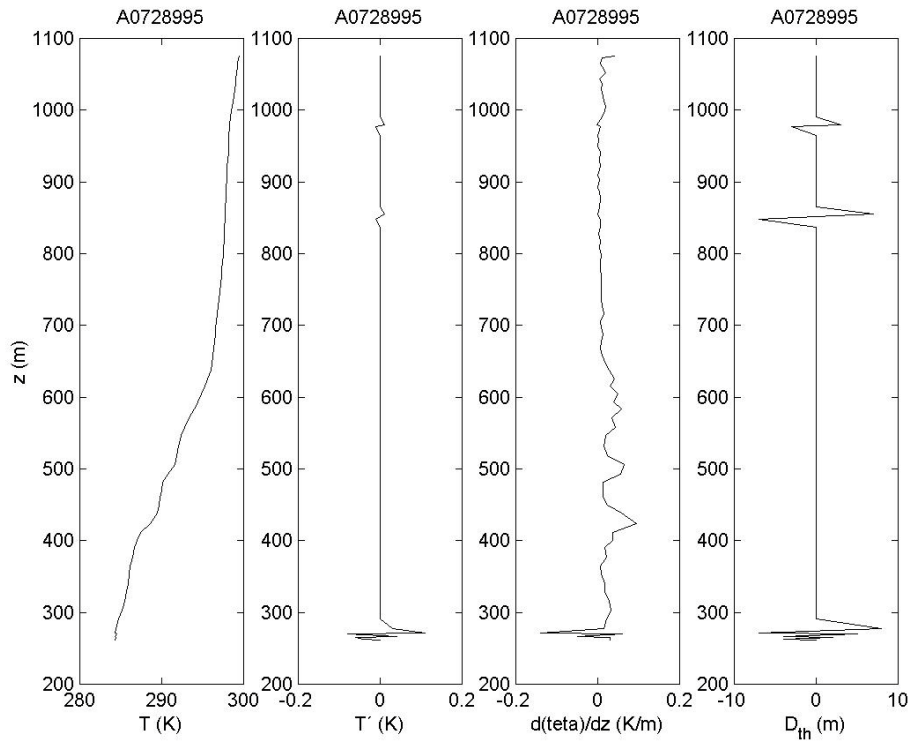


Figure 7. From left to right, each graph shows the potential temperature, the potential temperature fluctuations, the vertical temperature gradient and the Thorpe displacements profiles at 07 GMT (approximately at sunrise).

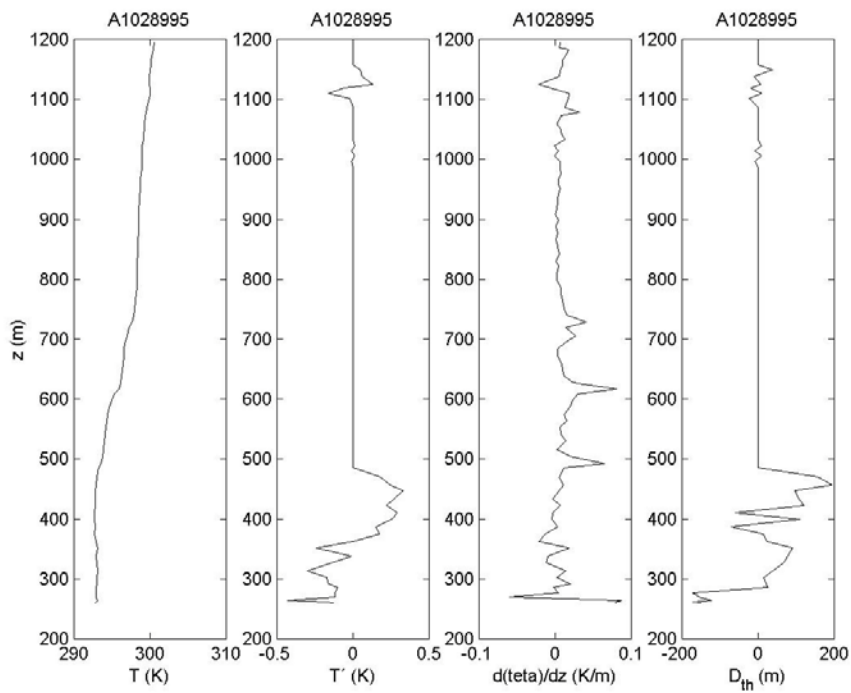


Figure 8. Idem as figure 7, but at 10 GMT.

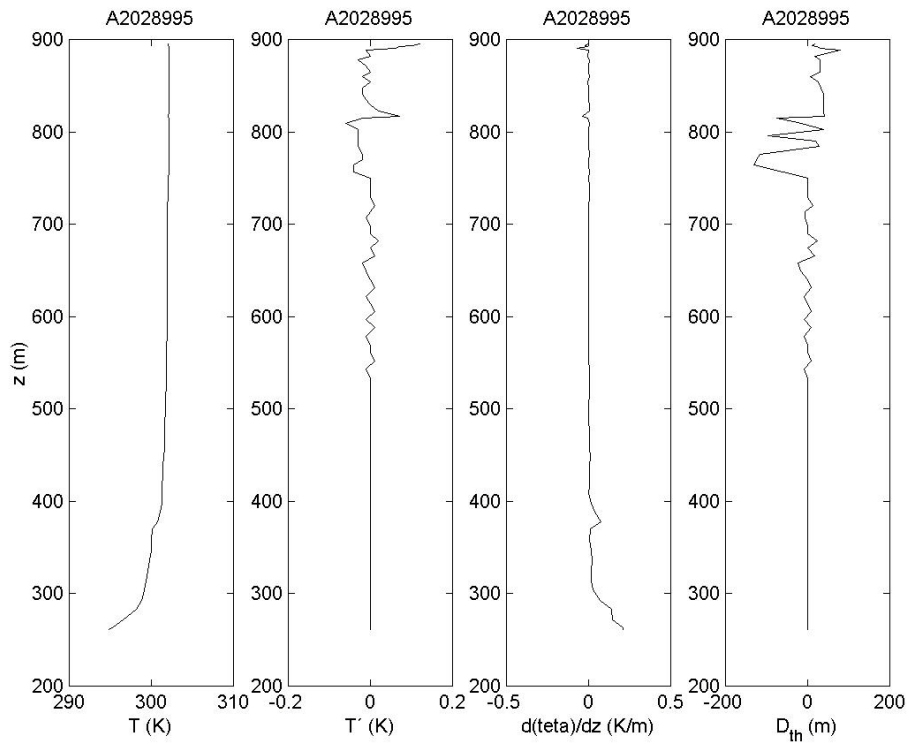


Figure 9. Idem as figure 7, but at 20 GMT (approximately at sunset).

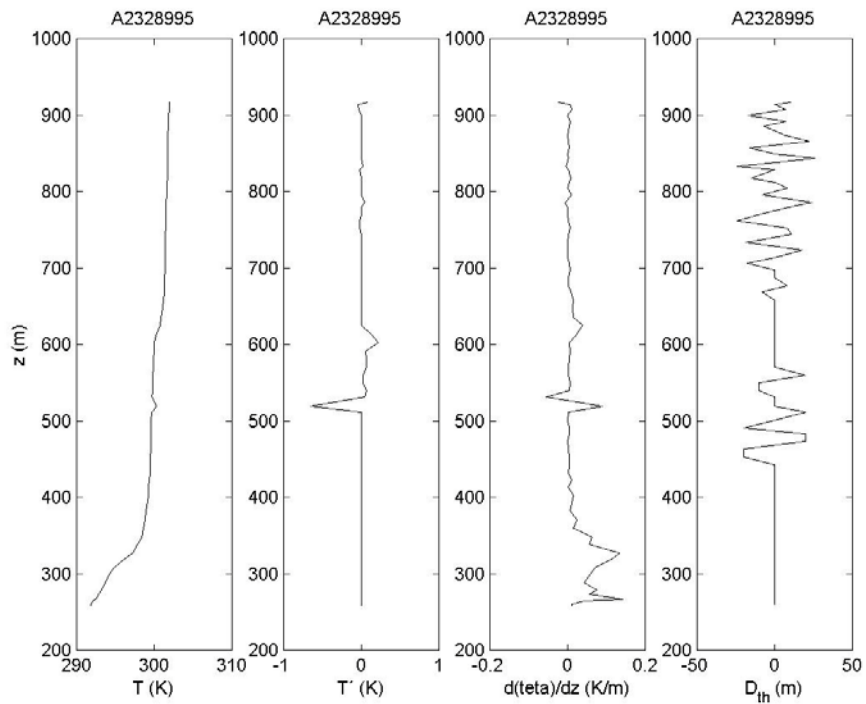


Figure 10. Idem as figure 7, but at 23 GMT.

Thorpe displacements observed at profiles could be qualitative classified in two groups: isolated Z patterns corresponding to discrete patches and non-zero Thorpe displacements. The isolated overturns are very few well-defined sharp overturns and they appear under stability conditions (at sunset, night and sunrise profiles). Usually, the signature which might be expected for a large overturning eddy is: sharp upper and lower boundaries with intense mixing inside - displacement fluctuations of a size comparable to the size of the disturbance itself are found in the interior -. While common in surface layers strongly forced by the wind, these large features are not always found as in our ABL case.

We really find other features that are smaller, some having an eddylike shape similar to the larger disturbances, some a random mix of small scale fluctuations without sharp boundaries. These are the non-zero Thorpe displacement regions with indistinct and distributed features which appear under convective and/or neutral conditions (at noon, afternoon and evening profiles) and are smaller.

4.2. Time evolution of Thorpe and Ozmidov scales

At section 3 we have defined the Thorpe scale (see equation (1)). The Thorpe scale's definition has an average $\langle \dots \rangle$ that has been chosen as a vertical average over a region of constant potential temperature gradient, that's meant, the absolute value of the gradient is less than 10 per cent of the maximum temperature gradient [3]. First, we have calculated the Thorpe displacements for every potential temperature profile. And later, we have deduced the Thorpe scale using equation (1).

Figure 15 shows the time evolution of the maximum Thorpe displacement, $(d_T)_{max}$, and the Thorpe scale, L_T during a day cycle.

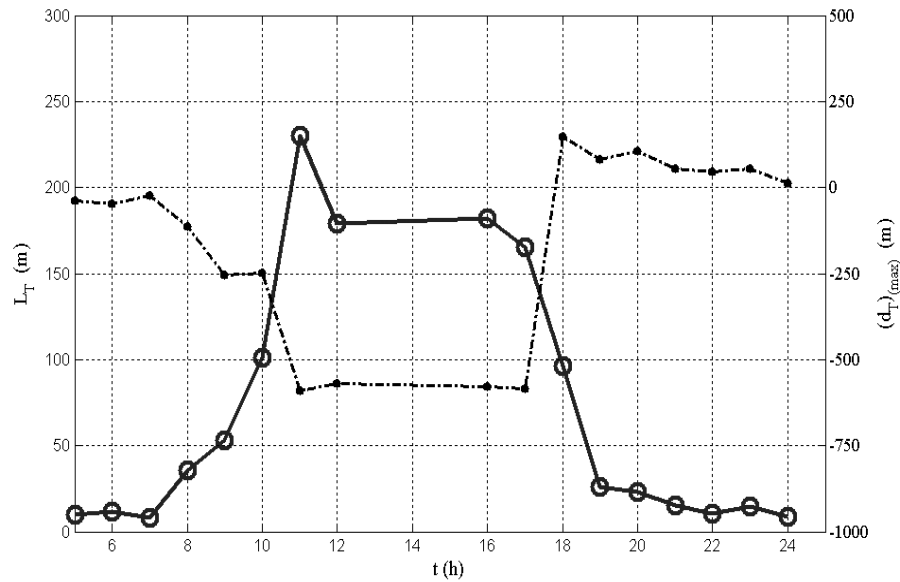


Figure 15. The dotted curve shows the time evolution of the maximum of Thorpe displacements, $(d_T)_{max}$. The continuous curve is the Thorpe scale time evolution.

The scale $(d_T)_{max}$, or maximum Thorpe displacement, is approximately zero under stability conditions (between sunset and sunrise); it reaches a minimum region at noon under convective conditions and it reaches a small maximum in the evening hours under neutral conditions. We can observe that the scale $(d_T)_{max}$ has always negative values when it is not approximately zero and it is only positive in the evening hours –from 18:00 to 24:00-. These results means as follows. We have

defined Thorpe displacements as the difference between the final height and the initial height of the fluid particle. Then, we have the following behaviour:

$$d_T = (z_m)_{final} - (z_n)_{initial} \begin{cases} \text{If } (z_m)_{final} > (z_n)_{initial} \Rightarrow d_T > 0 \\ \text{If } (z_m)_{final} < (z_n)_{initial} \Rightarrow d_T < 0 \end{cases} \quad (6)$$

Therefore, if $d_T > 0$ the fluid particle has to go up to reach its stable position, and if $d_T < 0$ the fluid particle has to go down to reach its stable point. Figure 15 shows the time evolution of the maximum of Thorpe displacements, $(d_T)_{max}$ and we can deduce the following behaviour during a day cycle. Fluid particles only go up in the evening hours under neutral conditions and under the rest of stratification conditions -convective-, the fluid particles go down.

The Thorpe scale L_T is approximately zero under stability conditions (between sunset and sunrise); it reaches a clear maximum under convective conditions at noon and a secondary maximum in the afternoon hours under convective to neutral conditions.

We observe that the Thorpe scale L_T reaches its maximum values -at 11:00 a.m.- approximately at the same time interval as $(d_T)_{max}$ reaches its minimum values - from 11:00 a. m. to 17:00 p.m.-. Both scales, L_T and $(d_T)_{max}$, are close to zero at sunset, midnight and sunrise. Moreover, the Thorpe scale is always positive (or approximately zero) during the day cycle, but the maximum Thorpe displacement is always negative except at evening (from 18:00 p.m. to 24:00 p.m.).

Then, there are two distinct behaviours with high ($L_T > 150$ m) and low ($L_T < 10$ m) magnitudes of the Thorpe scales. We propose that in most of the patches in inner layers the Thorpe scale does not exceed several meters and they appear under stable and neutral conditions when the Thorpe displacements are related to instantaneous density gradients. In contrast, under convective conditions, Thorpe scales are relatively large and may be related to convective burst.

The probability distribution of Thorpe scales was also analyzed using the estimates of L_T . The empirical probability of Thorpe scale follows approximately the exponential model, which assumes the highest probability for very small amplitudes of L_T [1]. Another researchers have suggested that exponential distribution can serve as a good approximation for the Thorpe scale, at least at lakes [10]. This hypothesis is approximately confirmed by our results [1]. The exponential model could be used for L_T distribution where turbulence is highly intermittent and generally weak, but it is not relevant for active turbulent regions such as permanent wind-induced turbulent zones where mixing generates turbulent eddies of the sizes proportional to L_T (probability of very small L_T is low) [1, 15].

As mentioned before, the maximum Thorpe displacement can be considered as an appropriate measure of the overturning scale, and some researchers have found a linear relationship between L_T and $(d_T)_{max}$ (section 3). Therefore, we also study such relationship. Figure 16 shows a graph which represents that for the ABL data we observe a linear relation between these two scales, L_T and $(d_T)_{max}$ with a correlation coefficient $r^2 = 0.9157$.

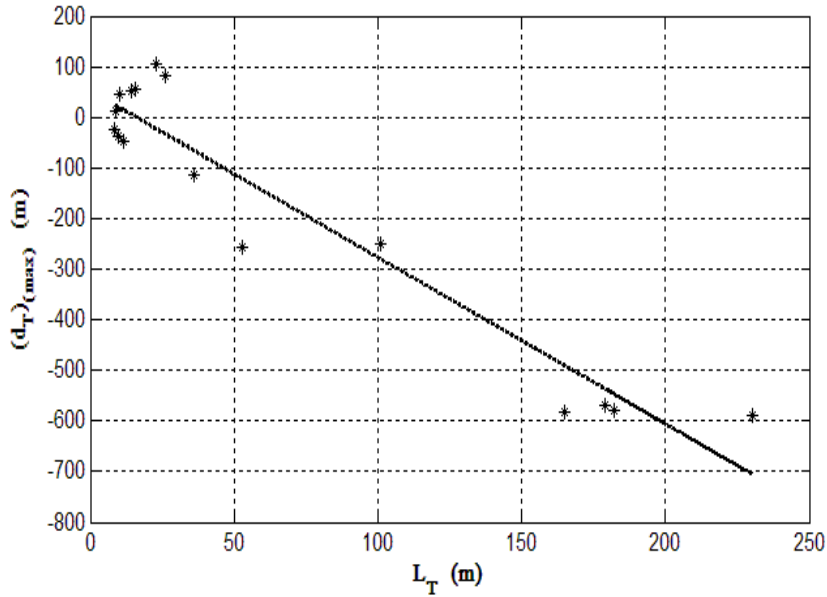


Figure 16. The maximum of Thorpe displacements, $(d_T)_{max}$ versus the Thorpe scale L_T . The corresponding linear adjustment is shown: $(d_T)_{max} = -3.29 L_T + 52.5$ with a correlation coefficient $r^2 = 0.91$.

In the future, we will choose to use the Thorpe scale rather than the maximum displacement because we only sample vertically while the turbulence is three dimensional and, therefore, the Thorpe scale or *rms* displacement is more likely to be a statistically stable representation of the entire feature. This is important because we have to choose an appropriate overturning scale to make a comparison with the Ozmidov scale at ABL data [5, 7, 10].

The vertical scale, at which the buoyancy force is of the same magnitude as the inertial forces, is called the Ozmidov scale L_o and defined by equation (3). The Ozmidov scale quantifies the maximum size of overturning eddies for a given level of turbulence (characterized by the dissipation rate of turbulent kinetic energy ε) and stratification (characterized by the stability frequency N). Experimental evidence indicates that the Ozmidov scale (L_o) and the Thorpe scale (L_T) are strongly related and about equal [5].

We have analyzed the time evolution of the Ozmidov scale during a day cycle and the results are shown in figure 17. The Ozmidov scale L_o is approximately zero under neutral to convective conditions – from 9:00 a.m. to 17:00 p.m.-. About 6:00 a.m. to 7:00 a.m. it reaches a secondary maximum under neutral conditions at sunrise and two maximums under convective to neutral conditions in the evening and night hours. We observe that the Ozmidov scale L_o reaches its maximum values –at 11:00 p.m.-.

Both scales, Thorpe and Ozmidov scales, are always positive during a day cycle but they have an opposite behaviour. Thorpe scale reaches its maximum values –at 11:00 p.m.- approximately at the same time interval as Ozmidov scale is approximately zero - from 10:00 a. m. to 19:00 p.m.-. Moreover, the Thorpe scale is approximately zero when the Ozmidov scale is different from zero and it reaches its maximum values (from 05:00 a.m. to 08:00 a.m. and from 19:00 p.m to 24:00 p.m.).

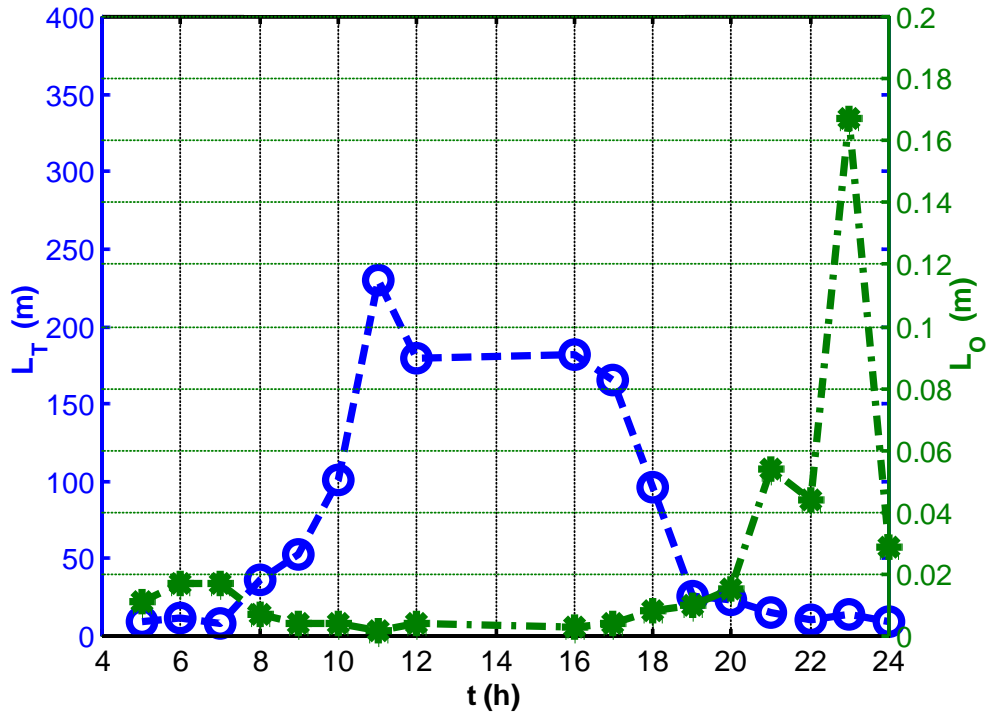


Figure 17. The dotted green curve shows the time evolution of the Ozmidov scale, L_O , during a day cycle and the dotted blue curve shows the time evolution of the Thorpe scale, L_T .

At future, we will study the influence of shear and buoyancy effects on the time evolution of the Thorpe and Ozmidov scales. The reason is that our ABL data are an example of geophysical turbulence and geophysical flows rarely conform to the simplifying Kolmogorov assumptions. There are three important classes of phenomena that modify small-scale turbulence: shear, stratification and boundary limits, and these effects have to be reflected on Thorpe and Ozmidov scales behaviour.

5. Conclusions

This paper presents some preliminary results related to the time evolutions of the ABL turbulent parameters d_T , L_T , $(d_T)_{max}$ and L_O during a day cycle, regarding different levels of stability/instability and the turbulence developed during a day.

Thorpe displacements observed at profiles could be qualitatively classified in two groups: isolated Z patterns corresponding to discrete patches of identified overturns and the non-zero Thorpe displacement regions with indistinct features. The isolated overturns appear under stability conditions. The distributed and indistinct features appear under convective and/or neutral conditions.

We can observe that the scales L_T and $(d_T)_{max}$ have an opposite behaviour under convective conditions –between 11:00 a. m. and 17:00 p.m.– because the Thorpe scale reaches a maximum at 11:00 a.m. and the maximum Thorpe displacement reaches a minimum at the same time –in fact, there appears a “minimum zone” between 11:00 a. m. and 17:00 p.m.–. Moreover, the Thorpe scale is always positive (or approximately zero) during the day cycle, but the maximum Thorpe displacement is always negative except in the evening (from 18:00 p.m. to 24:00 p.m.). Finally, both scales are close to zero at sunset, midnight and sunrise.

The varying height of the well mixed layer and the interaction of boundary layer roughness with the stratification is also directly related with the local entrainment as discussed by [16], terrain shape can interact with the ability of the ABL to produce local mixing very near the ground, this will need

further field work where different conditions are met. For example, the location of mixing events in a 3 or 4 dimensional parameter space formed by (L_O, L_T, L_{MO}, L_t) .

The assumption that the Thorpe scales have a universal probability distribution can be used to verify how accurately the Thorpe scales were computed. It is very likely that the distribution itself or its parameters depend on the governing background conditions generating Thorpe displacements, which are different in the boundary layers from those in the interior layers with intermittent mixing, such measurements of the intermittency of the forcing, as well as that of the actual scale to scale stratified turbulence cascade are discussed in [17, 18] and may be a combination of the boundary condition effects and of stability combining the 3D and 2D characteristics of scale to scale direct and inverse cascades.

The relation of the Thorpe scale L_T and the Ozmidov scale L_O is very interesting because if the Thorpe scale is nearly equal to the energy containing length scale, then the scale L_T can be used to estimate rates of dissipation of turbulent kinetic energy ε using the definition of the Ozmidov length scale, how the ABL affects, either by wind friction or due to thermal effects the upper layers, in spite of having been observed in laboratory experiments [19] is not well documented in the atmosphere..

All these are reasons why this relation will be studied in future research works for atmospheric data sets collected at the ABL. In the future, we will study Thorpe displacement profiles corresponding to ABL data in stable conditions. For this purpose, we will use a set of atmospheric data from SABLES2006 field campaign which took place from 19 June to 2 July 2006 at the CIBA site (Valladolid, Spain). We also intend to analyze the coupling in stable situations (using nocturnal ABL profiles collected at SABLES2006) between the lower Atmospheric Boundary Layer and the structure of the stratified structure taking place at higher altitudes studying the Thorpe displacement associated to shear-driven overturns together with the standard ABL Monin-Obukhov scaling. The complex interaction between the different length-scales in a stable ABL will probably set some limits to the forcing intermittent behaviour detected by [17] as well as producing some non-linear coupling between scales depending on the actual interaction between the forcing and the turbulent cascade.

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