BLLAST
Boundary-Layer Late Afternoon and Sunset Turbulence

White paper, October 2009
Updated July 2010

Authors: M. Lothon, D. Pino, W. Angevine

With the contribution of
R. Beare, F. Couvreux, H. Jonker, D. H. Lenschow, Z. Sorbjan, J. Vilà-Guerau de Arellano.

Part I - Introduction

Context

As an interface between the earth surface and the atmosphere, the planetary boundary layer (PBL) is a critical component of the earth system. It controls transfers of heat, momentum, humidity, and trace gasses between the surface and the atmosphere. The planetary boundary layer (PBL) is the largest sink of turbulent kinetic energy of the atmosphere. It also plays an essential role in the formation of cumulus and stratocumulus clouds with a significant impact on the global radiation and water budgets.

The PBL has still received only small attention at the global scale and has been usually modeled with simple schemes in spite of its complexity. Large discrepancies exist between observations and global models. However, the global climate models have now started to improve their representation of the PBL. A recent example is given by the results of including the eddy mass flux scheme in the global models (Rio et al, 2008).

The numerical simulation of the boundary layer at the mesoscale and smaller scales has been significantly improved during the last years. Intensive observations, remote sensing advances and sensor technologies allow a more accurate and dense exploration of this part of the atmosphere in space and time.

The issues that remain unsolved today and deserve attention due to the associated impacts on the environment are mainly the following: transitions linked with the diurnal cycle; coastal, complex terrain, nocturnal and urban boundary layers; entrainment at the PBL top; and boundary layer clouds.

The present project deals with the transition that occurs in late afternoon, which has an important impact on the interaction of scales (turbulence scales and mesoscales) and on the transport of scalars.

Issues

The growth (FIG. 1) of the convective planetary boundary layer (CBL) over land from morning through early afternoon due to solar heating has been extensively observed and relatively successfully modeled. But the evolution from mid-afternoon on, and the transition from the mixed layer convective boundary layer to a residual layer overlying a stably-stratified surface layer in late afternoon is still not well understood. Even the definition of the boundary layer at that time of the day is uncertain, since there is no consensus on what criteria to use and no simple scaling laws. Yet this transition to the nocturnal boundary layer plays an important role in such diverse atmospheric phenomena as transport and diffusion of trace constituents, wind energy production and storm initiation. The residual layer is, at least for transport purposes, part of the free troposphere, so that water vapour and pollutants emitted at the surface and diluted into the convective boundary layer (CBL) during the day can be introduced in the free atmosphere and transported over long distances often with no interaction with the surface. Moreover, many chemical compounds have also a strong diurnal cycle, and this transition is often an abrupt transition for their own cycle, with a change or reverse of sources and sinks.

At some point in the afternoon, the surface buoyancy flux is not large enough to maintain turbulent mixing, especially for a deep CBL. Yet, vertical motions of about 1 m s^{-1} extending horizontally over several km have been observed, most notably by free flight pilots. The pilots (Pagen, 1988) call this phase of the day the « glass off ». The reason for this large-scale uplift is unclear. According to the pilots, the better the day (large turbulent energy), the better the following ‘glass off’ (large enough positive air vertical velocity over 'large' area). It is described by Aupetit (1989) in the following words (translated from French): ‘The sources [...] of daytime heating decrease at different rates in the late afternoon. Those areas with slower decreases of surface heating produce ascending currents. Nonetheless, the heating rate is much slower than that occurring during daytime convection. Also the ‘glass off’ surfaces must be sufficiently large to be efficient. Ex: Forests, moist soils like wetlands, marshland, peat bogs, rising motion may be enhanced by the fact that humid air above is lighter than the neighboring dry air and so tends to be lifted upward even with no temperature difference between the two. This is why some sites close to the Great Lakes can have exceptional glass off conditions around 17-18 local time. One can quote Lac du Bourget, Lac d'Annecy, Lac de Sainte-Croix, and some peat bogs from Scotland or Ireland.’ This could mean that the super-adiabatic air at the surface, which is inherently linked with the existence of convective boundary layer, could last longer over certain surfaces while the air would start to stabilize due to surface radiative cooling. In any case, the reason for this large-scale uplift remains unclear; possibilities including surface variability and orography, which can induce mesoscale circulation, need to be explored. Cooling and stable stratification may form with even slight depressions of surface elevation while weak heating continues elsewhere.

The scale of these updrafts during the transition seems to be larger than the turbulent scales of vertical transfer during the middle of the day (a few km). Previous large-eddy simulation studies showed during that period of the day, a decoupled residual layer, within which turbulence is still active, develops above the stable surface layer and is characterized by larger-scale updrafts than the mid-day eddies (Sorbjan, 1997). They persist even when the surface buoyancy turns negative. These updrafts may generate smaller-scale eddies that are able to entrain free tropospheric air through the top of the residual layer.

If those structures exist and contribute to the energy and scalar transfers, it is necessary to properly model them with numerical simulations (including large scale of time and space). The eddy mass flux scheme (Rio and Hourdin, 2008) is more realistic than the parameterizations which use diffusion exchange coefficients or mixing length theory for convective boundary layers, but they still cannot take into account those structures that persist after the surface heating source shuts down.

Quantitative observational evidence for this circulation is lacking, partly due to the difficulty of measuring weak turbulence and mean circulations in transitory conditions and at larger scale. Angevine (2007) actually suggests the term 'unforced transition'...

It has been 20 or 30 years since the importance of a better understanding of this phase has been raised (André et al, 1978; Nieuwstadt and Brost, 1986), but up to now, it has remained largely unexplored, from both modeling and observational perspectives.

Part II - State of the art

1) PBL diurnal evolution

The evolution of the low troposphere region, more specifically, the atmospheric boundary layer (ABL) has been studied since the fifties. It can be said that there is an extensive knowledge of the diurnal evolution of the PBL (Sorbjan 2008) and its influence on the pollutant distribution (Vilà-Guerau de Arellano et al. 2004b, 2009; Casso-Torralba et al. 2008). Weather and air quality forecasts have largely benefited from these investigations by introducing new parameterizations that better describe the fundamentals processes involved.

The increasing knowledge of the PBL processes has been based on two main types of studies: the application of the theoretical concepts of turbulence (Batchelor 1967; Tennekes and Lumley 1973; Pope 2000) to perform numerical simulations of the atmospheric characteristics (Lilly 1967; Deardorff 1972; Lenschow 1974; Stull 1976; Moeng 1984; Jacobson 2000, Pielke 2002; Stensrud 2007), and detailed observations in different intensive campaigns: Wangara 1967, Kansas 1968, or Minnesota 1973 (Hess et al. 1981; Kaimal and Wyngaard 1990). Moreover, there are currently systematic observations for example at Lindenberg (Beyrich and Engelbart, 2008) or Cabauw (Van Ulden et al. 1996; Hurley and Luhar 2009; Baas et al. 2009; The Netherlands).

Examples of intensive campaigns to study different aspects of the PBL in combination with numerical simulation and theoretical studies is CASES 99, which has produced an extensive bibliography (Poulos et al. 2002); IHOP (Weckwerth et al. 2004; Couvreux et al. 2005a); LIFT/FLATLAND (Angevine et al. 1998; Cohn et al. 1998); TRAC98 (Bernard-Trottolo et al. 2003). Some of these projects included wind profiler and LIDAR, that have become useful instrument to study PBL evolution (Grimsdell and Angevine 2002; Shaw and Barnard 2002; Fitzjarrald et al. 2004; Angevine 2008).

Most of the studies were devoted to investigation of the PBL characteristics and the relevant processes during the day, when unstable or neutral conditions usually prevail (Kaimal et al. 1976; Mahrt and Lenschow 1976; Stull 1988; Moeng and Sullivan 1994; Cuijpers and Holtslag 1998), or at night when a stable atmosphere is usually found (Nieuwstadt 1984; Garrat 1992; Debyshire 1990; Cuxart et al. 2000; Beare 2006b).

There have been only a few observational studies of transitory processes in the cloud-free atmospheric boundary layer. Based on these studies (which mainly deal with the morning transition) one might conclude that the dynamics of diurnal transitions is controlled by the surface cooling and moistening, wind shear, subsidence, and the presence of clouds [e.g. LeMone and Grossman (1999), LeMone, et al.(2002), Poulos, et al. (2002)].
Acevedo and Fitzjarrald (2001) reported occurrences of specific humidity jumps during

evening transitions, and drops in surface temperature, accompanied by an abrupt decay in wind velocity. Similarly, Mahrt et al. (1999) observed that the latent heat flux during evening events decreased more slowly than the strength of turbulence and the boundary layer depth. This led to the significant moistening of the surface layer. Mahrt (1981), Mahrt et al. (1999) pointed out that the evolution of the stress divergence during evening transitions increased the ageostrophic flow, and led to the development of a low-level jet-stream, accompanied by the decoupling of the flow just above the surface.

Grimsdell and Angevine (2002) used the 915 MHz wind profiler reflectivity and spectral width data to characterize a general behavior of the afternoon transitions. They detected two categories of transitions, those with inversion layer separations, and those with inversion layer descends, which began hours before the sunset. The beginning of the transition period started well before sunset. They defined two types of transitions based on the presence of an intense return of the backscatter signal. However, more recently Angevine (2008) showed that there may be only one type of transition formed by a boundary layer mixing depth decrease with time and a residual inversion above at approximately constant height (FIG. 3). This structure can correspond to varying signals in the wind profiler depending on the temperature gradient and humidity.

During the late afternoon, the combination of the decrease of surface heat fluxes, local increasing stratification, wind shear and the increasing role played by radiation, subsidence or advection can yield to a complex scenario (Vilà-Guerau de Arellano 2007; Angevine 2008).

2) Turbulent energy decay

Several authors have previously studied the transition regimes of turbulence. Through laboratory experiments, Monin and Yaglom (1975) studied the decay of grid-generated turbulence under neutral conditions. During the decay, the turbulence maintains the initial isotropy, with the energy decay following a power law t^{-n} (FIG. 2), where t denotes time. Cole and Fernando (1998), by performing an experiment in a water tank, studied the decay of temperature and velocity fluctuations in a convective turbulent boundary layer in response to cooling at the surface. They found that, when the cooling rate is constant, the decay times of turbulent velocity and temperature scale with a time that is proportional to the temperature difference between the cooling surface and the mixed–layer temperature, and inversely proportional to the cooling rate. Stillinger et al. (1983) studied the decay of homogeneous turbulence in a uniform stratification showing that turbulence becomes highly anisotropic. On the other hand, turbulence decay has been also studied by using theoretical models (George 1992), large–eddy simulation models (LES) (Touil et al. 2002) or direct numerical simulations (DNS) (Biferale et al. 2003).

The first LES study of the decaying atmospheric convective mixed layer was performed by Nieuwstadt and Brost (1986). The authors analyzed an idealized case of the shearless, clear mixed layer, in which turbulence decayed as a result of a sudden shut-off of the upward surface heat flux. The process was described in terms of power time laws (t-t0)-n, similar to those obtained for grid-generated turbulence under neutral, unconfined conditions. In absence of external time scales, the volume integrated turbulent kinetic energy was demonstrated to depend only on the initial state and a dimensionless time $t/t*$ (where $t*$ is the convective time scale, equal to the PBL depth divided by the convective velocity w_*).

The study of Nieuwstadt and Brost was followed by Sorbjan (1997), who considered a gradual change of the heat flux with time, in response to the decreasing sun's elevation. The evolution of the decaying shearless mixed layer showed to be governed by two time scales: the external time scale τ_f , and t^*. When the ratio of these scales was small ($\tau_f /t^* \sim 0$), the decay of the volume averaged dimensionless turbulent kinetic energy was self-similar, and described by the power law $E \sim (t/t^*)^{-1.2}$. When τ_f /t* was large ($\tau_f /t^* \sim \infty$), the turbulent kinetic energy was approximately constant with time, $E \sim (t/t^*)_0$. For various values of τ_f/t^* in the range $0 < \tau_f /t^* < \infty$, the resulting functions describing the decay of the volume averaged TKE were confined between two curves obtained for $\tau_f = 0$ and $\tau_f = \infty$.

Acevedo and Fitzjarrald (2001) undertook a LES study in order to understand the effects of moistening close to the earth's surface during the early evening transition, when the net radiation and sensible heat fluxes at the surface change signs. They observed that residual turbulent activity in the newly forming stable surface layer promoted continuing evaporation. The resulting scalar flux convergence led to an identifiable jump in specific humidity and other scalars in the surface layer. The jumps were accompanied by the inflection point in the temperature time series, which identified the instant at which the surface layer became decoupled from the boundary layer.

The decay of convective turbulence during the afternoon in the atmospheric boundary layer has then been further analyzed by Goulart et al. (2003, 2007) by use of theoretical models; Goulart et al. (2003), Beare et al. (2006a), Pino et al. (2006) and Sorbjan (2008) made more LES studies; Edwards et al. (2006) used a single column model; and Shaw and Barnard (2002) used DNS to study the issue.
In the numerical studies, the decay of the turbulent kinetic energy (TKE) found of course depends in a large extent in the way that the decrease of the surface fluxes is prescribed (some authors consider a sudden drop to zero, while others consider a progressive decrease).

Caughey and Kaimal (1977), Grant (1997), Fernando et al. (2004), Fitzjarrald et al. (2004), Brazel et al. (2005) and Edwards et al. (2006) reported observations of the decay and in some cases compared with model results.
Finally, a particular decaying turbulence process has been studied by some authors during a solar eclipse (Dolas et al. 2002; Girard-Ardhuin et al. 2003; Anfossi et al. 2004; Amiridis et al. 2007; Mauder et al. 2007; Founda et al. 2007; Nymphas et al. 2009). Most of these studies have focused in studying the decay of the turbulent kinetic energy in the boundary layer and how the main variables evolve during the process.

3) Characteristic length scales of scalars and wind components, anisotropy of turbulence

There is a lack of agreement in the evolution the vertical velocity length scale during the late afternoon transition, partly due to the difficulty of addressing the issue, both with numerical studies and observation.

By using LES, Nieuwstadt and Brost (1986) found that the characteristic length scale, defined as the peak of the spectrum, of the vertical velocity variance remained constant during the decay process (which was considered by suddenly forcing surface heat flux down to zero). Sorbjan's study (1997) mentioned previously reflected that small eddies had a tendency to decay earlier than large eddies, as the surface heat flux progressively decreases. Consequently, organized convection persisted in the decaying mixed layer even when the heat flux at the surface became negative (FIG. 4)), and surface inversion was being developed near the earth's surface. These results were later confirmed by the direct numerical simulation of Shaw and Barnard (2002).

Pino et al. (2006) have shown that the characteristic length scale, based on a weighted integral of the density energy spectrum (see eq. 3 of their paper), can have different evolution during the decay, depending on the considered variable (FIG. 5). They found that for all variables, except the vertical velocity for which the scale remained almost constant, the characteristic length scale increases with time. Couvreux et al. 2005b found an increase of the water vapour lengthscale during the afternoon (FIG. 6) of a dry convective case of IHOP.

Grant (1997) dealt with this aspect intensively based on observations. He calculated the spectra of the vertical velocity component at different times and heights during an observational campaign during August 1990 at Cardington, England, and found that during the decay process the spectral peak of the vertical velocity spectra in the boundary layer shifts to smaller length scales.

Fitzjarrald et al. (2004) observed six cases of the evening transition from convective to stable boundary layers by means of flux towers and aircraft measurements. They clearly observed a decrease in the vertical velocity variance, though the horizontal velocity variances decreased more slowly or even increased for some of the nights. Therefore, the differences in the exponent of the variances of the three components of the velocity are a clear indication that the turbulence does not relax to an isotropic state during the decay process. This result was found by Pino et al. (2006) by means of LES.

With the TKE decay itself, the evolution of the characteristic length scales has been one of the main questions addressed in the past studies on the afternoon transition. But the scale issue remains unclear and only partly understood. If the scales in the (future residual) mixed layer seem to increase, it first has to be more thoroughly proved and explained, and the scales might oppositely decrease in the surface layer as the nocturnal boundary layer starts to build. Another important related question is the anisotropy of the turbulence. It has been found ‘squashed’ during the middle of a convective day (Lothon, 2006), but we do not know much how squashed it remains until sunset. The quoted studies above consider different characteristic scales (wavelength of the energy spectrum peak, integral scale, other scales defined with a weighted integral of the spectrum, etc…). During midday, they are often proportional (Lenschow, 1986), but this might change in late afternoon and the various scales have to be considered and their evolution studied.

4) Time scales

Regarding the time scales of the process, previous results (Sorbjan 1997) have shown that the turbulence afternoon decay is governing by two scales: one external time scale related to the time evolution of the surface heat flux and the convective time scale, which is equal to boundary layer depth divided by the convective velocity. The ratio between these two time scales has not been deeply investigated and it can control the setup of the nighttime stable boundary layer.

5) Clouds

Even less is known about the coupling of the afternoon transition turbulence decay and the fair weather cloud dissipation, from the point of view of the entire boundary layer.
Brown et al. (2002) performed a LES study of cumulus convection evolving in time during daytime hours, and compared the numerical results with observations collected during the ARM experiment in Oklahoma. The focus of their study was on the comparison results of eight LES models and the matching observations, rather than on the analysis of diurnal transitions.
Cumulus clouds can also turn to isolated deep convection at this time of the day. During CCOPE-81, LeMone et al (1988) measured the vertical velocity and horizontal wind in the vicinity and below congestus and cumulonimbus clouds, and Nicholls and LeMone (1980), during GATE, compared the heat flux vertical profiles below clouds versus clear atmosphere.
Clouds also modify the surface energy budget through shading. This has been studied at relatively large scale (Betts, 2005), but there is a few on their impact on the energy budget at surface at turbulence scale.

This is a huge and complex issue, that goes somehow beyond the scope of a fundamental study of the afternoon decay. But in the same time, it is a crucial issue, as we at least have to know how large does the cloud fraction and cloud size have to be to make a large impact on the boundary layer behavior.

6) Transport

Recent studies (Vilà-Guerau de Arellano et al. 2004a; Casso-Torralba et al. 2008; Gorska et al. 2008; Vilà-Guerau de Arellano et al. 2009) have shown that morning and afternoon transition are also important for the exchange of species (CO_2 and isoprenes for the cited articles). In early morning, when high entrainment rates have been observed (Vilà-Guerau de Arellano et al. 2004a; Beare et al. 2006), the remaining pollutants of the residual layer are introduced in the shallow boundary layer, increasing the concentration. In the evening, the residual part overlying the stable layer can be incorporated in the free troposphere, so that water vapor and pollutants emitted at the surface and diluted into the convective layer during the day can be introduced in the free atmosphere and transported at larger scale.

Local and regional air quality depends not only on vertical mixing of locally-emitted pollutants, but on horizontal transport from upwind sources. One common case is that pollutants in a well-mixed deep daytime boundary layer are transported in several layers (Banta et al. 1998; Berkowitz et al. 1998), beginning in the evening after the influence of surface friction is removed by the reduction of turbulent mixing. Vertical variations in timing of the afternoon/evening reduction of friction (Mahrt 1981) controls how far and in which direction these layers are transported. A similar effect occurs at upwind coastlines (Angevine et al. 2006; Angevine et al. 1996; Angevine et al. 2004).

Part III – Questions to be addressed

1) Definitions and scaling

Due to its transitional aspect, this phase puts several basic boundary layer definitions into question. The period that we are considering starts as soon as the surface buoyancy flux begins to sharply decrease (late afternoon transition), and it covers the change of sign of the flux (evening transition). The temperature profile close to the surface turns from superadiabatic to subadiabatic, stabilizing the surface layer by the end of the period. Within this context, the mixed layer, the residual layer and the surface layer are non stationary. As a consequence:

- The surface layer cannot be defined in the same way as the process evolves.
- The mixed layer evolves from a well-mixed layer (constant concentration of tracers, constant virtual or equivalent potential temperature) with vigorous turbulence during the previous hours of mid-day, to a stable layer.
- The residual layer during the night will be a weakly stable layer with intermittent turbulence (Tjernström et al, 2009). Both have different interactions with the entrainment zone and capping inversion above and the surface layer below.

During the day, in the convective conditions, most of the moments can usually be scaled based on the surface buoyancy flux (Stull, 1988), from surface up to about 2/3 of the PBL depth. Above this height, the entrainment process can make the scaling less robust. The convective velocity scale, largely used to scale the turbulence throughout the CBL, is a function of the CBL depth and the surface buoyancy flux. And scaling is the base of a robust parameterization in bulk models. On the contrary, during the afternoon transition, the surface buoyancy fluxes are small, and other small forcing processes come into play. So the usual scaling is not relevant anymore, while the stable boundary layer scaling, based on the surface wind stress (Banta et al, 2006), is not relevant yet.

The entrainment zone that overlies the CBL is a transition zone from the mixed-layer to the stable free troposphere, where air from above is engulfed into the mixed layer and vice versa. In LES, it is defined as the layer for which the buoyancy flux is negative, based on a buoyancy flux profile close to linear, positive close to the ground, and getting to negative at the top, with a modulus close to 1/5 of the surface buoyancy flux. But during the late afternoon transition, the buoyancy flux profile has a different shape, similar to a flat 'S', as noted by Sorbjan (1997), and due to non-stationarity. The part of the mixed layer (or residual layer) where the buoyancy flux is negative gets deep, without necessarily corresponding to the place where entrainment takes place (FIG. 7). Moreover, entrainment is usually parameterized through the ratio of the entrainment buoyancy heat flux to the surface buoyancy heat flux. Because the latter gets very small, this parameter, although commonly used for the study and modeling of entrainment, is not relevant either during the afternoon transition (FIG. 8).

From the observations mentioned above, we will need either to define a new scaling approach or to solve issues and work on parameterizations without scaling. And we will also need to clearly define and try to estimate the different layers that we are considering (surface layer, mixed layer, entrainment zone, buffer layer, residual layer, boundary layer top).

2) Observational issues

There are also some issues raised when trying to probe the boundary layer during that time:

- Weak and intermittent turbulence is difficult to measure with any in situ (aircraft, towers) or remote sensing device.
- If the turbulent characteristic length scales are larger, they require larger samples to be well probed. This fact increases the difficulty, especially during a transitory phase. The same holds true for intermittent turbulence. Moreover, the estimates of turbulent moments require homogeneous samples, which is less likely to encounter in both transitory conditions and for larger samples. Therefore, the weak S-shape buoyancy flux profile (FIG. 9) found by Sorbjan (1997) and Pino et al. (2006) and the growth of the characteristic length scales may be difficult to verify in reality, although necessary.
- UHF wind profilers and C-band radar are very useful tools for the study of the PBL. However, *odd* phenomena are usually observed during the late afternoon transition (as well as during the early morning transition), that can be either a sudden sharp drop of the sensitivity (FIG. 10), or a saturation of the receiver due to excessive signal. The latter is often explained by migrating birds, as many species migrate only during the night, starting around sunset and stopping around sunrise. The first phenomenon remains unexplained. But in both cases, the meteorological information is difficult to extract from the signal.

3) Key processes

As sensed from the state of the art drawn up previously, the role of several key elements will be important to study:

- Does the mixing height decrease in the afternoon ? If so, what processes are important for that decrease ? Is subsidence necessary, or are radiation and entrainment sufficient ?
- Turbulence: what are the characteristic of the turbulence during the afternoon transition ?
- How does entrainment evolve ? What processes control it ? What are its roles in the late afternoon transition ?
- Top inversion: what is the role of shear, stability, and large scale subsidence at the top of the mixed layer/residual layer?
- Surface: What is the role of the land-use, topography and heterogeneity variability of the surface? How heat storage and albedo come into play?
- Baroclinicity and advection: What is the contribution of baroclinicity in the observed circulations and larger scale turbulence and of horizontal advection?
- Clouds: Fair weather clouds play an important role in the atmospheric dynamics because they enhance vertical movements, favor their organization, participate to the entrainment, they have their own 'life'. Once formed, they create a new forcing from the top of the mixed layer that can last even when the original surface forcing has stopped. What is their role on the vertical transport, entrainment and horizontal circulation? What is their impact on the characteristic scales? What conditions during this phase, make them turn to deep convection or flatten and dissipate? How does their impact on the surface energy budget through shading come into play ?
- Radiation: Since the surface buoyancy flux is weak, radiation can have a relative significant contribution during this phase, both at surface and at the top of the ML.
- Gravity waves: What is the interaction between the waves that can develop in the stable layers (below and above) and the mixed-layer?

4) Impact

- How does the afternoon transition influence the atmospheric compounds distribution? This transition makes an abrupt variation on spatial distribution and vertical structure of gas, aerosols and scalars (FIG. 11). Their vertical profiles change rapidly from well-mixed to stratified, the sources and sink are changing, sometimes reversing; the dynamics that control them is driven by changing sources too.
- There is a strong influence of the dynamics on the various chemical regimes. The nature of the issue and impact depends on the compound of interest and its life time. The land/vegetation interaction is a key for carbon dioxide and methane, as well as the impact of the transition on their transport and dispersion. For ozone, nitrogen and biogenic compounds, the dynamic processes can affect their reactivity.
- What is the impact of the multi-layer setting (stable boundary layer, residual layer, buffer or entrainment layer) on the trace gas, aerosol, chemical compounds and scalars (as a function of their nature too)?
- In many regions, a low-level jet often sets up during the night until morning, generated by the withdrawal of the turbulence and baroclinity. How important is the transition for the setting nocturnal jet? And what is the impact on the transport of chemical compounds and scalars?
- Similarly, the late afternoon transition is also the time, -or precludes the time-, of reversal of land/see breeze, valley winds and slope winds. And the same questions raised for the nocturnal low-level jet can be addressed for sea breeze and valley wind. The horizontal variability has probably an increasing role during this part of the day, in the circulations observed.
- The afternoon transition sets the initial conditions for the nocturnal boundary layer, which is notoriously complex and difficult to understand. Forcings are weak and timescales are therefore long. It may be that the nocturnal boundary layer is rarely in equilibrium with the local surface, and quasi-stationarity cannot be assumed (Garratt 1992). The relationships of friction velocity (Basu et al. 2008) or geostrophic wind (Shi et al. 2005) are dual-valued and which branch of the solution is taken may depend on how the afternoon transition occurs. The resulting regime choice (strongly vs. weakly stable) (Mahrt et al. 1998) may be "sticky", that is, the atmosphere may persist in that regime for some or all of the night (Shi et al. 2005). The wind profile in the nocturnal boundary layer, including the low-level jet, depends on the stability regime, with obvious implications for wind energy and local air quality. The stability regime also influences the formation of fog or frost (Grant 1997).

The results of the fundamental study of the processes that occur during the afternoon transition and their impacts on the chemical compounds transport and diffusion and on the initial conditions for the next nocturnal stable layer have further impacts and applications in air quality, wind energy and in numerical weather prediction.

Improved understanding of the afternoon transition should lead to improvements in parameterizations of boundary-layer turbulence in numerical models, so that they can better predict the phenomena mentioned above. Some studies of parameterizations have included afternoon transitions (Steeneveld et al. 2006; GABLS2 paper to come from G. Svensson).

Centres such as ECMWF and the UK Met Office still adopt a first-order (non-prognostic) closure for the boundary layer. Others (e.g. Canadian Met Centre, Météo-France) adopt a prognostic (TKE scheme) to represent the convective boundary layer, like those described in Siebesma et al (2007), Soares et al (2004) or Pergaud et al (2009). Detailed observations and modelling of the evening transition (when the prognostic terms are significant) will determine whether prognostic schemes are required for weather and climate models.

Part IV – Strategy

We plan to use the combination of the observations in real world, laboratory experiment, large eddy simulation and mesoscale numerical simulations from the start of the project, in order to connect as much as possible our understanding of the observations to the improvements of the parameterisations of the processes in the bulk models.

1) Data mining in preparation of the field campaign

The work will start with the analysis of some previous datasets, since several campaigns made in the past for different objectives can be appropriate for the study of the afternoon transition. This data mining includes both data analysis and modeling. The considered experiments are:

- AMMA (2006): African Monsoon Multidisciplinary Analysis. The field campaign took place in West Africa, with several meso- or super- sites.

  Extensive observations were made at Niamey, Niger. We will focus on IOPs during which radiosounding balloons were launched 8 times per day. Remote sensing devices of several wavelengths were operated continuously during summer 2006, as well as surface flux stations.

  One LES case made for this experiment has been extended to the end of the day for this study.
- LIFT (1996): Lidar In Flat Terrain, concomitant with FLATLAND experiment, in Champaign, Illinois, USA. The HRDL (High Resolution Doppler Lidar) measured RHI and vertical profiles of Doppler velocity for several days during summer 1996. Previous works have been done on 11 cases when the lidar was pointing up, on the turbulence statistics of the air vertical velocity in the mid-day CBL: coherence, integral scales (Lothon et al, 2006), spectra (Lothon et al, 2009) and higher moments (Lenschow et al, 2009). The data allow us to extend our analysis to the late afternoon, with caution on sampling issues due to the transitory aspect of the considered period.
- CERES (2005): CarboEurope Regional Experiment. This experiment took place in the Landes Forest, in France, for the study of the carbon dioxide cycle. Extensive radiosoundings and flights at several times in the same day can be considered, with measurements of an instrumented tower as well.
- IHOP (2002): International H2O Project. This project was conducted in the Great Plains of Oklahoma and Kansas, USA. A LES made for a highly documented and analyzed study case of IHOP (Couvreux et al, 2007) has been extended to cover the late afternoon.
- CASES (1997-1999): Cooperative Atmosphere-Surface Exchange Study, for the study of the nocturnal stable boundary layer. During CASES-99, the observations started late in the day, except for one case that can be useful for our objective. During cases-97, radiosoundings were launched every 1h30 for 5 days in a row at three different sites, with surface fluxes at 5 different

locations, and also sodar and UHF wind profiles.

- Cabauw and Lindenberg are two observatories with an equipped tower, remote sensing and surface measurements with long term observations which allowed simulations testbed. Those are appropriate to consider, both for some selected cases (that were chosen before for LES and mesoscale models intercomparisons), and from a statistic point of view, in order to draw a climatology of the afternoon transition.

2) Field Campaign

No experiment ever gathered the airplanes, wind profilers, and Doppler lidar for the study of the late afternoon CBL processes. Those three platforms are very complementary and important for boundary layer issues, like entrainment. We plan to put them together along with surface measurements, tethered and radiosounding balloons, and with LES and mesoscale modelling in a parallel approach, to have the most complete understanding of the CBL processes, with measurements all day and over night for some instruments, and an extensive density of measurements during the afternoon transition.
There is a strong need of field campaigns for the study of the afternoon transition decay, recently expressed in the 2009 European Meteorology Conference (Pardyjak, 2009). Both complex terrain or heterogeneous surface and 'ideal' flat and homogeneous terrain are necessary to consider.

We plan a one-month experiment in 2011 (June or July) at Lannemezan in France, and one in 2012 at Lindenberg in Germany, during the LITFASS experiment.
The two sites present heterogeneous surface, and rather complex terrain (especially Lannemezan, which is on a plateau close to the Pyrénées ridge).
Those sites will allow us to study the effect of the surface heterogeneity, and in Lannemezan, the interaction of the transition with plain-mountain breeze system setting can be studied.

Site for 2011

In 2011, the experiment will take place in the vicinity of the instrumented site of Lannemezan, France.

Close to Pyrenees mountains, the site is located on apPlateau, with divergent hills starting from it. The site is aligned with a main S-N oriented valley, which starts to the south (FIG. 13).
Heterogeneous surface: Prairies, grasslands, crops, forests (FIG. 14).

This site has a 65 m instrumented tower, with 5 levels, including 3 levels with turbulence measurements (3 sonics, 1 rapid O3, one Licor, one KH-20).
Several other instruments at the site: UHF profiler, VHF profiler, standard meteorological station. Radiosoundings and tethered balloons + several ground meteorological stations which can be operated during field campaigns.
There are 12 standard Météo-France stations in a 50 km x 50 km square around the site (FIG. 13).
Toulouse aircraft facility and airport 100 km away, and there are several small airports closer to the site.

Instruments

Dynamics:

We wish to combine, at the same place and time, the in situ measurements made with balloons and airplanes with the remote sensing (LIDAR and RADAR) capability and with measurements of flux and mean variables at the surface.

The surface layer and entrainment zone could be extensively probed by one or two tethered balloons, with successive upward-downward round-trips through those layers, which are two key interfaces, with rapidly changing vertical structure during the afternoon transition.
The tethered balloons can also be fixed at a certain altitude, to make several punctual measurements of turbulence fluxes higher in the CBL than the tower can give.

The UAV is also a tool that could do that, but has not been used for this much yet. We would like to test its capability to probe thin interfaces.

Meanwhile, the airplanes will probe a larger horizontally extended area, with horizontal legs at different levels for the measurement of fluxes. The smaller the airplane, the closer to the surface.

The Doppler lidar can give the fine scale structure of one air velocity component, with smaller time interval and at higher spatial resolution. So turbulence statistics can be made on the wind.

The UHF profiler has been a very appropriate tool to give continuous vertical profiles of the mean wind and CBL depth evolution, -which are both indispensable for the analysis of the LIDAR and in situ turbulence measurements. It also gives estimates of the TKE dissipation rate profiles (and possibly humidity and temperature in certain conditions).

<u>Ground based in situ</u>

- Instrumented tower. 3 levels of turbulence, 5 of mean variables.
- Surface fluxes (10 m towers around the site), radiation, energy budget.
- Radiosounding balloons, PTUV probes (a system will allow frequent radiosoundings of the lower troposphere).
- Tethered balloon(s) with turbulence pod
- Smoke release

<u>Ground based remote sensing, to be operated at the tower site</u>

- Doppler lidar
- Aerosol lidar
- Raman lidar
- UHF wind profiler
- Camera to track the smoke released from the 65 m tower

<u>Airborne in situ</u>

As far as we can, we wish to have several airplanes flying at various levels at the same time. Sky arrows and UAVs are appropriate for flying low and slow, and they will probe the surface layer, miwed layer and PBL top in a complementary way.

Expected participating aircrafts:
- Ibimet sky arrow (Italy)
- Piper Aztec (France)
- UAV M2AV Carolo (Germany)

Main strategy: We will deploy 3 to 6 sites around the 65 m tower on different vegetation-type surfaces, and within around 5 km from the tower (FIG. 15).
Some sites will be aligned with the main wind flow in fair weather (northerly valley wind), and some along a transverse axis (and aligned with another common westerly flow). The first axis will be the path of a scintillometer system.

As much as possible, those sites will be equipped with a 10 m tower with turbulence measurements, a tethered balloon for sounding the lower troposphere, and a wind profiler like a sodar or UHF wind profiler.
The other instruments will be concentrated close to the 65 m tower of the main site, which has most of the infrastructure (lidars, smoke release experiment, radiosoundings…).

The UAVs will probe the low atmosphere in this 5 km radius area of concentrated measurements, and the Sky arrow and Piper will probe a larger area (40 km scale) (FIG. 16).

Modelling

By using DNS, LES and mesoscale models, we plan to combine ideal and real case simulations in order to better understand the key processes during afternoon transition: entrainment, land-surface interaction, clouds... Moreover, mixed-layer model can help us to interpret basic features during decay. Finally, some parameterizations will be suggested and included in large scale models. The validation of these parameterizations will be made by coupling the three types of simulations with observations, and further, by ensemble modeling, in the context of GABLS.

- First, DNS and LES of idealized transitions will be performed. In these simulations, the evolution of the CBL will be studied after a sudden or varying shut-off of surface fluxes. Varying the characteristics of the boundary layer (inversion strength, wind shear, surface characteristics …), will enable us to separate the main mechanisms which control the transition. Particular attention will be paid to the evolution of the characteristic length scales.
- A second step will be based on introducing in the LES the observed temperature, humidity and wind profiles as initial conditions, and the evolution of surface fluxes of some selected episodes. This study will lean on sounding data or LIDAR/wind profiler observations at the end of the day for these episodes.
- Mesoscale models will also be used to reproduce the atmospheric characteristics of the selected episodes and derive the importance of mesoscale circulations.
- Those numerical studies should enable us to improve the parameterization of

entrainment heat fluxes, for a better forecast of the evolution of the main variables during this phase of the diurnal cycle.

- The Météo-France forecast models will be used during the field experiment, and evaluated in the same occasion.

Models used, among others, will be the Dutch Atmospheric Large Eddy Simulation (DALES), Meso-NH, the Met-Office large eddy model (LEM), and Météo-France forecast models (ARPEGE, ALADIN, AROME).

Part VI: List of participants

LASTNAME	FIRSTNAME	LAB	CITY	COUNTRY
Angevine	Wayne	NOAA	Boulder, CO	US
Augustin	Patrick	LPCA	Dunkerque	FR
Bange	Jens	Technische Universitaet	Braunschweig	GE
Beare	Robert	University of Exeter	Exeter	GB
Beyrich	Frank	Meteorologisches Observatorium Lindenberg	Lindenberg	GE
Bousquet	Olivier	CNRM/LISA	Toulouse	FR
Campistron	Bernard	LA	Toulouse	FR
Couvreux	Fleur	CNRM	Toulouse	FR
Cuesta	Juan	LMD, Ecole Polytechnique	Paris	FR
Dabas	Alain	CNRM/LISA	Toulouse	FR
Delbarre	Hervé	LPCA	Dunkerque	FR
Durand	Pierre	LA	Toulouse	FR
Fernando	Harindra J.	University of Notre Dame	Notre Dame, IN	US
Gibert	Fabien	LMD, Ecole Polytechnique	Paris	FR
Gioli	Beniamino	IBIMET	Florence	IT
Guichard	Françoise	CNRM	Toulouse	FR
Hartogensis	Oscar	Meteorology and Air Quality Section, Wageningen University	Wageningen	NE
Heusinkveld	Bert	Meteorology and Air Quality Section, Wageningen University	Wageningen	NE
Holtslag	Bert	Meteorology and Air Quality Section, Wageningen University	Wageningen	NE
Jonker	Harm	Delft University of Technology	Delft	NE
Legain	Dominique	CNRM/4M	Toulouse	FR
Lemone	Margaret	NCAR	Boulder, CO	US
Lenschow	Donald H.	NCAR	Boulder, CO	US
Lohou	Fabienne	LA	Toulouse	FR
Lothon	Marie	LA	Toulouse	FR
Mahrt	Larry	COAS, College of Oceanic and Atmospheric Sciences	Corvallis, OR	US
Martin	Sabrina	Technische Universitaet	Braunschweig	GE
Moene	Arnold	Meteorology and Air Quality Section, Wageningen University	Wageningen	NE
Pardyjak	Eric	University of Utah	Salt Lake City, UT	US
Parlange	Mark	EPFL	Lausanne, Switzerland	SW
Pino	David	Technical University of Catalonia	Barcelona	SP
Saïd	Fredérique	LA	Toulouse	FR
Sorbjan	Zbigniew	Marquette University, Milwaukee	Milwaukee, WI	US
Steenveld	Gert-Jan	Meteorology and Air Quality Section, Wageningen University	Wageningen	NE
Vilà-Guerau de A.	Jordi	Meteorology and Air Quality Section, Wageningen University	Wageningen	NE

Participating groups

Ibimet (B. Gioli) : Sky arrow
> EUFAR

WUR (J. Vilà-Guerau de Arellano, B. Holtslag,O. Hartogensis, B. Heusinkveld, A. Moene, G.J. Steenveld) **:**
LES + surface/soil measurements, scintillometers, LW radiation
> part in LEFE, PHC proposal

Delft University (H.Jonker, E. de Beus, S. Boing, R. van Driel, S. de Roode, P. Siebesma): smoke exp. , LES
> part in LEFE

UPC (D. Pino, E. Blay): LES & MNH - integrated action

Univ. Tübingen (J. Bange, S. Maritn, Y. Breitenbach): UAV
> DFG

Univ. Utah (E. Pardyjak, M. Parlange, H. J. S. Fernando):
tethered balloon, surface stations (turbulence, CO_2 fluxes), Ra lidar
> NSF

Also D. H. Lenschow (USA, CO, NCAR), Wayne Angevine (USA, CO, NOAA), Z. Sorbjan (USA, Marquette University), B. Beare (UK, Univ. Exeter), L. Mahrt (USA, Oregon Univ.),… in modelling and/or expertises

Météo-France:
(1) Ground observation (D. Legain, A. Dabas, O. Bousquet)

- Remote sensing (Doppler lidar + UHF)
- Tethered balloon (with turbulence probe)
- Frequent radiosounding
- Surface turbulence measurements
- Scintillometer

(2) Modelling forecast (E. Bazile, Y. Seity)
(3) LES (F. Couvreux, F. Guichard)

LMD (Fabien Gibert): lidar (CO_2, Doppler) + sonic
> BQR + IPSL

LPCA (H. Delbarre, M. Fourmentin, P. Augustin, K. Deboudt, P. Flament :
aerosol lidar + sodar + aerosol UAV
> BQR

LA (M. Lothon, F. Lohou, P. Durand, F. Saïd) **:**

- Site with UHF, Tethered balloon, std radiosounding, tower
- Aicraft expertise
- LES

> BQR, LEFE, EUFAR

Part VII: References

Acevedo, O. and D. R. Fitzjarrald, 2001: The early evening surface-layer transition: Temporal and spatial variability. *J. Atmos. Sci.*, **58**, 2650-2667.

Amiridis, V., D. Melas, D. S. Balis, A. Papayannis, D. Founda, E. Katragkou, E. Glannakaki, R.E. Marmouri, E. Gerasopoulos and C. Zerefos, 2007: Aerosol lidar observations and model calculations of the planetary boundary layer evolution over Greece, during the March 2006 solar eclipse. *Atmos. Chem. Phys*., **7**, 6181-6189.

Anfossi, D., G. Schayes, G. Degrazia, and A. Goulart, 2004: Atmospheric turbulence decay during the solar total eclipse of 11 August 1999, *Boundary-Layer Meteor*., **111**, 301-311.

Angevine, W. M., M. K. Trainer, S. A. McKeen, and C. M. Berkowitz, 1996: Mesoscale meteorology of the New England coast, Gulf of Maine, and Nova Scotia: Overview. *J. Geophys. Res*., **101,** 28893-28901.

Angevine W. M., A. W. Grimsdell, L. M. Hartten, and A. C. Delany, 1998: The Flatland boundary layer experiments. *Bull. Amer. Meteorol. Soc.*, **79,** 419-431

Angevine W. M., H. K. Baltin, and F. C. Bosveld, 2001: Observations of the morning transition of the convective boundary layer. *Boundary-Layer Meteor*., **101**, 209–227.

Angevine, W. M., C. J. Senff, A. B. White, E. J. Williams, J. Koermer, S. T. K. Miller, R. Talbot, P. E. Johnston, S. A. McKeen, and T. Downs, 2004: Coastal boundary layer influence on pollutant transport in New England. *J. Appl. Meteor.*, **43,** 1425-1437.

Angevine, W. M., M. Tjernstrom, and M. Zagar, 2006: Modeling of the coastal boundary layer and pollutant transport in New England. *J. App. Meteorol. Climatol.*, **45,** 137-154.

Angevine, W. M., 2008: Transitional, entraining, cloudy, and coastal boundary layers, *Acta Geophys*., **56**, 2-20.

Bange, J., T. Spieß, and A. van den Kroonenberg, 2007: Characteristics of the early-morning shallow convective boundary layer from helipod flights during STINHO-2. *Theor. Appl. Climato*l. **90**, 113-126.

Banta, R. M., C. J. Senff, A. B. White, M. Trainer, R. T. McNider, R. J. Valente, S. D. Mayor, R. J. Alvarez, R. M. Hardesty, D. D. Parrish, and F. C. Fehsenfeld, 1998: Daytime buildup and nighttime transport of urban ozone in the boundary layer during a stagnation episode. *J. Geophys. Res.*, **103,** 22,519-22,544.

Baas, P., F. C. Bosveld, H. K. Baltink, and A. A. M. Holtslag, 2009: A climatology of nocturnal low-level jets at Cabauw. *J. Appl. Meteorol. Climatol.*, **48**, 1627-1642.

Banta, R. M., Y. L. Pichugina and W. A. Brewer, 2006: Turbulent velocity-variance profiles in the stable boundary-layer generated by a nocturnal low-level jet. *J. Atmos. Sci.*, **63,** 2700-2719.

Basu, S., A. A. M. Holtslag, B. J. H. van de Wiel, A. F. Moene, and G. -J. Steeneveld, 2008: An inconvenient "truth" about using sensible heat flux as a surface boundary condition in models under stably stratified regimes. *Acta Geophysica*, **56,** 88-99.

Batchelor, G. K., 1960: The Theory of Homogeneous Turbulence*. Cambridge University Press, London.*

Batchelor, G. K., 1967: An Introduction to Fluid Mechanics*. Cambridge University Press, London.*

Beare, R. J., J. M. Edwards, and A. J. Lapworth, 2006a: Simulation of the observed evening transition and nocturnal boundary layers: large-eddy modelling. Q*. J. R. Meteorol. Soc.* **132**, 81-99.

Beare, R. J., M. Macvean, A. A. M. Holtslag, J. Cuxart, I. Esau, J. C. Golaz, M. A. Jiménez, M. F. Khairoutdinov, B. Kosovic, D. C. Lewellen, T. S. Lund, J. K. Lundquist, A. McCabe, A. Moene, Y. Noh, S. Raasch, and P. P. Sullivan, 2006b: An intercomparison of large-eddy simulations of the stable boundary-layer, *Boundary-Layer Meteor*., **118**, 247-272.

Beare, R. J., 2007: Boundary layer mechanisms in extratropical cyclones. *Q. J. R. Meteorol. Soc*., **133**, 503-515.

Beare, R. J., 2008: The role of shear in the morning transition boundary layer. *Boundary-Layer Meteor.*, **129**, 395- 410.

Berkowitz, C. M., J. D. Fast, S. R. Springston, R. J. Larsen, C. W. Spicer, P. V. Doskey, J. M. Hubbe, and R. Plastridge, 1998: Formation mechanisms and chemical characteristics of elevated photochemical layers over the northeast United States. *J. Geophys. Res.*, **103,** 10,631-10,647.

Bernard-Trottolo S., B. Campistron, A. Druilhet, F. Lohou, F. Saïd, 2003 : TRAC98: Detection of coherent structures in a convective boundary layer using airborne measurements. *Bound.-Layer Meteor.,* **111,** 181-224.

Betts A., 2005: Land-surface, boundary layer, and cloud-field coupling over the southwestern Amazon in ERA-40, *J. Geophys. Res.*, 110, D14108, doi: 10.1029/2004JD005702.

Beyrich, F and D. A. M. Engelbart, 2008: Ten years of operational boundary-layer measurements at the Richard-Assmann Observatory Lindenberg: The role of remote sensing. 14th International Symposium for the Advancement of Boundary Layer Remote Sensing, 23-25 June 2008, Tech. Univ. of Denmark, Copenhagen, Denmark.

Biferale, L., G. Boffeta, A. Celani, A. Lanotte, F. Toschi, and M. Vergassola, M., 2003: The decay of homogeneous anisotropic turbulence*, Phys. Fluid.*, **8**, 2105-2112.

Brazel, A. J., H. J. S. Fernando, J. C. R. Hunt, N. Selover, B. C. Hedquist, and E. Pardyjak, 2005: Evening Transition Observations in Phoenix, Arizona. *J. Appl. Meteor.,* **44,** 99-112.

Brown R. A., R. T. Cederwall, A. Chlond, P. G. Duynkerke, J. C. Golaz, M. Khairoutdinov, D. C. Lewellen, A. P. Lock, M. K. McVean, C. H. Moeng, R. A. J. Neggers, A. P. Siebesma and B. Stevens, 2002: Large-eddy simulation of the diurnal cycle of a shallow cumulus convection overland, *Q. J. R. Meteorol. Soc.,* **128**, 1075-1093.

Casso-Tarralba, P. J., J. Vilà-Guerau de A., F. Bosveld. M. R. Soler, A. Vermeulen, C. Werner and E. Moors, 2008: Diurnal and vertical variability of the sensible heat and carbon dioxide budgets in the atmospheric surface layer. *J. Geophys. Res.*, **113,** D12119, doi:10.1029/2007JD009583.

Caughey, S. J., and J. C. Kaimal, 1977: Vertical Heat Flux in the Convective Boundary Layer*, Q. J. R. Meteorol. Soc.,* **103**, 811-815.

Cohn S. A., S. D. Mayor, C. J. Grund, T. M. Weckwerth, C. Senff, 1998: The lidars in flat terrain (LIFT) experiment, *Bull. Amer. Meteorol. Soc*., **79**, 1329-1343.

Cole, G. S., and H. J. S. Fernando, 1993: Some aspects of the decay of convective turbulence, *Fluid Dyn. Res.*, **23**, 161-176.

Couvreux F., F. Guichard, J. L. Redelsperger, C. Kiemle, V. Masson, J. P. Lafore, and C. Flamant, 2005a: Water-vapour variability within a convective boundary-layer assessed by large-eddy simulations and IHOP_2002 observations. Q. J. R. Meteorol. Soc., **113,** 2665-2693.

Couvreux, F., 2005b: phD, 'Water vapour variability in the convective boundary layer', *University Paul Sabatier, Toulouse, France.*

Couvreux F., F. Guichard, V. Masson, J. L. Redelsperger, 2007: Negative water vapour skewness and dry tongues in the convective boundary layer: observations and large-eddy simulation budget analysis. *Boundary-Layer Meteor*., **123**, 269-294.

Cuijpers J. W. M. and A. A. M. Holtslag, 1998: Impact of skewness and nonlocal effects on scalar and buoyancy fluxes in convective boundary layers *J. Atmos. Sci*., **55**, 151-162.

Cuxart, J., C. Yagüe, G. Morales, E. Terradellas, J. Orbe, J. Calvo, A. Fernández, M. R. Soler, C. Infante, P. Buenestado, A. Espinalt, H. E. Joergensen, J. M. Rees, J. Vilà, J. M. Redondo, I. R. Cantalapiedra, and 21 L. Conangla, 2000: Stable Atmospheric Boundary-Layer Experiment in Spain (SABLES 98): A Report. *Boundary-Layer Meteor*. **96**, 337-370.

Deardorff, J. W., 1972: Numerical investigation of neutral and unstable planetary boundary layers. *J. Atmos. Sci.*, **29**, 91-115.

Derbyshire, S. H., 1990: Nieuwstadt's stable boundary layer revisited. *Q. J. R. Meteorol. Soc.*, **116**, 127-158.

Dolas, P. M., R. Ramchandran, K. Sen Gupta, S. M. Patil, and P. N. Jadhav, 2002: Atmospheric surface-layer processes during the total solar eclipse of 11 august 1999. *Bound.-Layer Meteor.,* **104,** 445–461.

Edwards, R. J., Beare, and A. J. Lapworth, 2006: Simulation of the observed evening transition and nocturnal boundary layers: single column modelling*. Q. J. R. Meteorol. Soc.*, **132**, 61-80.

Fernando, H. J. S., M. Princevac, E. Pardyjak, and A. Data, 2004: The decay of convective turbulence during evening transition period. Proc. of the 11th Conference on Mountain Meteorology and MAP Meeting, Bartlett (NH), USA.

Fitzjarrald, D. R., J. M. Freedman., M. J. Czikowskyand, R. K. Sakai, O. C. Acevedo and O. L. L. Moraes, 2004: Momentum and scalar transport during the decay of CBL turbulence. Proc. of the 16th Symposium on Boundary Layers and Turbulence, Portland (ME), USA.

Founda, D., D. Melas, S. Lykoudis, I. Lisaridis, E. Gerasopoulos, G. Kouvarakis, M. Petrakis, and C. Zerefos, 2007: The effect of the total solar eclipse of 29 March 2006 on meteorological variables in Greece. *Atmos. Chem. Phys.*, **7**, 5543–5553.

Garratt, J. R., 1992: *The atmospheric boundary layer*. Cambridge University Press, 316 pp.

George, W. K., 1992: The decay of homogeneous isotropic turbulence, *Phys. Fluid.*, **4**, 1492-1509.

Girard-Ardhuin F., B. Benech, B. Campistron, J. Dessens, and S. Jacoby-Koaly, 2003: Remote sensing and surface observations of the response of the atmospheric boundary layer to a solar eclipse. *Boundary-Layer Meteor.*, **106**, 93-115.

Gorska M., J. Vilà-Guerau de Arellano, M. A. LeMone, and C. C. van Heerwaarden, 2008: Mean and flux horizontal variability of virtual potential temperature, moisture, and carbon dioxide: aircraft observations and LES study. *Mon. Weather Rev.*, **136**, 4435-4451.

Goulart A., G. Degrazia, U. Rizza, and D. Anfossi, 2003: A theoretical model for the study of convective turbulence decay and comparison with Large-Eddy Simulation data, *Boundary-Layer Meteor.*, **107**, 143-155.

Goulart A. G., M. T. Vilhena, G. A. Degrazia, and D. T. Flores, 2007: Vertical, lateral and longitudinal eddy diffusivities for a decaying turbulence in the convective boundary layer. *Ecol. Model.*, **204**, 516-522.

Grant, A. L. M., 1997: An observational study of the evening transition boundary layer. *Q. J. R. Meteorol. Soc.*, **123,** 657-677.

Grimsdell, A. W. and W. M. Angevine, 2002: Observations of the afternoon transition of the convective- boundary layer, *J. Appl. Meteorol.,* **41**, 3-11.

Hess, G. D., B. B. Hicks and T. Yamada, 1981: The impact of the Wangara experiment. *Boundary-Layer Meteor.,* **20**, 135-174.

Hurley, P., and A. Luhar, 2009: Modelling the meteorology at the Cabauw tower for 2005. *Bounary.-Layer Meteor.*, **132,** 43-57.

Jacobson, M. Z., 2000: Fundamentals of Atmospheric Modeling. *Cambridge University Press, 666 pp.*

Jonker, H. J. J., P. G. Duynkerke and J. W. M. Cuijpers, 1999: Mesoscale fluctuations in scalars generated by boundary layer convection*. J. Atmos. Sci.,* **56**, 801-811.

Jonker, H. J. J., T. H, 2008: Laboratory experiments of entrainment in dry convective boundary layers, , Proceedings of the 18th symposium on Boundary layer and Turbulence, Stockholm, Sweden, 9-13 June.

Kaimal, J. C., J. C. Wyngaard, D. A. Haugen, O. R. Coté, Y. Izumi, S. J. Caughey and C. J. Readings, 1976: Turbulence Structure in the Convective Boundary Layer*, J. Atmos Sci.*, **33**, 2152-2169.

Kaimal, J. C., and J. C. Wyngaard, 1990: The Kansas and Minnesota experiments. *Boundary-Layer Meteor*., **50**, 31-47.

Lapworth, A., 2006: The morning transition of the nocturnal boundary layer. *Boundary-Layer Meteor*., **119**, 501-526.

LeMone, M. A., L. F. Tarleton, and G. M. Barnes, 1988: Perturbation pressure at the cloud base of cumulus clouds in low shear. *Mon. Wea. Rev.*, **116**, 2062-2068.

LeMone M. A. and R. Grossman, 1999: Evolution of potential temperature and moisture during the morning: CASES-97. Proceedings, 13th Symposium on Boundary-Layers and Turbulence,500-5003 and Proceedings, 14th Conf. on Hydrology, 446-449, Dallas, TX, 10-15 Jan.

LeMone M. A., R. Grossman R. T. McMillen, K. N. Liou, S. Oh, S. McKeen, W. Angevine, K. Ikeda and F. Chen, 2002: CASES-97: late-morning warming and moistening of the boundary-layer over the Walnut River watershed. *Boundary-Layer Meteorol.,* **104**, 1-52.

Lenschow, D. H., 1974: Model of the height variation of the turbulence kinetic energy budget in the unstable planetary boundary layer, *J. Atmos. Sci.,* **31**, 465-474.

Lenschow, D. H., and Stankov,1986: Length scales in the convective boundary layer. *J. Atmos. Sci.*, **43**, 1198-1209.

Lilly, D. K., 1967: The representation of small-scale turbulence in numerical simulation experiments. Proc. of IBM scientific computing symposium on environmental sciences, Yorktown heights, USA.

Lothon M., D. H. Lenschow, and S. D. Mayor, 2006: Coherence and scale of vertical velocity in the convective boundary layer from a Doppler lidar. *Boundary-Layer Meteor.*, **132**, 205-226.

Lothon M., D. H. Lenschow, and S. D. Mayor, 2009: Doppler lidar measurements of vertical velocity spectra in the convective planetary boundary layer. *Boundary-Layer Meteor.*, **132**, 205-226.

Mahrt, L. and D. H. Lenschow, 1976: Growth dynamics of the convectively mixed layer*, J. Atmos. Sci.,* **33**, 41–51.

Mahrt, L., 1981: The early evening boundary layer transition. *Q. J. R. Meteorol. Soc.*, **107,** 329-343.

Mahrt, L., J. Sun, W. Blumen, A. C. Delany, and S. Oncley, 1998: Nocturnal boundary-layer regimes. *Bound.-Layer Meteor.*, **88,** 255-278.

Mahrt, L., 1999: Stratified atmospheric boundary layers. *Boundary-Layer Meteorol.* **90**, 375–396.

Mauder, M., R. L. Desjardins, S. P. Oncley, and I. MacPherson, 2007: Atmospheric Response to a Partial Solar Eclipse over a Cotton Field in Central California. *J. Appl. Meteorol. Climatol.,* **46,** 1792-1803.

Moeng, C. -H., 1984: A large-eddy-simulation model for the study of planetary boundary-layer turbulence. *J. Atmos. Sci.,* **41**, 2052–2062.

Moeng, C. -H., and P. P. Sullivan, 1994: A comparison of shear and buoyancy driven planetary-boundary-layer flows*. J. Atmos. Sci.*, **51**, 999–1022.

Monin, A. S., and A. M. Yaglom, 1975: Statistical Fluid Dynamics, Vol. 2, J. L. Lumley, *Ed., The MIT Press, Massachusetts, 874 pp.*

Nicholls S. and Lemone P., 1980: The fair weather Boundary Layer in GATE: The relationship of ths subcloud Fluxes and Structure to the distribution and enhancement of cumulus clouds, *J. Atmos. Sci.,* **37**, 2051-2066.

Nieuwstadt, F. T. M., 1984: The turbulent structure of the stable, nocturnal boundary layer. *J. Atmos. Sci.* **41**, 2202–2216.

Nieuwstadt, F. T. M., and R. A. Brost, 1986: The decay of convective turbulence, *J. Atmos. Sci.,* **43**, 532- 546.

Nymphas, E. F., M. O. Adeniyi, M. A. Ayoola, and E. O. Oladiran, 2009: Micrometeorological measurements in Nigeria during the total solar eclipse of 29 March, 2006. *J. Atmos. Sol.-Terr. Phys.*, **71**, 1245-1253.

Pardyjak, E. R. and Fernando, H. J. S., 2009: The effect o fsurface type on the decay of turbulence in the surface layer during evening transition: 9th EMS Annual Meeting, 28 September – 2 October 2009, Toulouse, France.

Pergaud J., V. Masson, S. Maladrel and F. Couvreux, 2009 : A parameterization of dry thermals and shallow cumuli for mesoscale numerical weather prediction. *Boundary-layer Meterol.*, **132**, 83-106.

Pielke, R. A., 2002: Mesoscale Meteorological Modeling. *Academic Press*, 676 pp.

Pino, D., J. Vilà-Guerau de Arellano, H. J. J. Jonker, and A. Dosio, 2006a: Role of the shear and inversion strength during sunset turbulence over land: Characteristic length scales. *Boundary-Layer Meteor.*, **121**, 537-556.

Pope, S. B., 2000: Turbulent Flows, *Cambridge University Press, Cambridge, 771 pp.*

Poulos, G. S., W. Blumen, D. C. Fritts, J. K. Lundquist, J. Sun, S. P. Burns, C. Nappo, R. Banta, R. Newsom, J. Cuxart, E. Terradellas, B. Balsley, and M. Jensen, 2002: CASES-99: A comprehensive investigation of the stable nocturnal boundary layer. *Bull. Amer. Meteorol. Soc.,* **83,** 555-581**.**

Rio, C and F. Hourdin, 2008: A thermal plume model for the convective boundary layer: Representation of cumulus clouds*, J. Atmos. Sci.,* **65**, 407-425.

Shaw, W. J. and Barnard J. C., 2002: Scales of turbulence decay from observations and direct numerical simulations. Proc. of the 15th Symposium on Boundary Layers and Turbulence. Wageningen, The Netherlands.

Shi, X., R. T. McNider, M. P. Singh, D. E. England, M. J. Friedman, W. M. Lapenta, and W. B. Norris, 2005: On the behavior of the stable boundary layer and the role of initial conditions. *Pure Appl. Geophys.*, **162,** 1811-1829.

Siebesma, A. P., P. M. M. Soares and J. Texeira, 2007: A combined eddy-diffusivity mass-flux approach for the convective boundary layer. *J. Atmos. Sci*., **64**, 1230-1248.

Soares P. M. M. P. M. A. Miranda, A. P. Siebesma and , 2004: An eddy-diffusivity/mass-flux parameterization for dry and shallow cumulus convection. . *Q. J. R. Meteorol. Soc.*, **130,** 3365-3383.

Sorbjan, Z., 1996: Effects caused by varying the strength of the capping inversion based on a large eddy simulation model of the shear-free convective boundary layer, *J. Atmos. Sci.,* **53**, 2015-2024.

Sorbjan, Z., 1997: Decay of convective turbulence revisited, *Boundary-Layer Meteor.,* **82**, 501-515.

Sorbjan, Z., 2007: A numerical study of daily transitions in the convective boundary layer, *Boundary-Layer Meteor.,* **123**, 365-383.

Steeneveld, G.-J., B. J. H. Van de Weil, and A. A. M. Holtslag, 2006: Modelling the evolution of the atmospheric boundary layer coupled to the land surface for three contrasting nights in CASES-99. *J. Atmos. Sci.*, **63,** 920-935.

Stensrud, D. J., 2007: Parameterization Schemes. *Cambridge University Press,* 459 pp.

Stillinger, D. C., K. N. Helland, and C. W. Van Atta, 1983: Experiments on the transition of homogeneous turbulence to internal waves in a stratified fluid*, J. Fluid Mech.,* **131**, 91-122.

Stull R. B., 1976: Energetics of entrainment across a density interface*. J. Atmos. Sci.,* **33,** 1260-1267.

Stull R. B., 1988: An Introduction to Boundary Layer Meteorology, *Kluwer Academic Press, Dordrecht,* 688 pp.

Tennekes, H. and J. L. Lumley, 1973: A First Course in Turbulence. *The MIT Press, Cambridge, Massachussets.*

Tennekes, H. and A. G. M. Driedonks, 1981: Basic entrainment equations for the atmospheric boundary layer. *Boundary-Layer Meteorol.*, **20**, 515-531.

Touil, H., J. -P. Bertoglio, and L. Shao, 2002: The decay of turbulence in a bounded domain, *J. Turbul.,* **3**, 049.

Tjernstrom, M., B. B. Basley, G. Svensson, C. J. Nappo, 2009: The effects of critical layers on residual layer turbulence. *J. Atmos. Sci.,* **66**, 468-480.

Van Ulden, A. P. and J. Wieringa, 1996: Atmospheric boundary layer research at Cabauw. *Boundary-Layer Meteor.* **78**, 39–69.

Vilà-Guerau de Arellano, J., B. Gioli, F. Miglietta, H. J. J. Jonker, H. Klein Baltink, R.W. A. Hutjes and A. A. M. Holtslag, 2004a: The entrainment process of carbon dioxide in the atmospheric boundary layer. *J. Geophys. Res.,* **109**, D18110.

Vilà-Guerau de Arellano, J., A. Dosio, J. -F. Vinuesa, A. A. M. Holtslag, and S. Galmarini, 2004b: The dispersion of chemically reactive species in the atmospheric boundary layer. *Meteorol. Atmos. Phys.,* **87,** 23-28.

Vilà-Guerau de Arellano J., 2007: Role of nocturnal turbulence and advection in the formation of shallow cumulus over land. *Q. J. R. Meteorol. Soc.*, **133**, 1615-1627.

Vilà-Guerau de Arellano, J., J., K. van den Dries, D. Pino, 2009: On inferring isoprene emission surface flux from atmospheric boundary layer concentration measurements. *Atmos. Chem. Phys.,* **9,** 3629-3640.

Weckwerth T. M., D. B. Parsons, S. E. Koch, J. A. Moore, M. A. LeMone, B. B. Demoz, C. Flamant, B. Geerts, J. H. Wang, and W. F. Feltz, 2002: An overview of the International H2O Project (IHOP_2002) and some preliminary highlights, Bull. Amer. Meteorol. Soc., **85,** 253.

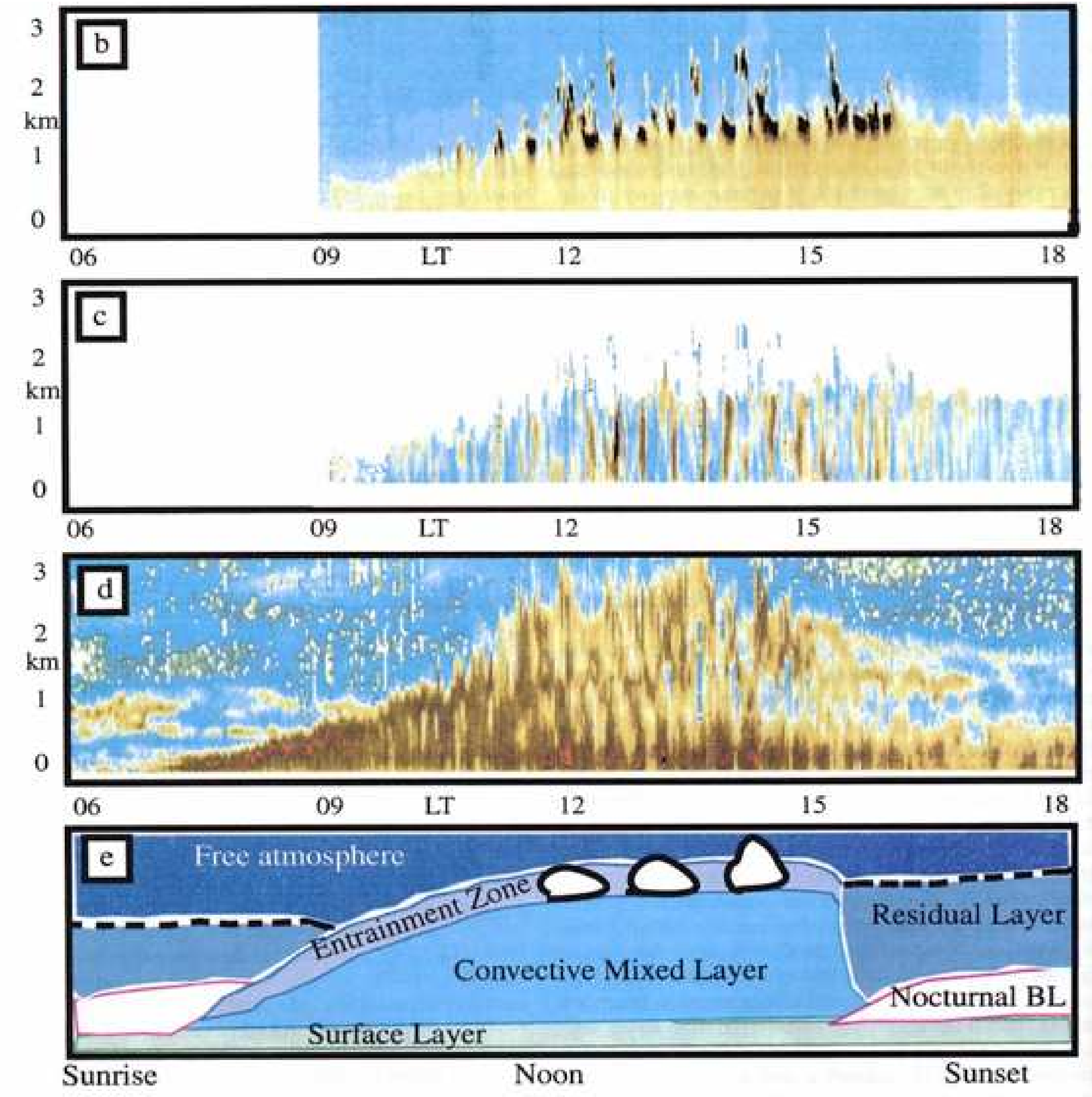

FIG. 1: After Cohn and Angevine, 2000. b)): lidar backscatter observed with the HRDL on 12 August 1996 during FLATLAND experiment. c)): Vertical velocity measured with the HRDL. d)): UHF wind profiler reflectivity for the same day. e)): Sketch by Stull, 1988.
Typical growth of the CBL and afternoon transition decay can be seen on lidar backscatter or UHF radar reflectivity, and on the thermal activity. In mid-day, each thermal was capped by a fair-weather cloud.

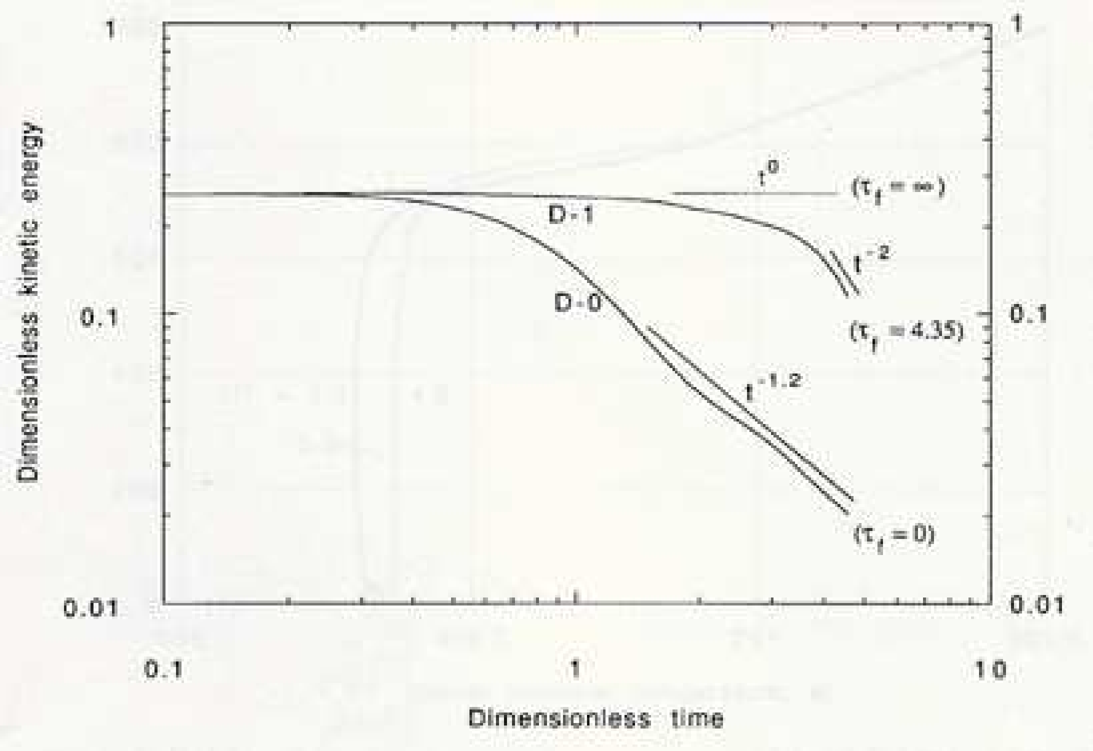

FIG. 2: Decay of the TKE during the afternoon, for various decrease of the surface heat flux, from sudden (τ_f=0) to infinitely slow (τ_∞=0). D-0 and D-1 are two different runs for two different time scales of the surface flux decrease. After Sorbjan (1997).

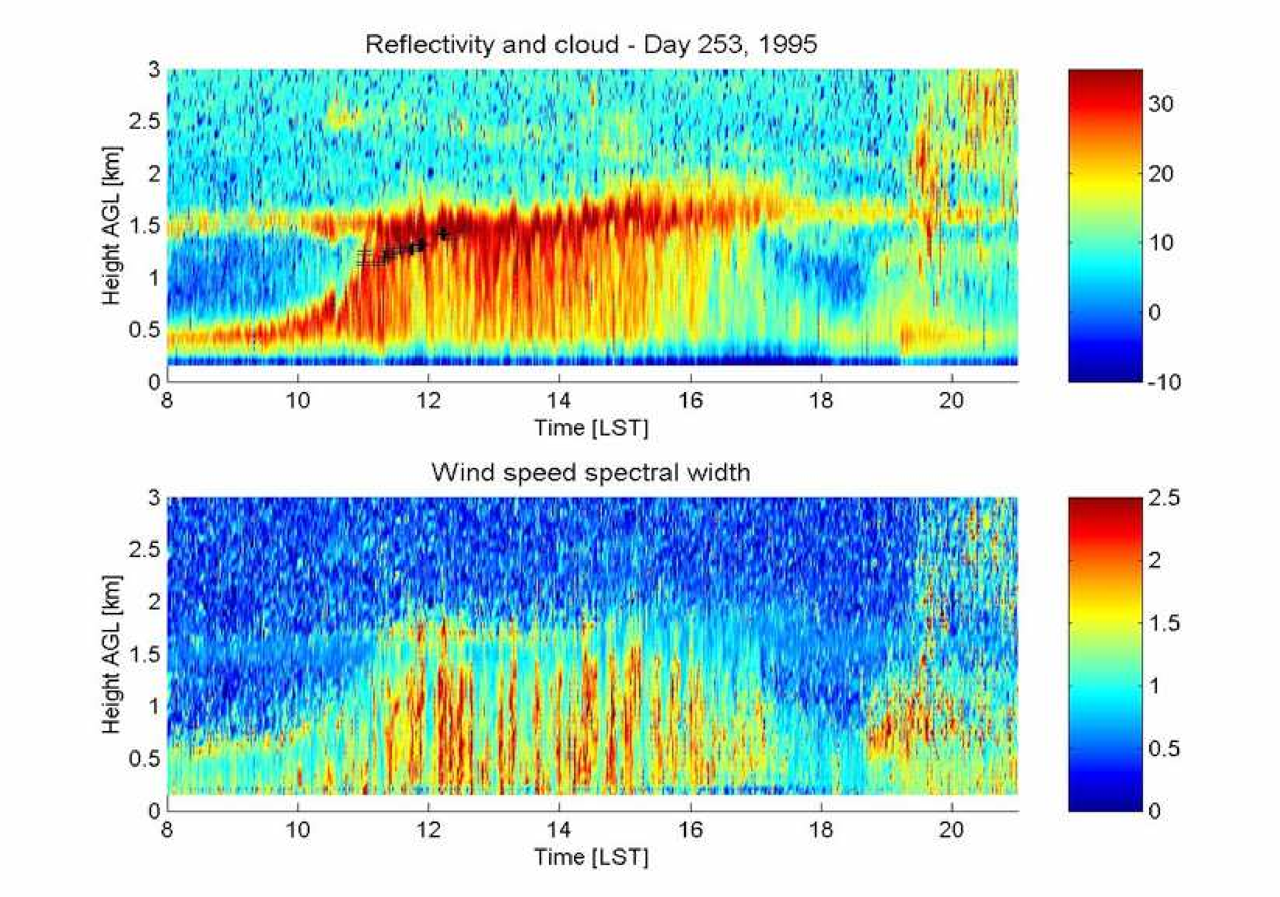

FIG. 3: After Angevine et al, 1998. Observation of a UHF wind profiler during FLATLAND: reflectivity at the top and Doppler spectral width at the bottom. This is a case with a strong inversion that did not move much all day (at around 1500 m). The TKE decay is seen with the spectral width, with a decreasing of the turbulent ML depth starting at about 15 LST until the night.

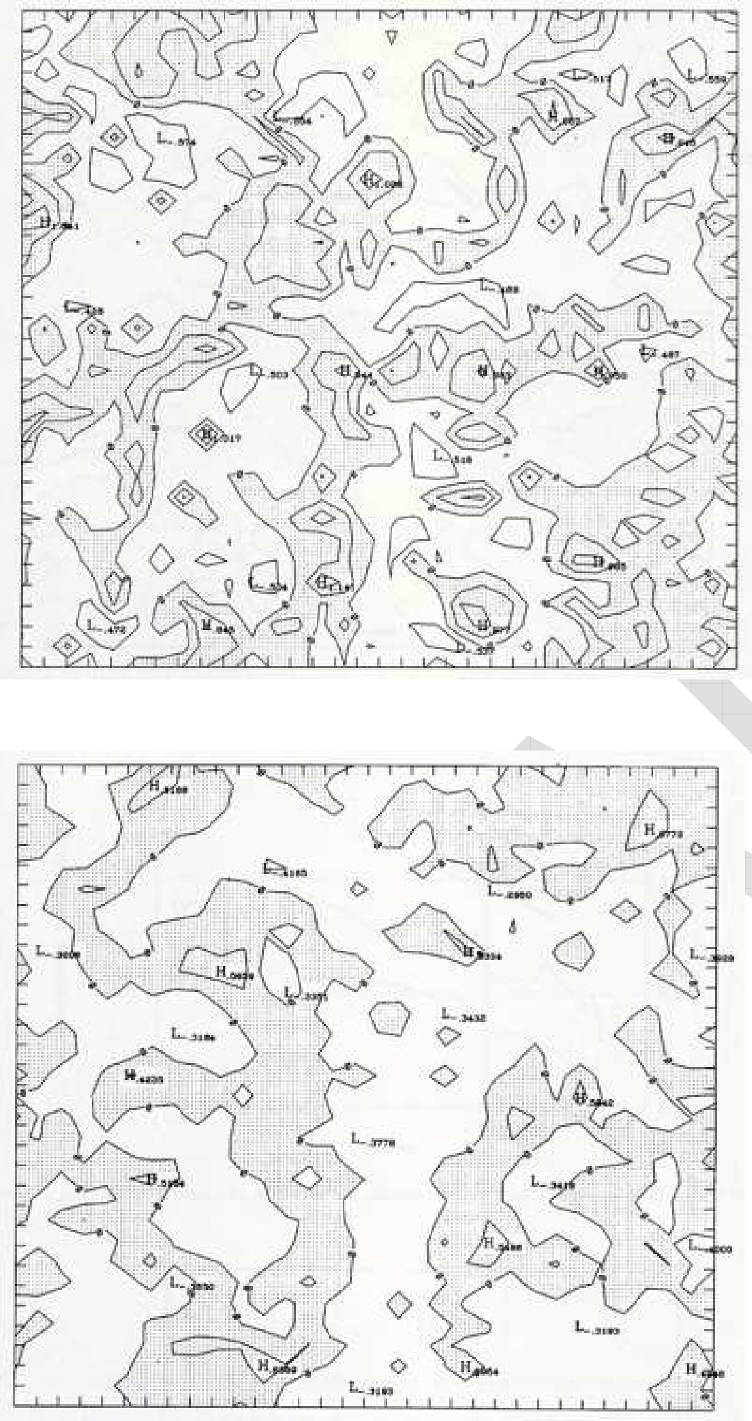

FIG. 4: After Sorbjan, 1997. Horizontal cross section of the vertical velocity in a LES at a height of 0.3 time the PBL depth, at two different times along the surface flux decrease: the characteristic scale of w increases from t=0 (top) to t=4.5 times the convective scale.

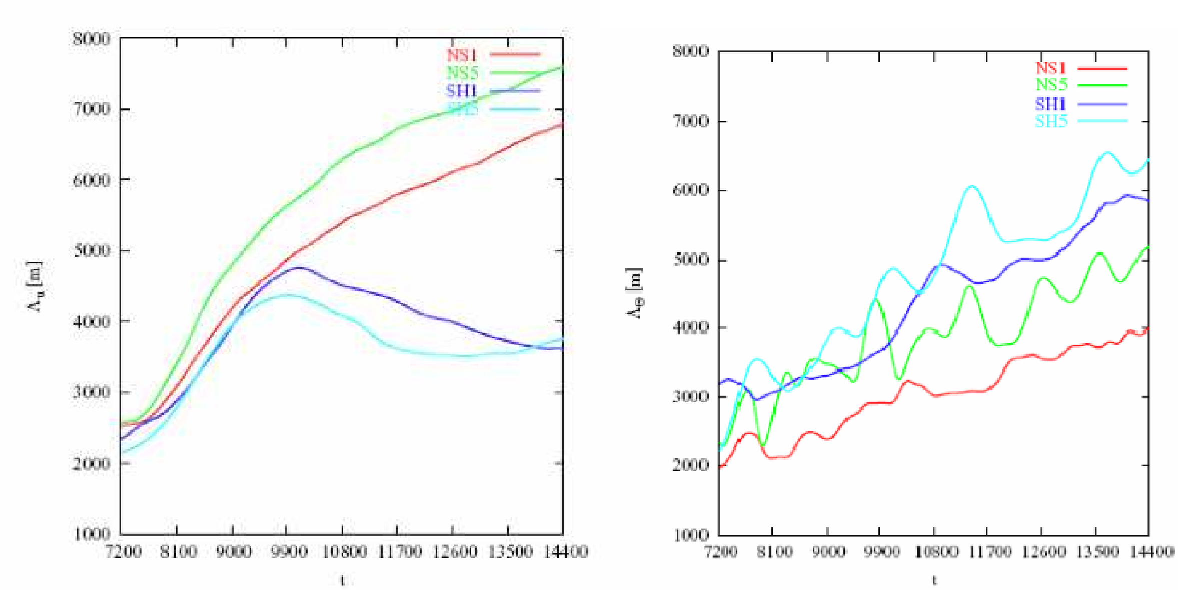

FIG. 5: After Pino et al, 2006. Evolution of the scale of the wind u-component (left) and potential temperature (right) during a LES. Temperature scale seems to grow indefinitely, while the u-scale stops to increase when shear exists. NS1 and NS5 have no-shear, SH1 and SH5 have a 1ms^{-1} wind speed jump across the PBL top. NS5 and SH5 have a larger potential temperature jump.

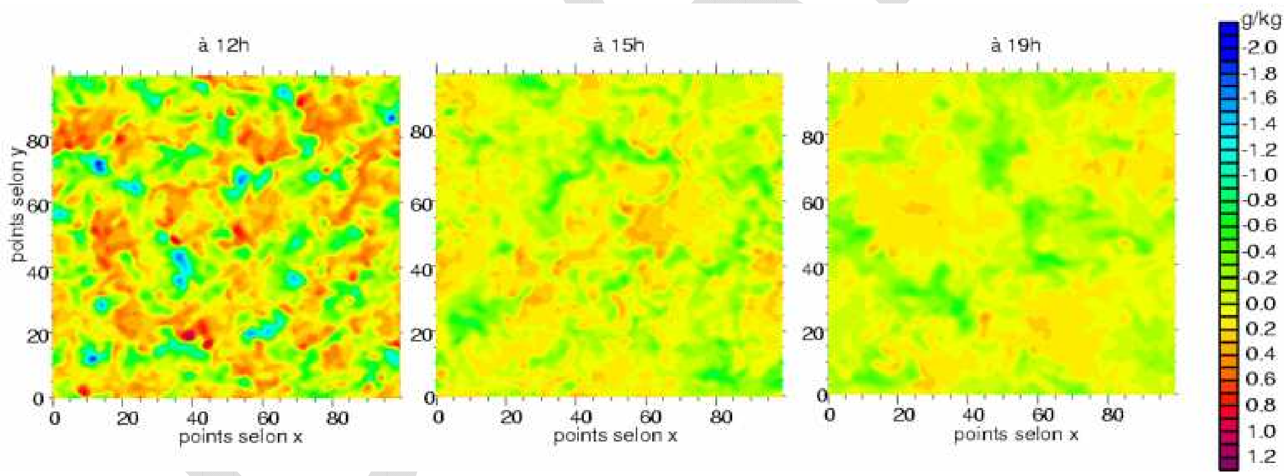

FIG. 6: LES of a IHOP case: Fluctuations of the water vapour at 12, 15 and 19 LT in the middle of the boundary layer. The characteristic scale of the water vapour mixing ratio increases as well during the late afternoon decay (the decrease in variance can be seen too). After Couvreux et al, 2005b.

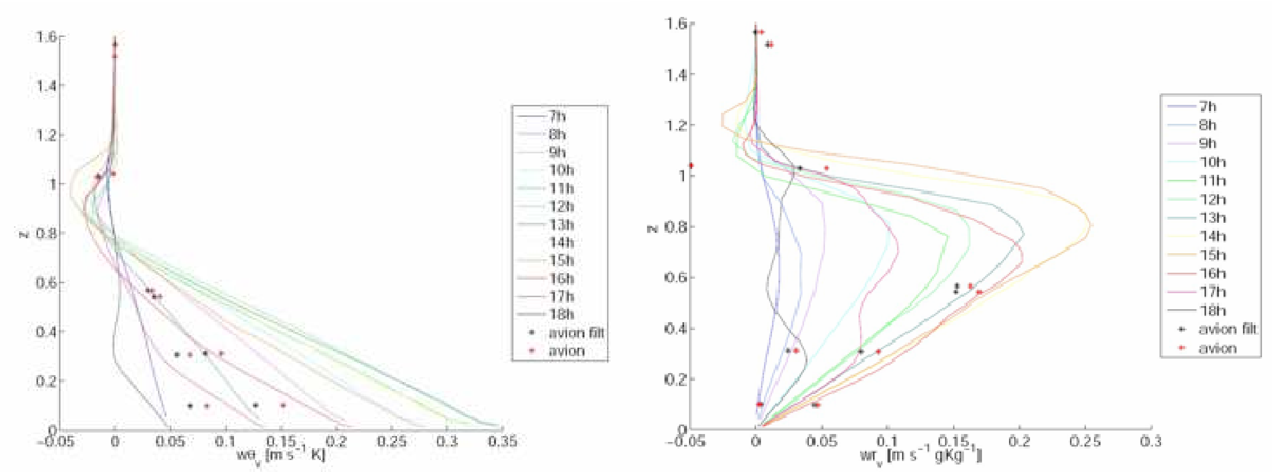

FIG. 7: Evolution of the vertical buoyancy (left) and latent heat (right) flux modeled with a LES for a case of AMMA (5 June 2006). Dots are the aircraft-measured fluxes. Courtesy of Canut and Couvreux, 2009.

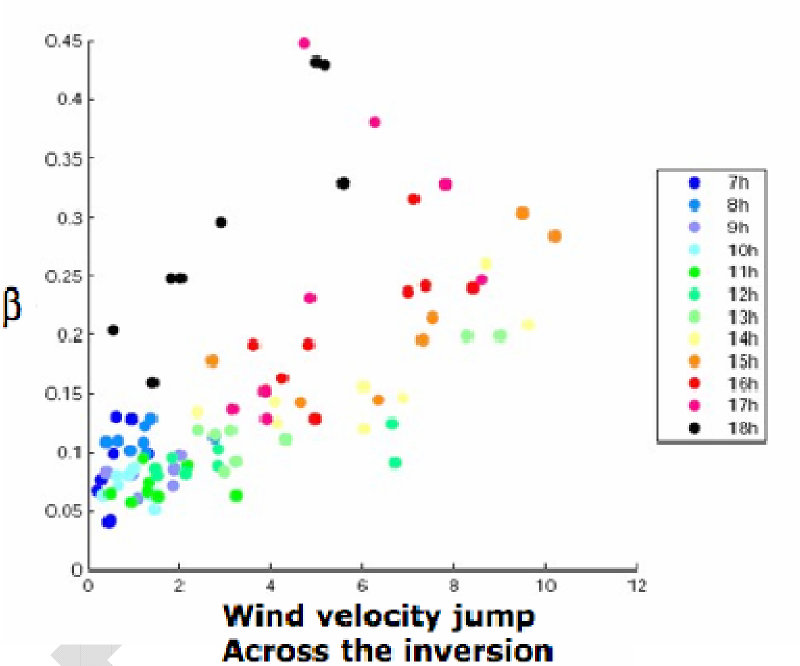

FIG. 8: The same LES as in Fig. 9 is considered, but with different initial conditions (wind and thermodynamic), which makes a set of various LES with constant surface flux but varying PBL vertical structure. Here the ratio of the buoyancy entrainment flux to surface flux for is plotted as a function of the shear at the top inversion (x-axis) and as a function of time (colors). Beta increases with increasing shear. Due to very weak fluxes at the end of the simulation, beta starts to be an unreliable parameter for the study and parameterization of entrainment. Courtesy of Canut and Couvreux, 2009.

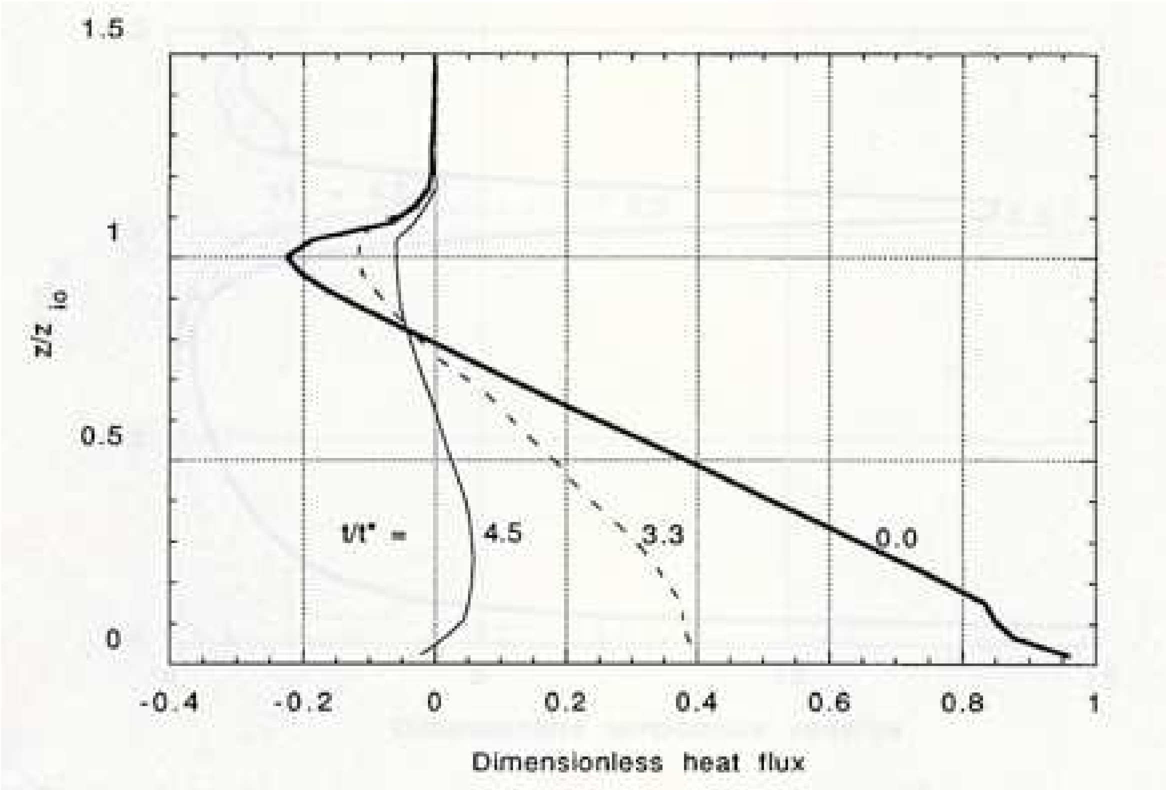

FIG. 9: After Sorbjan, 1997. The linear vertical profile of buoyancy flux turns to S-shape of weak fluxes in the late afternoon (t/t*=0 when the surface flux starts to decrease).

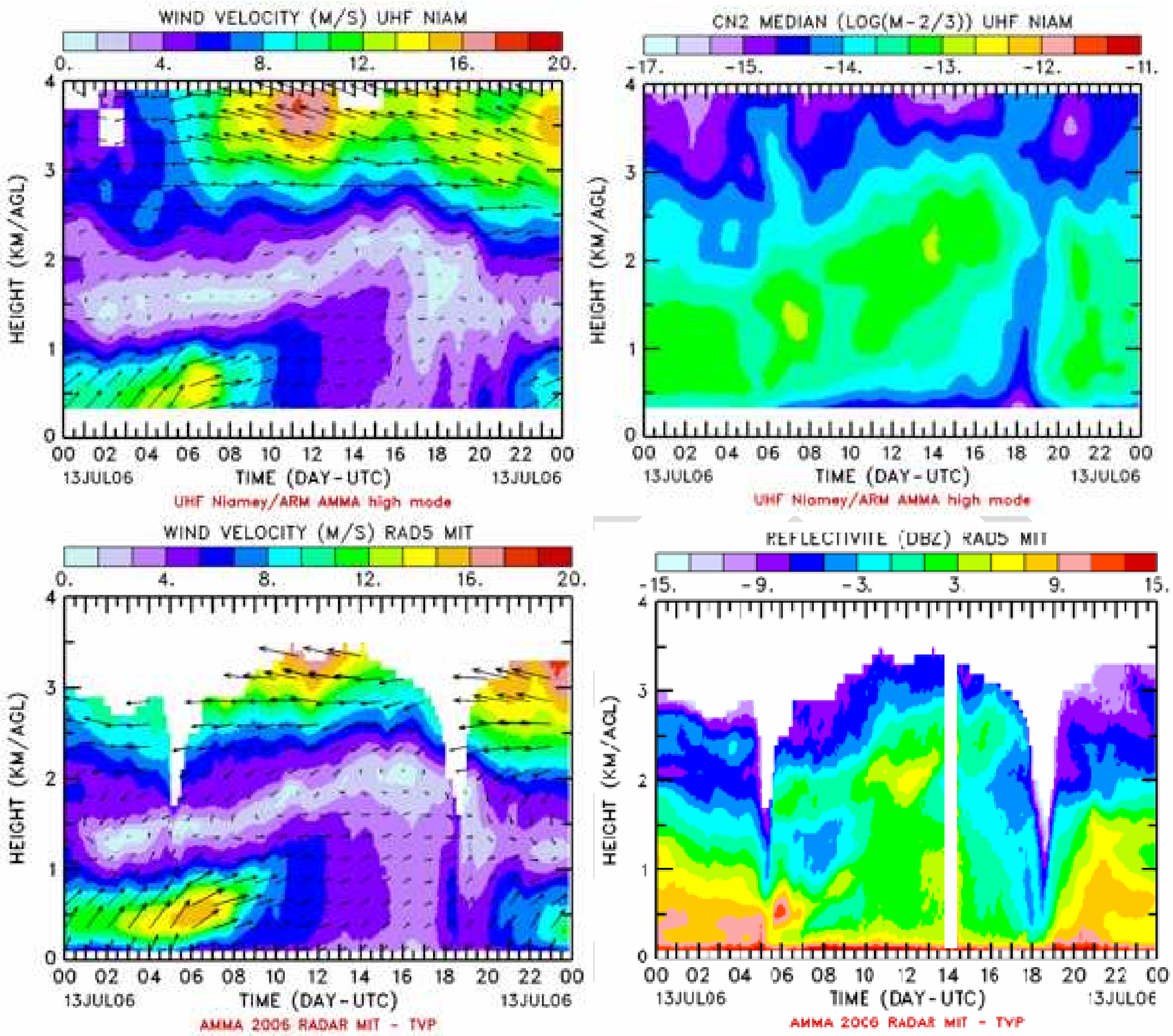

FIG. 10: Horizontal wind (left side) and reflectivity (right side) measured over Niamey, Niger, during the AMMA experiment by a UHF wind profiler (top) and a C-band radar (bottom). A nocturnal low-level jet sets up around 22 pm, until morning. A very sharp decrease of sensitivity on the MIT reflectivity was observed every day during the whole summer 2006 of the special observing period. The UHF reflectivity shows some decrease of reflectivity at that time too.

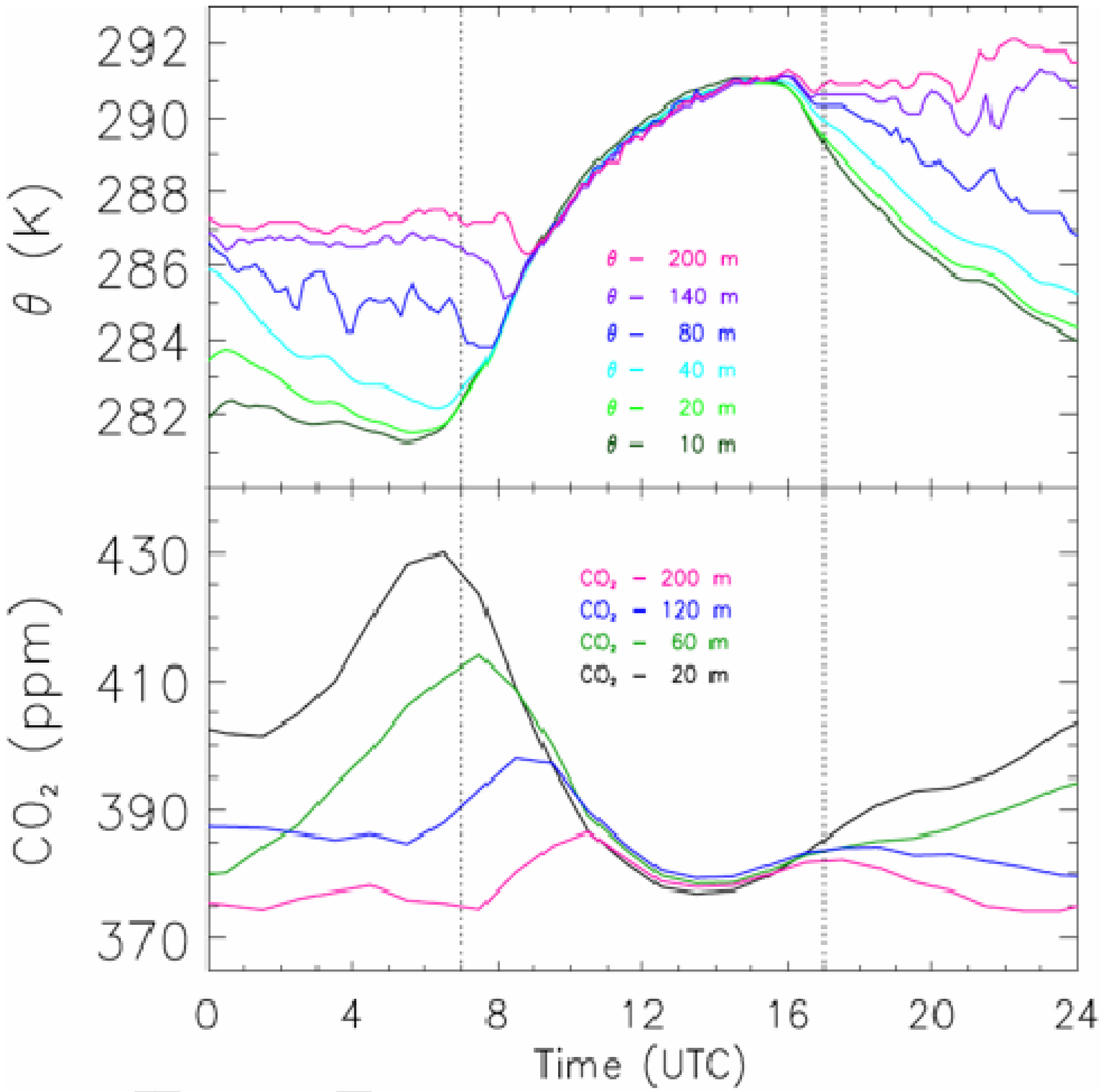

Fig. 11: After and complement of Casso-Terralba, 2008. Evolution of potential temperature and carbon dioxide concentration during an entire day, at several levels of the instrumented tower of Cabauw. The change from well-mixed scalars to stratification is very sharp.

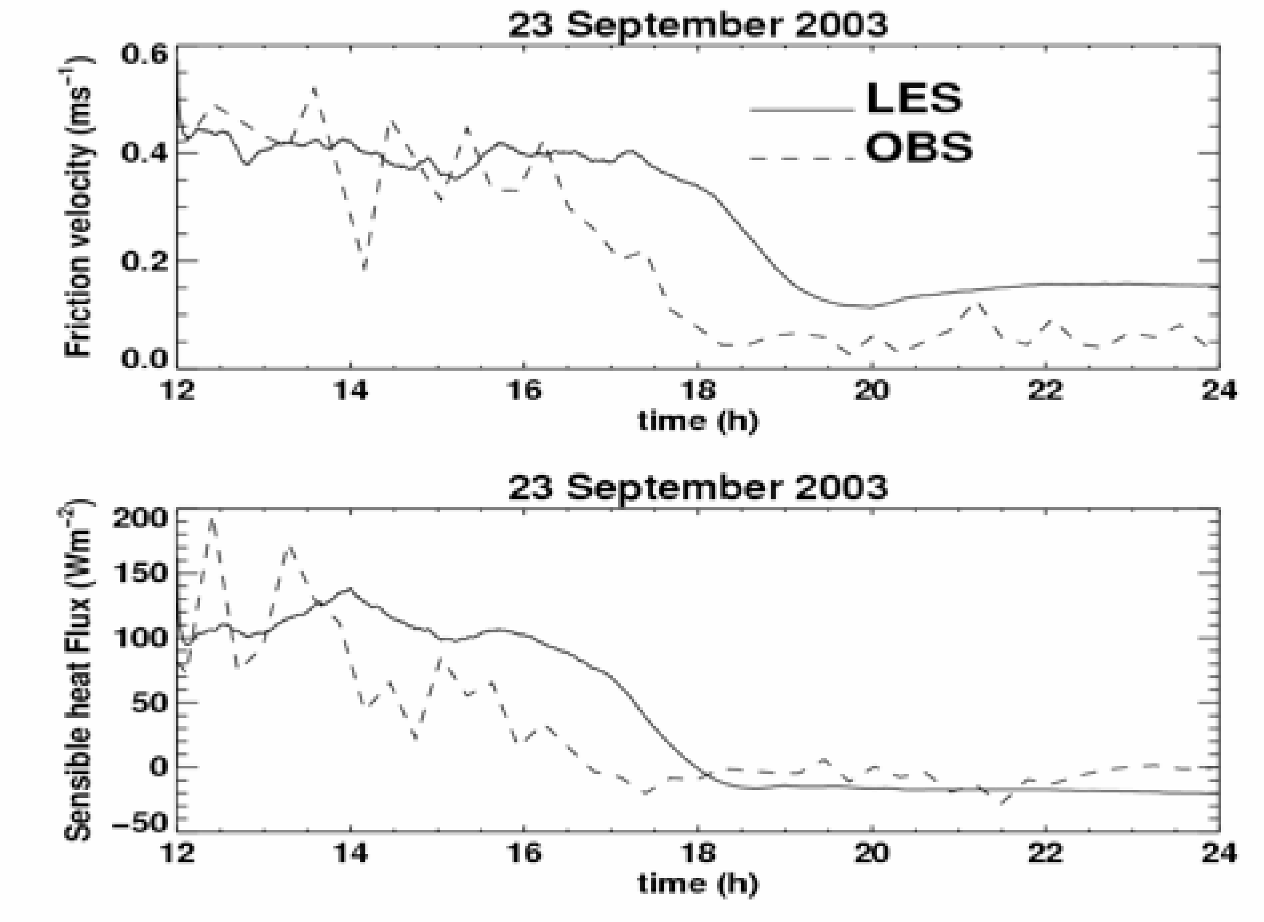

Fig. 12: After Beare et al, 2006. Comparison of the friction velocity and sensible heat flux between LES and observation. The late afternoon transition decay is delayed in the LES relative to the observations. A large improvement was found when assimilating the observations.

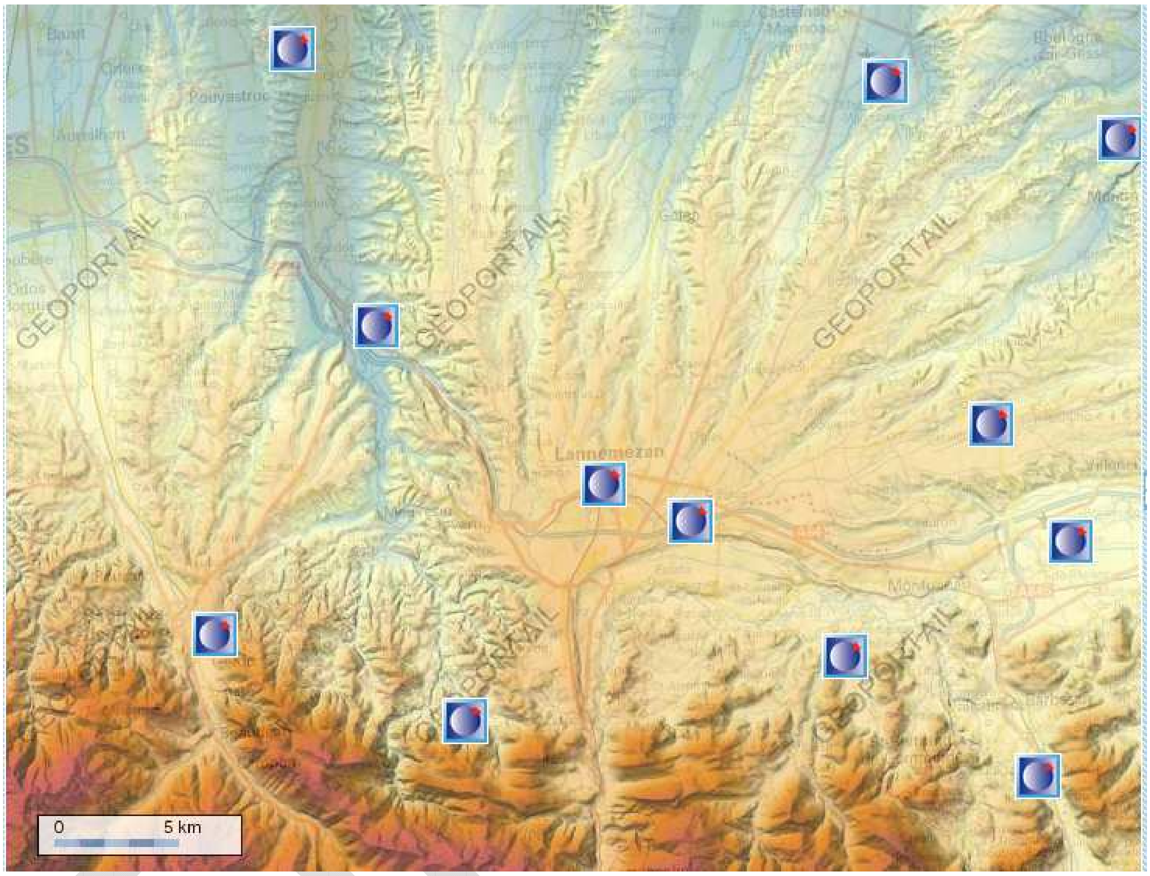

FIG. 13: Topography of the Lannemezan area, and synoptic Météo-France stations located around.

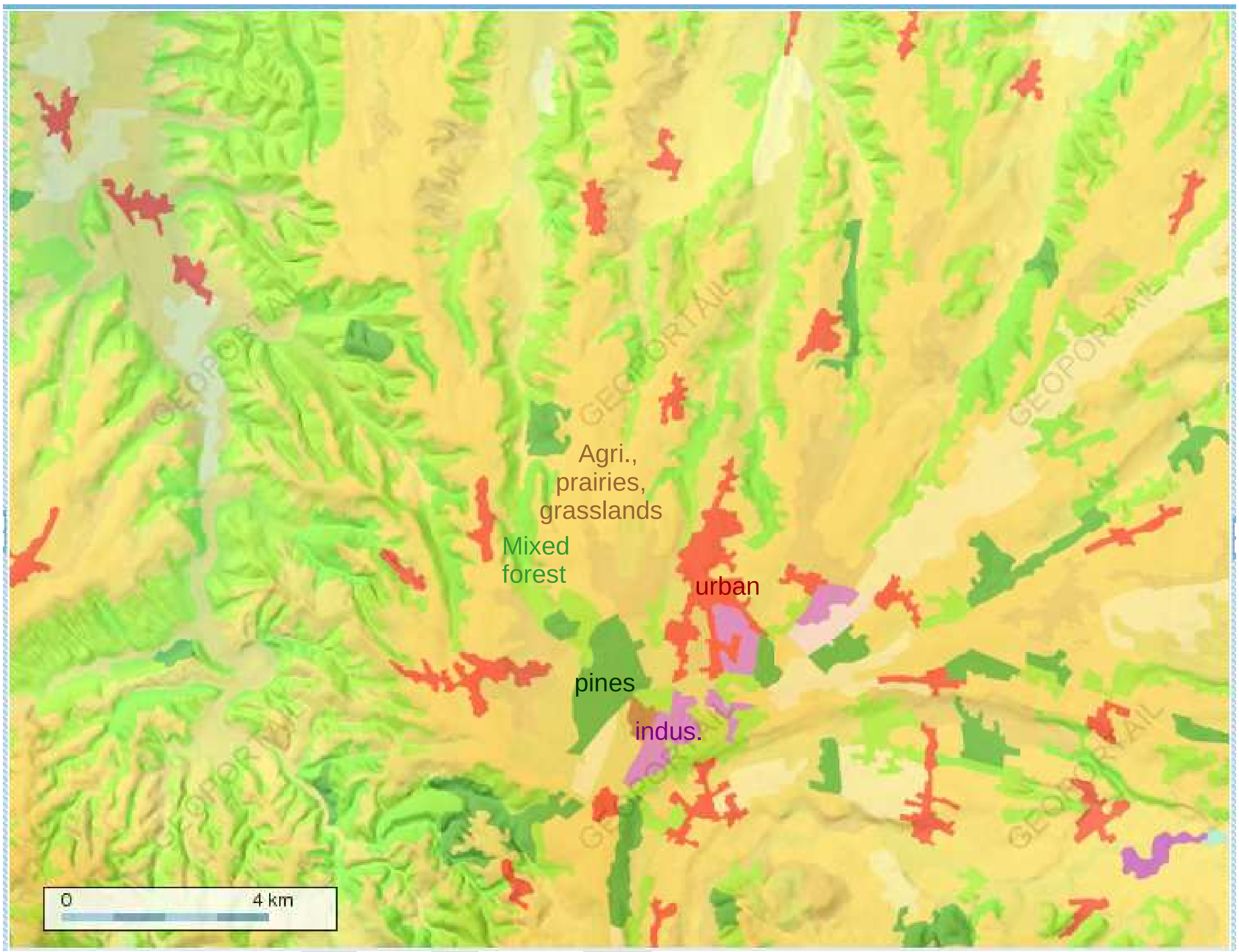

FIG. 14: Land use in the area.

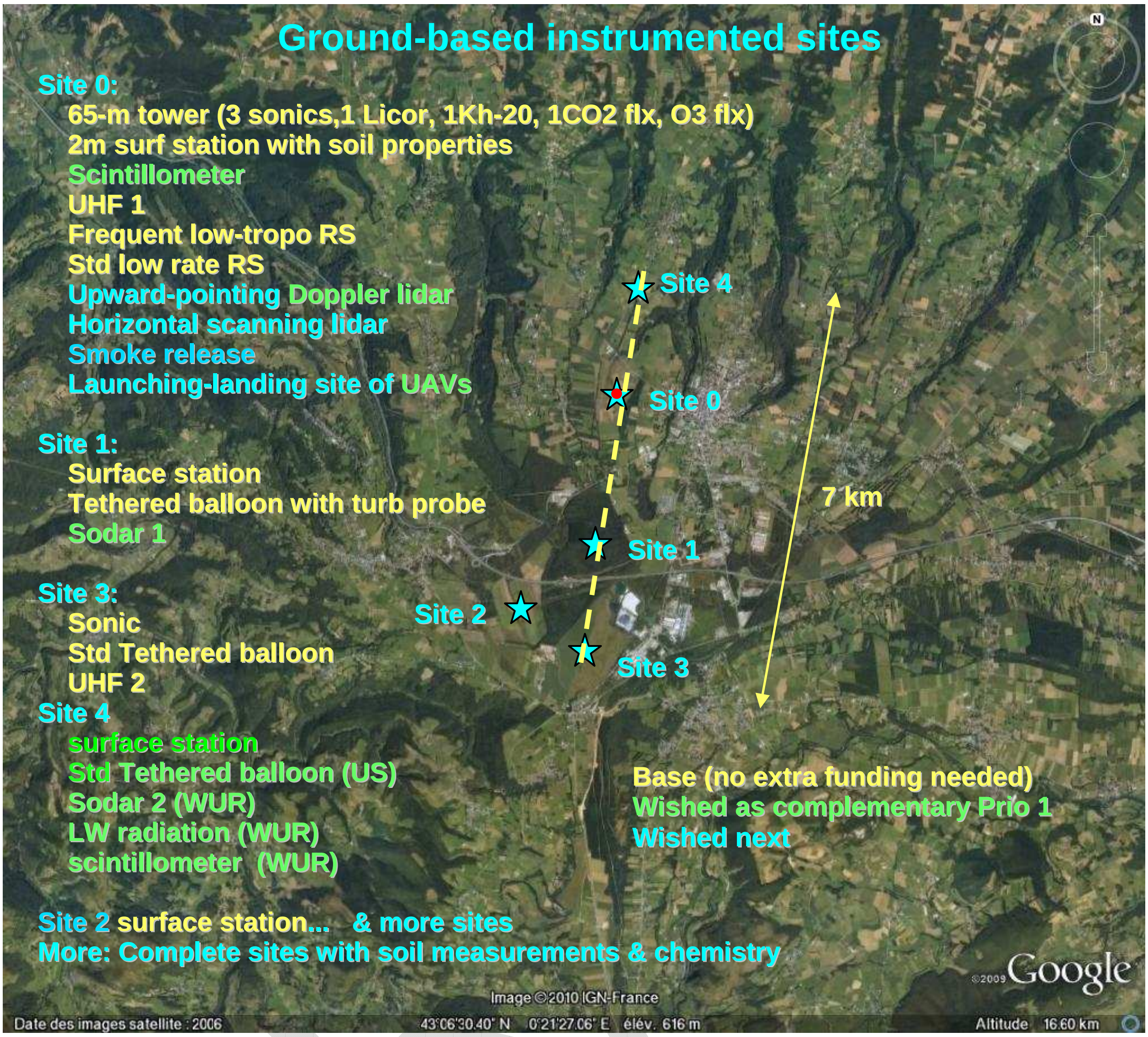

FIG. 15: Possible deployment of the ground-based sites around the main 65-m tower site.

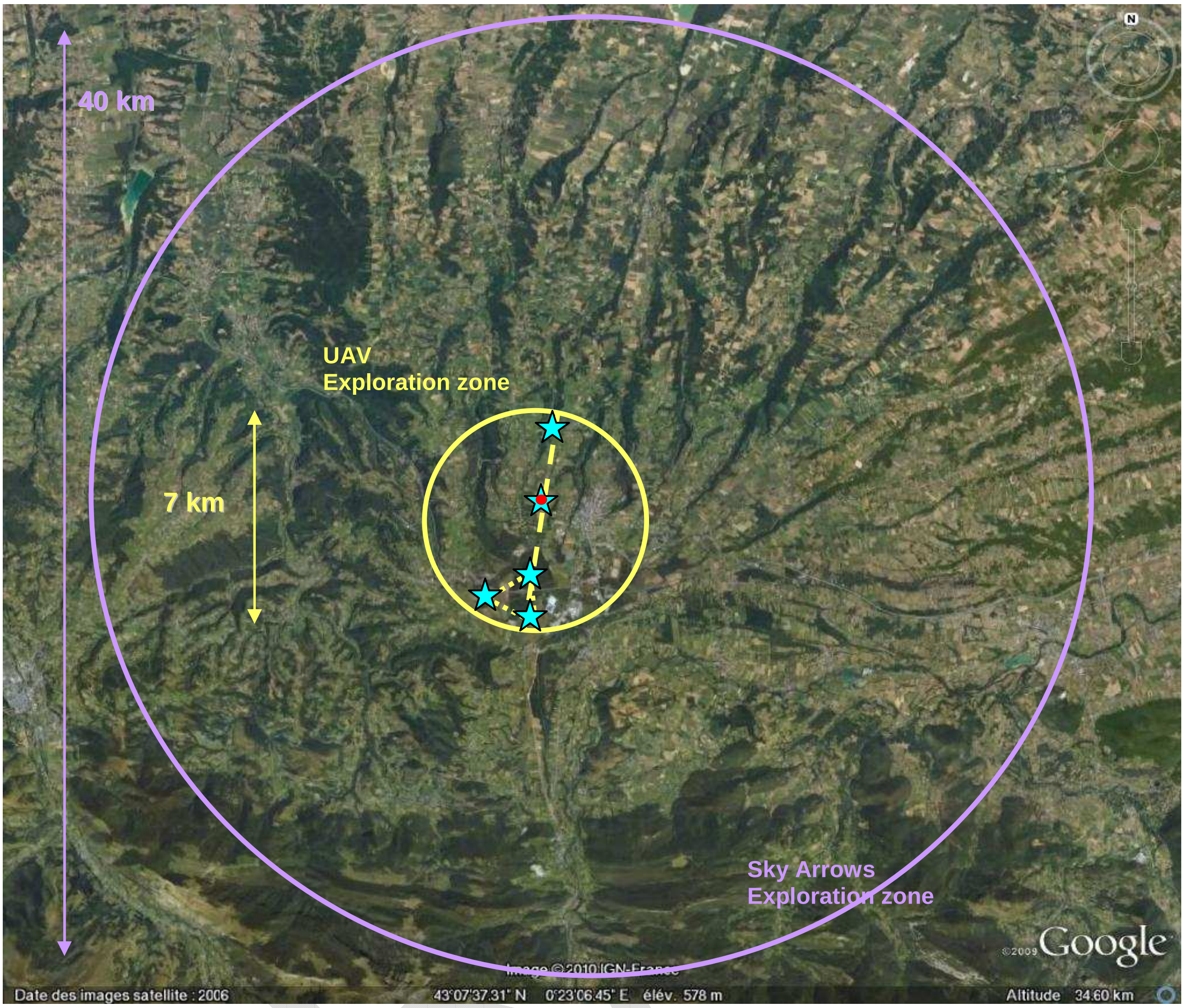

FIG. 16: Exploration zones of the sky arrow aircrafts and UAVs around the ground-based sites.