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Francesco Tampieri

Turbulence and Dispersion in the Planetary Boundary Layer

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ISSN 1610-1677 ISSN 1865-0678 (electronic)
Physics of Earth and Space Environments
ISBN 978-3-319-43602-9 ISBN 978-3-319-43604-3 (eBook)
DOI 10.1007/978-3-319-43604-3

Library of Congress Control Number: 2016950206

Printed on acid-free paper

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The registered company is Springer International Publishing AG Switzerland

Preface

The purpose of this book is to give an updated and problematic description of the atmospheric turbulence and the dispersion in the planetary boundary layer, for students, and perhaps to researchers, interested in the atmospheric sciences. Well-settled arguments coexists with topics investigated in the last years or even under discussion: thus, the reader will find differences in the treatment of the arguments that reflect these aspects. The hope is to give a comprehensive view of the physics of the planetary boundary layer, with the certainties and the uncertainties, to raise interest and to stimulate future research.

A couple of observations (which should become suggestions to the reader). The first one: many boring computations are put into the exercises. Most of them refer to analytic solutions of simplified problems. Although numerical modelling is of increasing use, the exact solutions may be a benchmark for models and are useful to better evidence the basic mechanisms and approximations. The student should not overlook this aspect.

The second one: in the Appendix to the Introduction, some data sets are cited, and reference is given to download the data. Note that the list is not exhaustive; other data sets are freely available. The data typically refer to variables averaged over some time interval. Most of the figures of the book are made using such data, so that every figure can be understood as an exercise for the student, who can remake, modify and possibly improve it. (As a general rule for the figures, the data are plotted directly, with various symbols, or binned in intervals of the independent variable and reported in terms of the median and the 10th and the 90th percentiles, with error bars.)

I would like to acknowledge here all who contributed to the genesis of the book: first, all my students who pushed me to build (as far as possible for me) a clear and unifying picture of the topic and the colleagues for discussions and criticism, in particular those who gave me data and special images. Special thanks to Domenico Anfossi, Alessandra Lanotte, Silvia Trini Castelli and Sergej Zilitinkevich, who suggested specific arguments and supplied the proper material. Last but not least,

I must remember two people who addressed my research and thus are in part responsible for the genesis of this book: Ottavio Vittori, who taught me not to cross the road on zebra crossing, and Julian Hunt, who introduced me to the mysteries of turbulence.

Contents

Acronyms and Symbols

CBL	Convective boundary layer
DNS	Direct numerical simulation
FPE	Fokker-Planck equation
IBL	Internal boundary layer
K41	Inertial subrange paradigm, from Kolmogorov (1941)
LE	Langevin equation
LES	Large eddy simulation
LSM	Lagrangian stochastic model
MOST	Monin-Obukhov similarity theory
NSE	Navier-Stokes equations
NWP	Numerical weather prediction
PBL	Planetary boundary layer
RL	Residual layer
SBL	Stable boundary layer
TKE	Mean turbulent kinetic energy
TPE	Mean turbulent potential energy
UBL	Urban boundary layer
UCL	Urban canopy layer
WMC	Well-mixed condition

a	A generic scalar variable; the inertial particle radius, Eq. (5.103)
$\mathscr{B}$	Buoyancy term in the TKE equation, Eq. (3.7)
$\tilde{c} = C + c$	Concentration of the scalar c, Sect. 2.2.4
c_{g}	Geostrophic drag coefficient, Eq. (3.77)
c_{d}	Drag coefficient for velocity in the UCL, Sect. 4.6.1.2
c_{p}	Specific heat at constant pressure, Sect. 2.2.5. For the dry air $1012\,\mathrm{N\,kg^{-1}\,K^{-1}}$

c_{U}	Drag coefficient for velocity, Eq. (3.32)
C_{K}	Inertial subrange Eulerian constant, reference value 2, Eq. (2.96)
C_{S}	Mean (average) concentration from the source S, Eq. (5.2)
C_0	Inertial subrange Lagrangian constant, reference value 6.2, Eq. (2.111)
C_θ	Inertial subrange Eulerian constant for scalars, Eq. (2.116)
D_{ij}	Eddy diffusion coefficient in transport processes, Eq. (5.10)
D_{uu}	Structure function for velocity of order 2, Eq. (2.95)
D_{uuu}	Structure function for velocity of order 3, Eq. (2.97)
E_{ab}	Cospectrum or Fourier transform of R_{ab}, Eq. (2.55)
E_{F}	Kinetic energy of the filtered field, Eq. (6.39)
E_{ij}	Velocity cospectrum, or Fourier transform of R_{ij}, the covariance between velocity components u_i and u_j, Sect. 2.3.3.8
$E_{\mathrm{k}} = \frac{1}{2}\langle u_i u_i \rangle$	Mean turbulent kinetic energy, Sect. 2.4.3.1
E_{P}	Mean turbulent potential energy, Eq. (3.12)
E_{S}	Sub-grid scale kinetic energy, Eq. (6.40)
f	Coriolis parameter, reference value $10^{-4}\,\mathrm{s}^{-1}$, Sect. 2.2.1; a generic function
f_{E}	Eulerian probability density function of the velocity, Eq. (5.3)
f_α	Probability density function of the random variable α, Sect. 2.3.1
$\boldsymbol{F}^{(c)}$	Flux of the scalar c, Sect. 2.2.4
$F_{\mathrm{r}} = \mathrm{U}/N\mathrm{L}$	Froude number, Sect. 2.2.7
g	Gravity acceleration, reference value $9.81\,\mathrm{m\,s^{-2}}$, Sect. 2.2.1; Richardson law constant, reference value 0.6, Eq. (5.34)
h	PBL depth, Sects. 1.3.3, 3.2, 3.5.2, 3.7.4
h_{b}	Blending height, Sect. 4.4.1
h_{c}	Roughness element height, Sect. 3.4.1; for an urban canopy, building height, Sect. 3.4.1
h_{i}	IBL depth, Sect. 4.2.1
h_{s}	Surface layer height scale, Sect. 3.2
H_{t}	Scale height of the topography, Sect. 4.3
H	Vertical height scale of the motion

k Wave number
k_{b} Wave number scale of the buoyancy-dominated range, Eq. (3.136)
K_{Q} Eddy diffusion coefficient for heat, Eq. (2.85)
K_{Δ} Eddy diffusion coefficient in LES, Eq. (6.55)
K_{τ} Eddy diffusion coefficient for momentum, Eq. (2.80)
Ku_{a} Kurtosis of the variable a, Eq. (2.34)

l Length in the inertial subrange, Sect. 2.5
l Instantaneous value of the parcel displacement, Sect. 2.4.6
l_{K} Viscous length scale, or Kolmogorov scale, Eq. (2.92)
l_{τ} Mixing length, Sect. 2.4.6
l_{Δ} Length scale for mixing in LES, Eq. (6.55)
L Obukhov length in the surface layer, Eq. (3.19)
L_{ab} Integral length scale independent on direction (for isotropic conditions), Sect. 2.3.3.8
$L_{\mathrm{ab}}^{(k)}$ Integral length scale in the k direction, Eq. (2.43)
$L_{\mathrm{A}i}$ Side of the averaging box, Sect. 2.3.2.3
l_{t} Inner layer depth over topography, Eqs. (4.15), (4.22), (4.23)
L_{t} Streamwise scale length of the topography, Sect. 4.3
L_{c} Penetration length scale in an urban canopy, Eq. (4.55)
L_{E} Integral Eulerian length scale in isotropic conditions, Sect. 2.5.4
L_{ij} Integral length scale for velocity components, Sect. 2.3.3.8
L_{r} Scale length of the roughness change, Sect. 4.4.1
L Global length scale of the motion

n_{S} Number of tracer parcels, Eq. (5.1).
N Brunt-Väisälä frequency, Eq. (2.29). Reference value 0.01 s^{-1}

$p_{\mathrm{f}} = p_{\mathrm{a}} + \tilde{p}$ Pressure of the air, reference value 1013 $\mathrm{Kg\,m^{-1}\,s^{-2}}$, Sect. 2.2.1
$P_{\mathrm{r}} = \nu/\kappa_{\mathrm{H}}$ Prandtl number, Sect. 2.2.7
$P_{\mathrm{t}} = K_{\tau}/K_{\mathrm{Q}}$ Turbulent Prandtl number, Sect. 3.4.2.3
$\mathscr{P}$ Shear production term in the TKE equation, Eq. (3.6)

$\tilde{q}_{\mathrm{c}} = \tilde{c}/\rho_{\mathrm{f}}$ Mixing ratio of the scalar c, Sect. 2.2.4
Q_{h} Horizontal kinematic heat flux, Eq. (2.86) and Sect. 3.4.3.4

$\boldsymbol{r}$	Separation vector
R	Gas constant for dry air, 287 $m^2\ s^{-2}\ K^{-1}$, Sect. 2.2.2
R_{ab}	Covariance between the quantities a and b, Eq. (2.42)
R_B	Bulk Richardson number, Eq. (3.21)
$R_e = UL/\nu$	Reynolds number, Sects. 1.2 and 2.2.7
R_f	Flux Richardson number, Eq. (3.9)
R_{fc}	Critical flux Richardson number, Sect. 3.1
R_g	Gradient Richardson number, Eq. (3.20)
R_{gc}	Critical gradient Richardson number, Sects. 3.7.1.3 and 3.7.3
R_{ij}	Covariance between velocity components u_i and u_j, Sect. 2.3.3.8
$R_o = U/(fL)$	Rossby number, Sect. 2.2.7
R_v	Gas constant for water vapour 641.5 $m^2\ s^{-2}\ K^{-1}$, Sect. 2.2.5.3
S	Tracer source for WMC, Eq. (5.3)
$S_c = \nu/\kappa_c$	Schmidt number, Sect. 2.4.1
S_t	Stokes number, Sect. 5.7
S_{ij}	Mean rate of strain, Eq. (6.4)
Sk_a	Skewness of the variable a, Eq. (2.33)
t	Time
T	Integral Lagrangian time scale in isotropic conditions, Sect. 2.5.4
T_{ab}	Integral time scale, Eq. (2.44)
T_A	Averaging time, Sect. 2.3.2.2
T_E	Integral Eulerian time scale in isotropic conditions, Eq. (2.121)
$T_f = T_a + \tilde{T} = T_{00} + T_0 + \tilde{T}$	Absolute temperature of the air, in Kelvin degrees, Sect. 2.2.2
T_{ij}	Integral time scale for velocity components, Sect. 2.3.3.8
T_K	Kolmogorov time scale, Eq. (2.93)
T_p	Inertial particle relaxation time (Stokes time), Eq. (5.103)
T_s	Characteristic time of the shear, Sects. 5.2.2.1 and 5.6.3.3
T_v	Virtual temperature, Sect 2.2.5.3
$\mathscr{T}$	Third-order moments in the TKE equation, Eq. (3.8)
$\tilde{\boldsymbol{u}} = \boldsymbol{U} + \boldsymbol{u}$	Eulerian velocity, Sect. 2.2.1
$u_* = \tau^{1/2}$	Friction velocity, Eq. (3.15)
u_{**}	Velocity scale in the transition sublayer, CBL case, Eq. (3.37)
$\boldsymbol{U} = (U, V, W)$	Mean velocity, Sect. 2.4.1

$\boldsymbol{U}_{\mathrm{g}} = (U_{\mathrm{g}}, V_{\mathrm{g}}, 0)$	Geostrophic wind, Sect. 2.4.1.1
U_{U}	Upstream (unperturbed) velocity, Eq. (4.3), Sect. 4.3.2
U	Global velocity scale of the motion
U_{l}	Velocity scale in the inertial subrange, at scale l, Sect. 2.5
$\boldsymbol{v}$	Lagrangian velocity of a parcel or a particle, Eq. (2.2)
$\mathbf{V} = U + \imath V$	Horizontal mean velocity in complex form, Sect. 3.1
$\mathbf{V}_{\mathrm{g}} = U_{\mathrm{g}} + \imath V_{\mathrm{g}}$	Geostrophic wind velocity in complex form, Sect. 3.1
v_{s}	Terminal velocity for particles, Eq. (5.104)
$w_*(z)$	Free-convection velocity scale in the surface layer, Eq. (3.35)
w_{e}	Entrainment velocity, Eq. (3.92)
$W_* = w_*(h)$	Global free-convection velocity scale, Eq. (3.83)
$\boldsymbol{x} = (x, y, z)$	Space coordinates; parcel/particle position, Eq. (2.1)
$\mathbf{X} = \langle uw \rangle + \imath \langle vw \rangle$	Turbulent vertical momentum flux in complex form, Sect. 3.1
z_{d}	Displacement height for the mean velocity profile, Eq. (3.33)
z_0	Roughness length, Eqs. (3.30), (3.31)
$z_{0\mathrm{e}}$	Effective roughness length, Sect. 4.4
z_{r}	Reference height, Sect. 2.2.5.2
ε	Viscous dissipation of TKE, Eq. (2.74)
ε_θ	Viscous dissipation of temperature variance, Eq. (2.77)
ΔS	Speed-up, Eq. (4.24)
Δu	Perturbation velocity in the inner layer, Sect. 4.3.4; velocity difference for parcel pairs, Sect. 5.3
ΔU	Perturbation to the mean velocity profile, Eq. (4.16)
$\Delta \tau$	Perturbation to the shear stress, Eq. (4.17)
$\Delta \theta$	Temperature jump at the inversion level for CBL, Sect. 3.6.2
$\zeta = \lvert z/L \rvert$	Nondimensional vertical coordinate in the unstable surface layer, Sect. 3.4.2
$\zeta = z/\Lambda$	Nondimensional vertical coordinate in the stable boundary layer, Sect. 3.7
θ_*	Temperature scale in the surface layer, Eq. (3.18)
θ_{**}	Temperature scale in the free-convection sublayer, Eq. (3.36)

$\theta_f = \theta_a + \tilde{\theta}$ $= \theta_{00} + \theta_0 + \tilde{\theta}$ $= \theta_{00} + \theta_0 + \Theta + \theta$	Potential temperature of the air, in Kelvin degrees, Sect. 2.2.5.2
$\Theta_f = \theta_{00} + \theta_0 + \Theta$	Average temperature of the flow, Sect. 2.4.4
Θ_{**}	Temperature scale in the CBL, Eq. 3.88
κ	von Karman constant, reference value 0.4 , Sect. 3.3
κ_c	Molecular diffusivity of the scalar c [$\mathrm{m}^2\,\mathrm{s}^{-1}$], Sect. 2.2.4
κ_H	Thermal diffusivity, Sect. 2.2.6. For the air 2.1 10^{-5} m^2 s^{-1} at 20 C
Λ	Local value of the Obukhov length, Eq. (3.102)
ν	Kinematic viscosity of the fluid, Sect. 2.2.1. For the air at the standard temperature of 20 C and the pressure of 1000 hPa $\nu = 1.5\ 10^{-5}\,\mathrm{m}^2\,\mathrm{s}^{-1}$
$\rho_f = \rho_a + \tilde{\rho}$ $= \rho_{00} + \rho_0 + \tilde{\rho}$	Density of the air, Sect. 2.2.1
ρ_p	Inertial particle density, Sect. 5.7
τ	Time lag; $\tau = \|\mathbf{X}\|$: modulus of the turbulent vertical momentum flux, Sect. 3.1
τ_{ij}	Area averaged value of $-\langle u_i u_j \rangle$, Eq. (4.56)
Φ_U	Nondimensional vertical gradient of mean velocity, Eq. (3.38)
Φ_Θ	Nondimensional vertical gradient of mean potential temperature, Eq. (3.47)
σ	Square root of the variance of a stochastic variable, Eq. (2.32)

Chapter 1
Introduction

Abstract In the Introduction the planetary boundary layer (PBL) is described in general, as the part of the atmosphere where turbulence acts driving exchange processes and dispersion. Attention is paid to field and laboratory measurements, as well as to the use of numerical experiments as a further tool for knowledge.

1.1 The Basic Definition of the Planetary Boundary Layer

The planetary boundary layer (PBL) is the lower part of the troposphere, where the interactions with the surface of the Earth occur.

Similarly to all the boundary layers that develop as a fluid flows over a surface, the PBL is (under suitable conditions, which normally are verified for the atmosphere) characterized by the turbulence, that affects the exchange processes. For this reason, in this textbook we shall discuss about turbulence. The interaction with the surface occurs due to the exchange of momentum, of heat and of scalars (like the water vapour): the surface (the bare ground, a vegetative or a urban canopy, the sea) is a sink of momentum, but can be a source or a sink of heat or other scalars. Understanding these interactions is an important step for the proper modelling of weather and climate, and, in general, of the dynamics of the atmosphere.

The turbulence affects the transport of tracers (pollutants), which is a relevant issue in the air quality applications, and, more generally, in the study of the composition of the atmosphere (linked with climate). Besides the applications, the transport problem is related to the intimate nature of the turbulent flows, so that it deserves special attention also from a fundamental (theoretical) point of view.

Turbulent boundary layers are not limited to atmospheric flows: the general findings are relevant in geophysics as well as in engineering. However, some features are specific for the atmosphere, and will be detailed as possible.

F. Tampieri, *Turbulence and Dispersion in the Planetary Boundary Layer*,
Physics of Earth and Space Environments, DOI 10.1007/978-3-319-43604-3_1

1.2 A Few Words About Turbulence

An exhaustive treatment of this issue is well beyond the science of the author and the scope of this book. The reader must refer to the many textbooks starting from Monin and Yaglom (1971, 1975) for a classical introduction, with attention to geophysical applications, to the apparently simple Tennekes and Lumley (1972), to Pope (2000) in particular for turbulence modelling, to Wyngaard (2010) again focusing on the atmosphere. A suggested reading to go into the specific, but widely quoted, argument of the inertial subrange is Frisch (1995).

Here a short summary is presented, for the aspects of direct interest for the present study. A turbulent flow is characterized by random features of the state variables (velocity, temperature, or tracer concentration, for instance), by the existence of a wide range of scales of the motion (in terms of space and time), and by mixing properties.

The Reynolds number $R_e = \mathrm{UL}/\nu$ characterizes the flow of a viscous fluid. Here U is a velocity scale, for instance the average velocity, and L a length scale, for instance the depth of the fluid, or the width of the channel. The internal, molecular, friction is measured by the kinematic viscosity ν. If $R_e \sim 1$ the flow is laminar (and predictable). As R_e increases beyond 1000, say, the hydrodynamic instabilities make unpredictable some features of the flow. The velocity, for instance, is continuous (the fluid is viscous), but the accelerations can be quite large, and for many practical purposes the flow velocity can be considered a stochastic variable. The consequence is the need to give statistical descriptions of many phenomena characterizing the turbulent flow, and to refer to the probability density functions of the state variables. A time record of the velocity components and of the temperature in an atmospheric boundary layer is reported in Fig. 1.1, which qualitatively illustrates the random features of the variables characterizing the flow.

A second relevant feature of the turbulent flows is the presence of a range of time and space scales of the motion; in other words, the stochastic variables are not white noise, but are correlated in time and space (there is an underlying structure). This aspect can be qualitatively understood looking at the time patterns of Fig. 1.1: the high frequency fluctuations are superimposed to a fluctuating trend, characterized by longer time scales. The motion is organized in eddies, with spatial scales that range from those imposed by the boundaries down to scales small enough that their specific (computed from their length and velocity scales) Reynolds number is small, and viscosity dominates.

The internal structure of the turbulent flow produces important effects on the mixing. At small R_e the transport of scalars is described by the Fick law, i.e. the small scale molecular motion produces a large scale transport of the scalar in the direction of minus the gradient of the mean concentration (down-gradient transport): the paradigmatic case of diffusion occurs. At large R_e the eddies produce mixing, but the existence of a continuous range of scales inhibits the straightforward application of the diffusive, Fick law, approach. In a turbulent flow, the eddies may be as large as the scale of the gradient, and the transport may become non-Fickian.

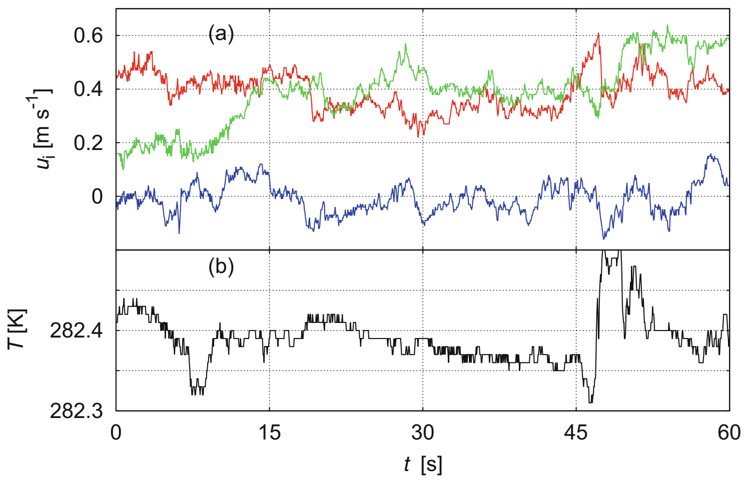

Fig. 1.1 Time series of the along-wind (*red line*), transversal (*green*) and vertical (*blue*) components of the air velocity fluctuations (**a**) and of the temperature (**b**), measured by a sonic anemometer at 7.5 m above the ground (CCT data). The high frequency record highlights the fluctuating behaviour of the observations and the different mean values of the three velocity components. Courtesy Mauro Mazzola, CNR ISAC

The space and time resolution of the observations and/or of the numerical models allow to resolve (i.e., explicitly describe) the larger scales of the motion. This aspect will be considered in detail in the following chapters; here it may be interesting to note that traditionally the motion of the atmosphere has been divided in mean wind and turbulence, just because the anemometers take an average over some time interval (of the order of minutes), so that the resolved variable (the averaged one) is the wind, the unresolved part is the turbulence. The arbitrariness is evident: from the point of view of seasonal dynamics of the atmosphere, the mid-latitude perturbations are eddies, i.e. turbulence.

1.3 The Structure and Evolution of the PBL

Quite schematically, the energy balance of the atmosphere in the PBL is related to the effect of the large scale motions of the atmosphere itself (winds and waves) and to the exchanges of momentum and heat at the surface. These exchanges are related to the radiation balance (incoming and outcoming radiation) and to the heat flux into the ground. Moisture effects can be relevant.

1.3.1 Local Equilibrium

As the vertical fluxes are large and the horizontal conditions pretty homogeneous, the PBL dynamics is dominated by the local conditions, i.e. its properties are only function of the distance from the surface, while the effects of the horizontal heterogeneities can be neglected. The local equilibrium is the basic paradigm for the study and the understanding of the phenomenology of the PBL, and will be discussed in Chap. 3.

Over a solid surface, if radiation is weak (overcast sky), the wind drives the turbulence and thus the exchanges. The turbulence is mainly produced by the shear, while thermal effects are minor: these are called quasi-neutral conditions (perfectly neutral conditions, occurring in absence of heat exchange, are probably realized only in the laboratory). In presence of a diurnal cycle, over the land, we shall take into account the time evolution of the radiative flux: the radiative flux during the day heats the ground and thus the air (turbulence increases, leading to the so-called unstable conditions) and cools both during the night (decreasing turbulence, leading to stable conditions). Intense heating of the ground gives origin to eddies of vertical size of the order of the PBL depth, with quite large positive vertical velocity: this is the onset of convection, a very efficient mechanism of mixing of the entire layer. Strong cooling damps the turbulence and the related exchanges, layers of air at different heights may be nearly independent on each other, the surface may become almost unimportant while phenomena occurring aloft have a relevant influence on the dynamics. The effect of radiation depends on the heat capacity of the surface, so that it is enhanced over the desert and almost negligible over the sea. The diurnal cycle disappears at high latitudes, leading to PBL characterized by a slow time evolution.

Remote sensing techniques based on the detection of density fluctuations and tracers in the atmosphere (SODAR and LIDAR respectively: see Sect. 1.5 for references to the instruments) allow the visualization of the vertical structure of the PBL: see Figs. 1.2, 1.3, 1.4 and 1.5.

Figures 1.2 and 1.3 depict the daily evolution of the PBL at midlatitudes, for a winter and a summer case, using a SODAR. During the night stable conditions prevail, and the density fluctuations (evidenced by the dark areas) appear to be highly variable in the vertical, and coherent in time, especially for the winter case (almost undetectable in the summer case). Note also the wave pattern at about 400 m from 3 h to 5 h on Feb. 8.

During the day the heating at the ground leads to the onset of convection, with large scale turbulent motions (of vertical size as the PBL itself) evidenced by the signals (dark lines) rapidly variable in time and vertically coherent. The convection starts around 6 h in the summer case and just before 10 h the plumes are so high to go beyond the vertical range of measure (i.e. the PBL depth in this case is greater than 800 m). In the winter case, convection begins around 8 h and is weaker: we can guess that the PBL depth reaches about 800 m for a few hours, around 14 h.

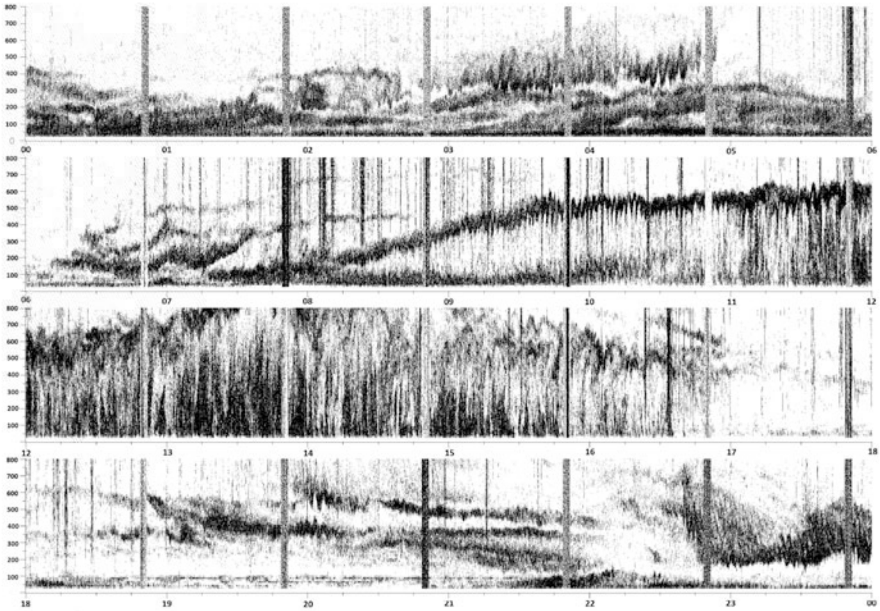

Fig. 1.2 SODAR echogram of the daily cycle in a winter day (February 08, 2011) at Castelporziano, a rural site near Roma, Italy. The local time (in hours) is reported in abscissa, the height (m above the ground) in ordinate. The *vertical gray stripes* correspond to calibration time intervals. Courtesy Angelo Viola, CNR ISAC, Roma

The development of a convective boundary layer is visualized also using LIDAR (which gives an estimate of the presence and the concentration of aerosol particles): see Fig. 1.4. Thanks to the larger measuring range, it shows the single convective cells that extends up to about 1.5 km in a typical summer sunny day.

A further SODAR sounding of the long lasting stable PBL in Antarctica is shown in Fig. 1.5. As expected, the absence of any time modulation contrasts with the mid-latitude cases; it is also worth noting the small depth of the layer, which reduces to a tenth of meters at the end of the period.

The presence of the diurnal cycle underlines the importance of the time evolution of the forcing mechanisms: the transitions from stable to unstable/convective conditions and vice versa, or the nocturnal radiative cooling, causing turbulence of decreasing intensity. Note that in general these phenomena occur on a time scale greater than the typical time scales of the turbulent flow, which then usually adjusts to the changing conditions.

Over the sea, some features characterize the PBL and distinguish it from the continental counterpart: the presence of a mobile lower boundary, which adjusts to a certain extent to the dynamics; the constant presence of moisture; the easier occurrence of homogeneous and steady conditions (apart from the coastal regions or in presence of cold/warm outbreaks); the diurnal cycle is small as well as the

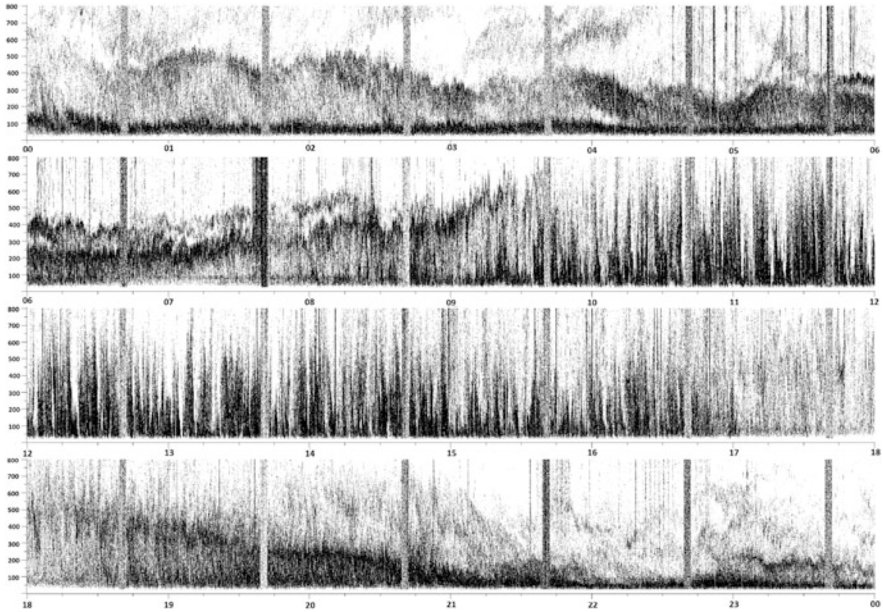

Fig. 1.3 As in Fig. 1.3, but for a summer day (August 31, 2010). Courtesy Angelo Viola, CNR ISAC, Roma

departures from near-neutral conditions (the air mass is in thermal equilibrium with the surface); large-scale eddies appears in form of rolls (organized motion).

1.3.2 Heterogeneities and Unsteadiness

In the real world, horizontal heterogeneities and unsteadiness on short time scales often occur, which means that the picture of the previous section must be revised, and the paradigm of local equilibrium loses its general validity, as discussed in detail in Chap. 4.

Broadly speaking, we can recognize two different situations in which heterogeneity is relevant, which we call 'large scale' and 'small scale', that undergo different treatments. The term 'large scale' means that we are able to make measurements and/or to realize numerical simulations and/or physical models which describe explicitly this situation. Its counterpart (the 'small scale' effects) occurs for heterogeneities/unsteadyness on space/time scales smaller than those we are able, or we want, to consider explicitly.

Large scale effects can be analyzed in detail; for instance, topographic features (hills and valleys, coastal borders) affect the wind field and the heating of the ground, thus produce horizontal variations of the forcing. Local circulations arise

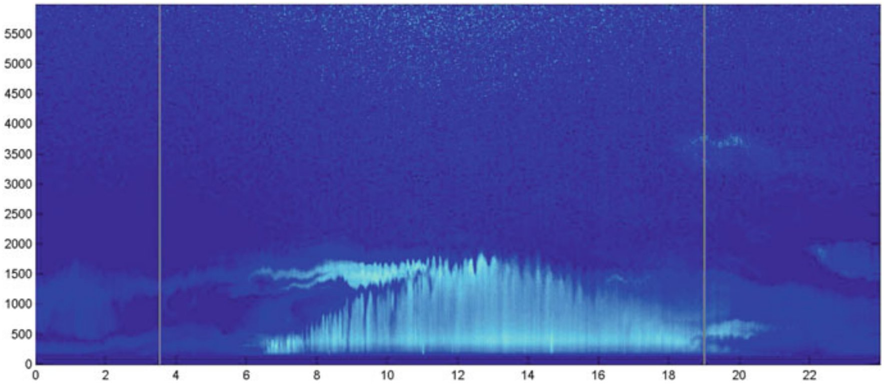

Fig. 1.4 LIDAR sounding of the convective boundary layer, for June 27, 2012, at San Pietro Capofiume, a site in the Po Valley, Italy. On the abscissa, time in hours, UTC. On the ordinate, the height in m above the ground. Courtesy Gianpaolo Gobbi, CNR ISAC, Roma. The presence of aerosols is evidenced by the almost *white areas*; *blue* means no particles. The *two vertical lines* correspond to sunrise and sunset. Further informations about the structure of the atmosphere on this day are reported in Sect. 3.6

(which are outside the scope of this book, but must be taken into account to understand the real, not too idealized, PBL). Changes of surface characteristics (grass to trees, land to sea) produce internal boundary layers; for instance, near the coastal line, during the night in the cold season, in presence of a land breeze, cold (stable) air blows over a warmer sea, generating an unstable internal PBL. The transitions related to the diurnal cycle (cited above) are a typical example of large scale unsteadiness.

Small scale effects are considered in terms of modifications of the rules characteristic of the local equilibrium paradigm: we abdicate the universality of the rules, maintaining some formal features and parameterizing the dependence on the small scale in the numerical value of the coefficients which appears in the formulas. The main consequence is the difficulty in finding general formulations of the parameterization.

The trend for tackling heterogeneous conditions is to refine the scale of the description, moving from parameterization to explicit treatment, thanks to the increasing computational power and the improvement of the observations. On the other hand, it is sometimes almost impossible, or unconvenient, to deal explicitly with all the details, while an averaged description may be all that we need. Situations like the wind and the vertical exchanges in a forest, or the meandering of the wind when the wind itself is quite low are examples of problems that typically do not require (or even do not permit) a detailed description. Note that both the 'large scale' and the 'small scale' approaches occur in the investigation on the urban PBL, and the different descriptions coexist in the common practice.

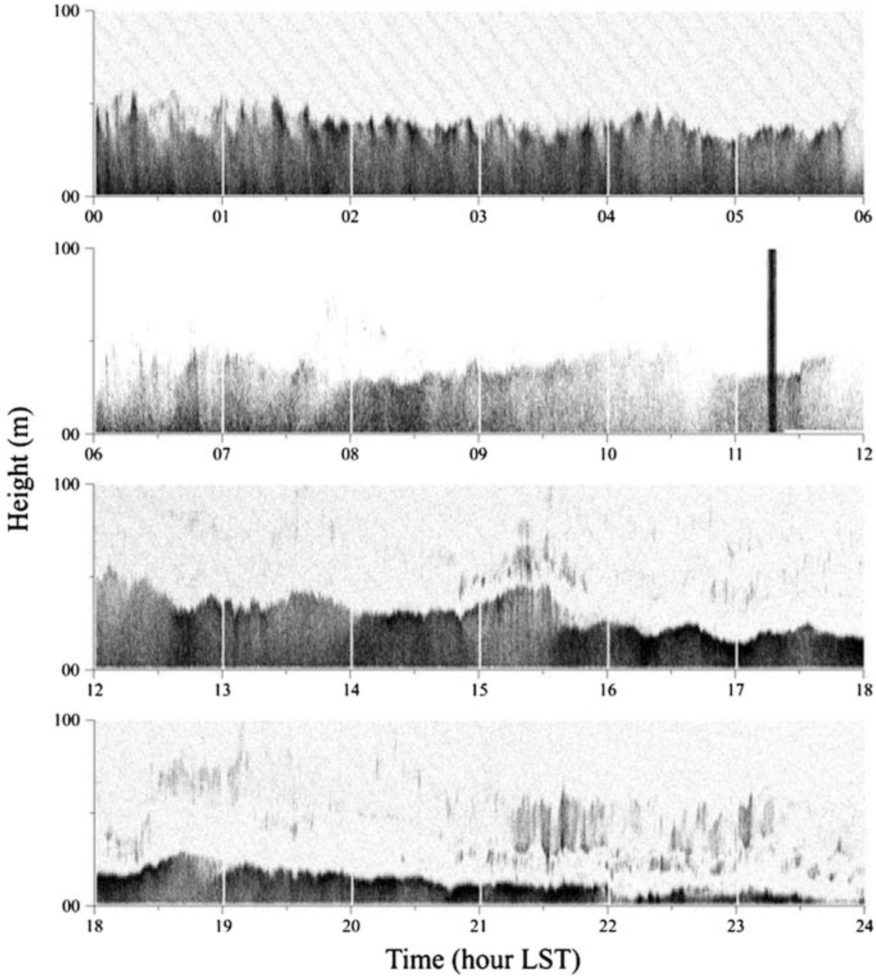

Fig. 1.5 Echogram of the stable boundary layer during the winter (August 20, 2012) at Concordia Station, Dome C, Antarctica, by an high resolution Surface-Layer MiniSODAR. Heights in m, local time in hours. Courtesy Stefania Argentini, CNR ISAC, Roma

1.3.3 The Boundary Layer Depth

The PBL depth h can be broadly defined as the height at which the interaction of the tropospheric flow with the surface becomes negligible. In spite of the fact that it is not a directly measurable quantity, a lot of words are spent about its determination, essentially because of the practical importance in modelling applications. This depth is quite evident in Fig. 1.4 as the level at which the aerosol sharply disappears, or in

Fig. 1.2 or Fig. 1.3 as the level at which convection stops, during daytime. Note thus that the depth is well identifiable in some conditions, and less well in others.

The traditional analysis of the PBL is largely based on the assumption that the surface fluxes play the major role in its dynamics and evolution, and h is the height at which these surface fluxes become negligible. The straightforward consequence is that the turbulent fluxes (at least of momentum and heat, but not only) are functions of the height z.

However, the surface fluxes may be small, or not relevant. In convective conditions, the mixing at the PBL top (i.e., the downwards heat flux from the troposphere) contributes to the growth of the layer. In stable conditions, the vertical fluxes may become negligible at the surface, while the turbulence is produced by the wind shear or by the waves aloft. In such cases the basic definition remains correct, but h cannot be estimated from surface fluxes.

Earth rotation introduces a further limitation to the PBL growth. So the final statement is that h is related to the surface fluxes, to the stability and wind conditions aloft (at its top), and to the Coriolis effect, if the geometry of the surface (hills and valleys) can be neglected.

1.4 The Transport Problem and the Turbulent Dispersion

The behaviour of a tracer advected by a turbulent flow shows complex, chaotic features. The tracer may be a pollutant, may be the temperature, may be a microorganism living in the sea. Some tracers are characterized by an almost infinite living time: they do not react with the environment; others combine by chemical reactions. Radioactive tracers change their properties according to their decay time.

Some tracers behave like the fluid molecules, i.e. their velocity is at any time equal to that of the flow: we shall refer to as fluid parcels. A fluid parcel is supposed to be an ideal small volume of fluid which can be identified: in general gaseous pollutants behave as parcels.

Other tracers are subject to the gravity acceleration and have their own dynamics, due to the inertia, like aerosols or water droplets: we shall refer to as inertial particles. In general, their velocity is different from that of the flow, and at least in certain conditions this aspect becomes critical.

Last but not least, some substances interact with the dynamics of the flow: for instance, temperature changes can occur associated with chemical reactions (this aspect will not be treated in this book).

The transport problem is tackled by computing the trajectories of the parcels, or of the particles, which means that their velocities must be known as function of time. In a turbulent flow, characterized by a wide range of scales of motion, this computation can be done in general for the large scales, but not for the small scales (a notable exception is given by the use of direct numerical simulations (DNS) of the flow). The problem is solved in a statistical sense, that is, in the computation of some moments of the positions of the parcels (or particles), or of their relative