3 The Boundary Layer and Related Phenomena

In this chapter we focus on a qualitative overview of the boundary layer, describe the inclusion of turbulence in the dynamical equations, discuss several related atmospheric phenomena, and consider applications of boundary layer theory.

3.1 Introduction

3.1.1 Definition

What is the *boundary layer*?

Stull (1988):

> *Planetary boundary layer is the part of troposphere that is directly influenced by the presence of the earth surface and responds to surface forcings with a timescale of about an hour or less.*

Sorbjan (2012):

> *The lowest portion of the atmosphere, which extensively exchanges mass (water), momentum, and heat with the Earth's surface.*

The boundary layer, also known as the atmospheric boundary layer (ABL) or planetary boundary layer (PBL), is a disturbance of the lower atmosphere induced by the underlying surface of the Earth. In other words, the boundary layer is an interfacial layer between the troposphere and the ground. The forcings that originate at the surface include frictional drag, evaporation and transpiration, heat transfer, pollutant emissions, and terrain-induced flow modifications.

In the lowest few millimeters of the boundary layer, conduction between the air and the ground is important. This small layer is also known as the *viscous sublayer*. Above this layer, the effects of molecular diffusion are ignored since they are negligible compared with the mixing effects of turbulent eddies.

Though molecular viscosity effects do not directly influence boundary layer motions above a few centimeters, the existence of the viscous sublayer is crucial in the formation of boundary layer eddies. In other words, the boundary layer would not exist without an underlying surface.

Specifically, as a result of molecular viscosity the flow vanishes at the surface (*no-slip boundary condition*). Consequently, large vertical wind shear is generated near the ground. This can be true even with light winds. This shear is critical in generating the turbulent eddies that act to transfer momentum, heat, and moisture to the lower boundary layer. These turbulent motions have spatial and temporal variations at scales much smaller than those resolved by the meteorological observing network.

The boundary layer flow is dominated by these turbulent eddies, as well as by those that result from surface heating. While these eddies are most often ignored in the free atmosphere, their consideration is a necessity for boundary layer flows. The transfer of momentum, heat, and moisture by these turbulent eddies must be accounted for in the dynamical equations in order to accurately describe the evolution of temperature and moisture fields and the relationship between the pressure gradient and wind.

3.1.2 The Importance of the Boundary Layer

The boundary layer is often perceived as a boring, esoteric topic to meteorology students. This is likely due to the fact that much of the theory historically stems from fluid mechanics and other engineering-focused areas of research.

Admittedly, boundary layer theory is laden with long equations and complex parameterizations. It is understandable why so many students' eyes glaze over when they are introduced to the boundary layer.

We will soon cover why these equations and parameterizations are necessary to properly describe the evolution of mass, momentum, and moisture fields in the lowest portion of the atmosphere. First, let us put aside rigorous math exercises and consider real-world examples that illustrate the importance of the boundary layer.

Why should you care about the boundary layer?

- We live in the boundary layer!
- Forecasts of dew, frost, minimum and maximum temperatures (to name a few) are really boundary layer forecasts.
- Pollution is trapped and dispersed within the boundary layer.
- Fog occurs in the boundary layer
- The primary energy source for the atmosphere is solar radiation, which is generally absorbed by the surface. Boundary layer processes act to transmit this energy to the rest of the atmosphere.
- $\sim 90\%$ of the net radiation absorbed by oceans causes evaporation. The latent heat stored in water vapor accounts for $\sim 80\%$ of the fuel that drives atmospheric motions!
- Crops are grown in the boundary layer, where pollen is distributed within.
- Cloud nuclei are sent into the air from the surface by boundary layer processes.

- Almost all water vapor that reaches the free atmosphere is transported through the boundary layer by turbulence and advection.
- Thunderstorm/hurricane evolution is tied to the inflow of moist boundary layer air.
- Downward turbulent transport of momentum through the boundary layer to the surface is the single most important atmospheric momentum sink.
- Turbulence/gustiness affects the design of structures.
- Wind turbines extract energy from boundary layer flows.
- Wind stress on the ocean surface is the primary energy source for ocean currents.

The list is not exhaustive, but it shows several tangible examples in which the boundary layer directly impacts our lives, in addition to indirect effects through its influence on weather.

3.1.3 Comparison of Boundary Layer and Free Atmosphere Characteristics

Another way to illustrate the importance of the boundary layer is to compare its characteristics with the overlying free atmosphere. The differences outlined in Table 1 highlight the role of the boundary layer and its importance to the entire atmosphere.

Table 1: Comparison of boundary layer and free atmosphere characteristics (adapted from Stull 1988).

Property	Boundary Layer	Free Atmosphere
Turbulence	Almost continuously turbulent over its whole depth.	Turbulence in convective clouds and sporadic clear air turbulence in thin layers of large horizontal extent.
Friction	Strong drag against the earth's surface. Large energy dissipation.	Small viscous dissipation.
Dispersion	Rapid turbulent mixing in the vertical and horizontal	Small molecular diffusion. Often rapid horizontal transport by the mean wind.
Winds	Near logarithmic wind speed profile in the surface layer. Subgeostrophic, cross-isobaric flow is common.	Winds nearly geostrophic.
Vertical transport	Turbulence dominates.	Mean wind and cumulus-scale dominate.
Thickness	Varies between ~ 100 m to 3 km in time and space. Diurnal oscillations over land.	Less variable at 8-18 km. Slow time variations.

3.2 Structure and Evolution of the Boundary Layer

The following figure illustrates the typical diurnal cycle of the boundary layer. This pronounced cycle is one of the defining features of the boundary layer owing to its fast response time to surface forcings. The diurnal evolution of the boundary layer, as well as the associated structural changes, are discussed in detail below.

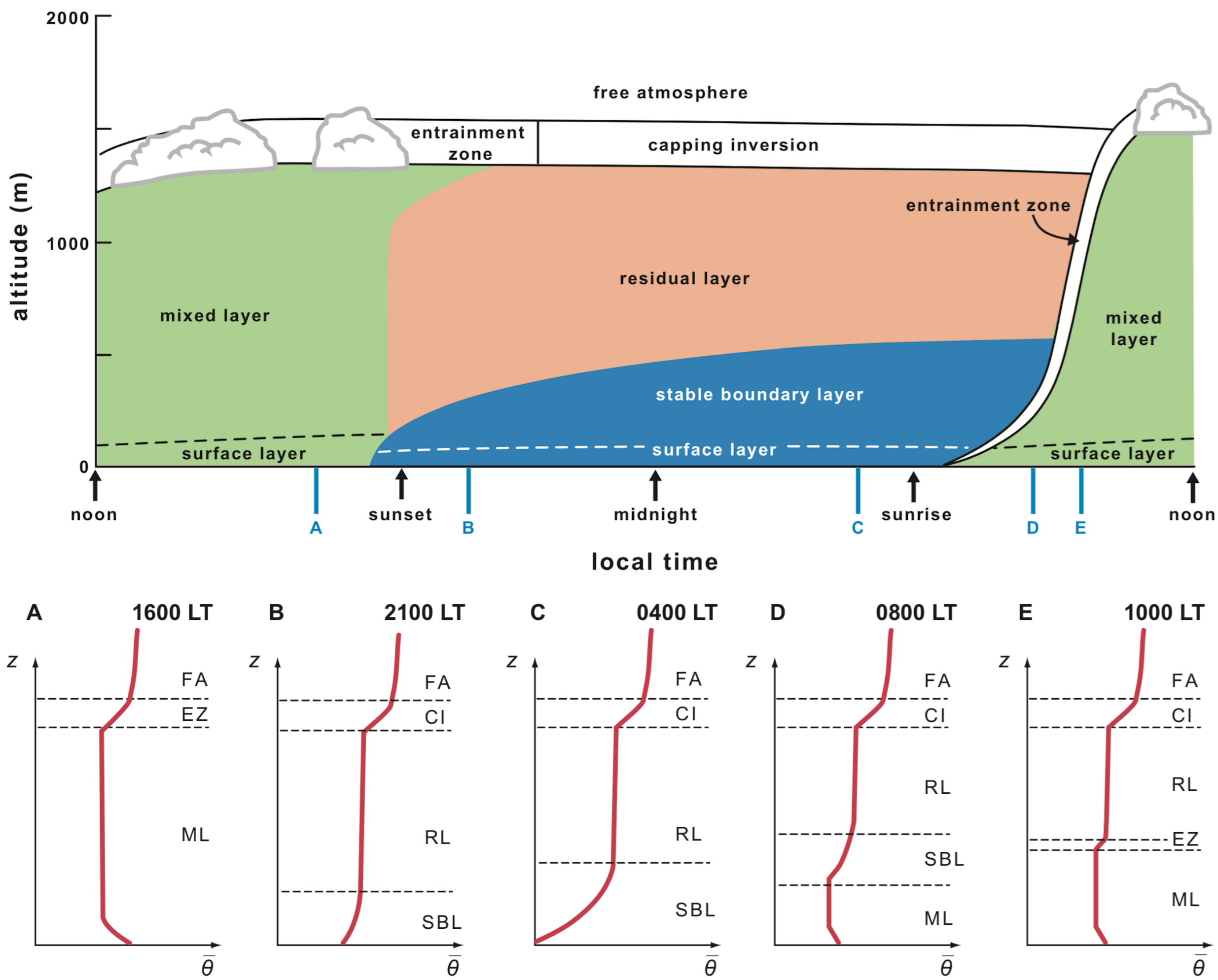

Figure 1: The boundary layer in relatively tranquil conditions over land consists of three major parts: a very turbulent mixed layer, a less turbulent residual layer consisting of former mixed-layer air, and a nocturnal stable boundary layer of sporadic turbulence. The letters A–E identify the times of the soundings. The vertical profiles of potential temperature show the evolution of the boundary layer during a diurnal cycle starting at about 1600 local time. The free atmosphere (FA), entrainment zone (EZ), capping inversion (CI), mixed layer (ML), residual layer (RL), and stable boundary layer (SBL) are indicated. [From Markowski and Richardson]

3.2.1 Boundary Layer Depth: Dependencies

- morning atmospheric profile of temperature
- intensity of turbulent mixing, which itself depends on
  - the amount of insolation and sensible heat flux (buoyancy-driven turbulence)
  - mean vertical wind shear (mechanically-driven turbulence)

3.2.2 Boundary Layer Depth: Variations

- stable, nighttime conditions
  - as shallow as a few tens of meters
  - intermittent turbulence can exist
- unstable, daytime conditions
  - as deep as several kilometers
  - layer is typically superadiabatic and dominated by convective motions
- spatial changes
  - deeper boundary layers are favored in warmer, drier climates
  - depth can exceed $4\ \mathrm{km}$ in the desert

3.2.3 General Boundary Layer Evolution

- After sunrise, the boundary layer is heated by the underlying surface through sensible heat flux.
- This heating drives the air in contact with the ground toward the dry adiabatic lapse rate $(\partial\overline{\theta}/\partial z = 0)$.
- Within a shallow layer in contact with the ground, superadiabatic conditions induce vertical mixing and turbulence.
- Mixing promotes homogeneity, which leads to moisture and wind speed profiles that are approximately constant with height.
- Thermals rise and penetrate the stably-stratified atmosphere.
- This *penetrative convection* induces mixing through an even deeper layer, leading to a deepening of the boundary layer.
- These overshooting thermals lead to cooling at the top of the mixed layer within the penetrated stable region, which in turn creates a capping inversion.
- Continued surface heating generates more thermals, deeper mixing, and a growing boundary layer.
- The maximum boundary layer depth is attained near sunset.
- After sunset the surface sensible heat flux reverses sign.
- As a result of the negative surface sensible heat flux, cooling starts and a stable region forms near the surface.

3.2.4 Daytime Boundary Layer Structure

- The layer of the daytime boundary layer extending from the surface to the *entrainment zone* (EZ; which tops the boundary layer) is often referred to as the *mixed layer* (ML) or *convective boundary layer* (CBL).
- Profiles of potential temperature ($\overline{\theta}$) are approximately constant with height, except within the lowest 10% of the CBL. This layer is called the *surface layer* (SL).
- Transfer of momentum, mass, and moisture within this layer is able to overcome mixing effects (see Fig. 2).
- Consequently, $\overline{\theta}$ may decrease by 1-2 K from the ground to the top of the SL (*i.e.*, super-adiabatic conditions). In addition, moisture ($\overline{q}$) may also decrease with height within this layer, while wind speed ($\overline{u}$) logarithmically approaches zero from the top of the layer to the ground.
- Turbulent fluxes generally have their largest magnitude at the surface and decrease in magnitude with height, where they become negligible at the base of the free atmosphere (see Fig. 2).
- An exception to this is the moisture flux ($\overline{w'q'}$), which can have a maximum magnitude at a level significantly above the surface (it is still negligible at the base of the free atmosphere).
- Since ($\overline{u}$) generally increases with height, the momentum flux ($\overline{w'u'}$) tends to be negative.
- Thus, rising thermals are associated with negative velocity perturbations as they carry lower momentum upward. Conversely, the downward motions are associated with positive velocity perturbations as they bring higher momentum downward.
- $\overline{w'u'}$ tends to decease linearly with height from the surface to the top of the CBL. The magnitude decreases rapidly from the base of the CBL to a value near zero at the top of the EZ.
- Since $\overline{w'u'}$ is negative at the surface and the magnitude decreases with height, $\partial(\overline{w'u'})/\partial z > 0$. This implies a net drag on the mean winds within the boundary layer for $\overline{u} > 0$.
- The kinematic heat flux ($\overline{w'\theta'}$) is generally positive in the CBL as a result of rising air being positively buoyant.
- ($\overline{w'\theta'}$) generally decreases linearly with height from the surface to the top of the CBL.
- Within the EZ, $\overline{w'\theta'}$ is generally negative and $\sim$ 10-20$\%$ of its surface magnitude.
- Under these conditions, $\partial(\overline{w'\theta'})/\partial z < 0$ in the CBL, which implies mean warming.
- Conversely, $\partial(\overline{w'\theta'})/\partial z > 0$ in the EZ, which points to mean cooling. Penetrating thermals are responsible for cooling this layer.
- $\overline{w'q'}$ also changes linearly with height, though the change can either be negative or positive (see Fig. 2) depending on whether there is drying or moistening of the CBL.
- The sign of $\partial(\overline{w'q'})/\partial z$ largely depends on the surface moisture characteristics.
- $\partial(\overline{w'q'})/\partial z < 0$ in the EZ, which implies net moistening in the layer.

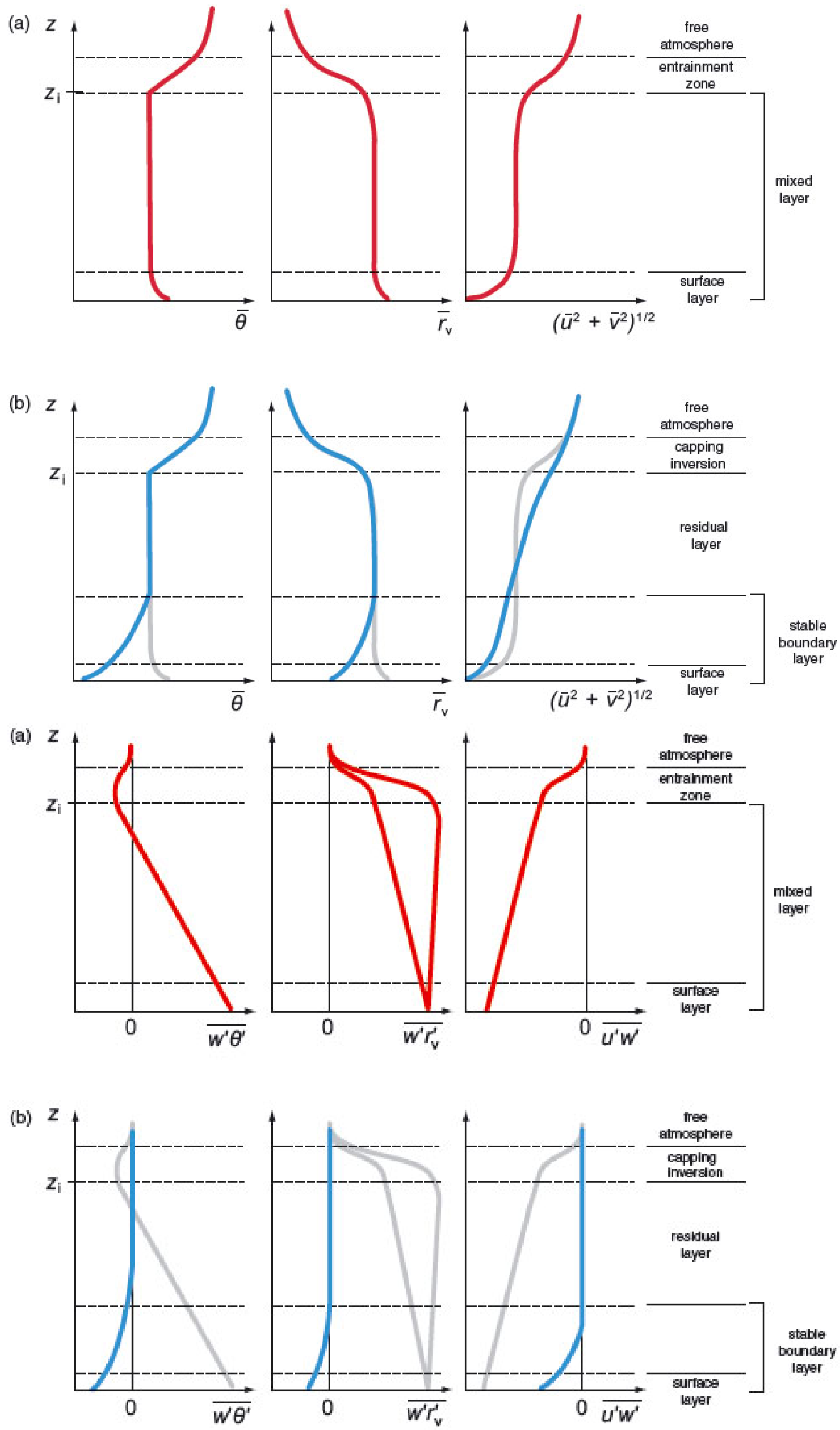

Figure 2: Typical vertical profiles of mean potential temperature, water vapor mixing ratio, horizontal wind speed (upper panels), vertical turbulent kinematic heat flux, moisture flux, and momentum flux (lower panels; it is assumed that $\overline{u} > 0$) during (a) daytime and (b) nighttime. The daytime profiles also appear in (b) in gray to facilitate day–night comparisons. [From Markowski and Richardson]

3.2.5 Nighttime Boundary Layer Structure

- A shallow inversion layer forms near the surface at night due to heat loss by the ground (Fig. 2).
- This layer is called the *stable boundary layer* (SBL) or *nocturnal boundary layer* (NBL).
- The SBL becomes *decoupled* from the ML because the inversion inhibits mixing.
- The decoupled "leftover" ML from the daytime CBL is called the *residual layer* (RL).
- The RL is characterized by nearly constant $\overline{\theta}$ and $\overline{q}$.
- The daytime's EZ is referred to as the capping inversion at night since there is not much exchange between the free atmosphere and the residual layer.
- Turbulent fluxes are generally negative in the SBL (see Fig. 2)).
- The maximum magnitude of each turbulent flux is located at the surface, while each flux becomes negligible at the top of the layer.
- Put another way, $\partial(\overline{w'u'})/\partial z$, $\partial(\overline{w'\theta'})/\partial z$, and $\partial(\overline{w'q'})/\partial z$ are all > 0.
- Thus, the SBL is characterized at night by cooling and drying, while acting as an effective drag on the mean wind.
- The shallow inversion layer grows in steps, which are often interrupted by intermittent turbulent events. These events are the result of small, mechanically-generated eddies.
- These intermittent turbulence events act to deepen the SBL and reduce its stability.
- Relative humidity increases in the SBL, forming dew, which may act to reduce $\overline{q}$ through condensation.
- Radiational cooling is strongest on nights with weak winds and infrequent turbulence events. On these nights, the inversion is rather strong and shallow, leading to especially low near-surface air temperature.
- The depth of the RL compared to the depth of the SBL is largely a function of mean wind shear.
- The SBL can be as shallow as $\sim 10\ \mathrm{m}$ on clear nights with minimal winds.
- Conversely, the SBL depth may be of the same order as the daytime CBL on cloudy nights with strong winds.
- $\overline{q}$ at the surface can either increase or decrease after sunset depending on the degree of dew formation versus the degree to which moisture was reduced in the daytime SL due to mixing.
- Winds at the top of the SBL and within the RL accelerate at night as surface drag is effectively eliminated.
- As a result, a nocturnal low-level wind maximum forms. This is often called the *low-level jet*. We will discuss the dynamics of the low-level jet in an upcoming section.
- At the surface, negatively buoyant air tends to sink toward lower elevations. This results in so-called *drainage winds*, which are cold-air-runoff winds that are produced when air in contact with terrain surfaces is cooled and flows downslope (katabatic) and/or downvalley.
- Drainage winds also refer to gravity winds that drain cold air into frost hollows, river valleys, and other lower-lying terrain.

3.3 The Nature of Turbulent Fluxes

Before considering specific boundary layer phenomena, we need to develop a set of equations that properly account for turbulent motions.

3.3.1 Reynolds-Averaging

Consider the time series plot of the wind speed shown in Fig. 3. The trace contains many high-frequency (fast) fluctuations. Such fluctuations are due to small turbulent eddies, which do not reliably represent the mean flow.

To obtain a wind speed measurement representative of the large-scale flow, one must obtain take an average over a time period long enough to smooth over the fluctuations but still short enough for keep the trend. Such averaging was first proposed by *Reynolds*, and is therefore named after him.

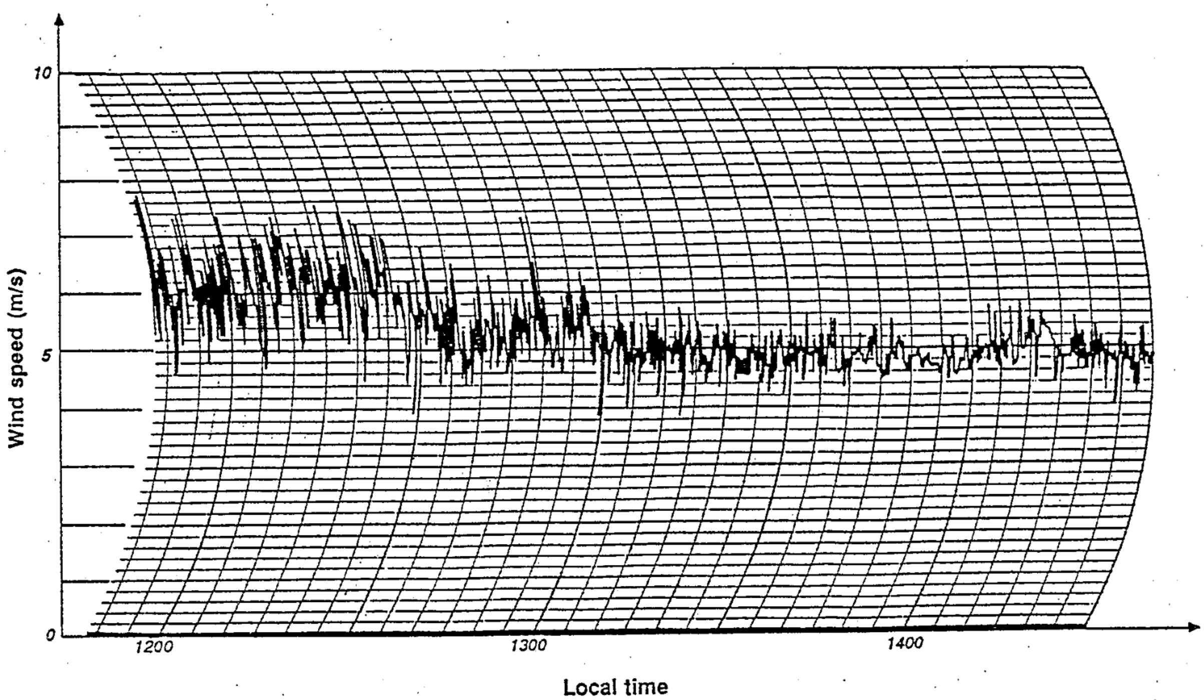

Figure 3: Trace of wind speed observed in early afternoon. [From Stull 1988]

Consider two variables A and B. We will define each variable such as comprising a slowly varying mean ($\overline{A}$ and $\overline{B}$) and a rapidly fluctuating perturbation that is associated with a turbulent eddy ($A^{'}$ and $B^{'}$):

$$A = \overline{A} + A^{'} \tag{1}$$

$$B = \overline{B} + B^{'} \, . \tag{2}$$

The $\overline{(\;)}$ operator represents an average in time or space, *i.e.*,

$$\overline{A} = \frac{1}{N}\sum_{i=1}^{N} A_i \;, \tag{3}$$

where A_i is the ith of N observations of A.

Specifically,

$$\overline{(A)} = \overline{(\overline{A} + A')} = \overline{\overline{A}} + \overline{A'} = \overline{A} + \overline{A'} \;. \tag{4}$$

There are two immediate takeaways from the above expression. First,

$$\overline{\overline{A}} = \overline{A} \tag{5}$$

because it is simply the average of the average. If we assume a homogeneous and stationary flow, then the average is not dependent on space or time. Thus, further averaging has no effect. Second,

$$\overline{A'} = 0 \tag{6}$$

by definition. This means that the area above and below a mean line are equal. Be careful, though. The average of the product of these turbulent fluctuations is not necessarily zero. That is,

$$\overline{A'B'} \neq 0 \;. \tag{7}$$

As an example, consider $\overline{w'\theta'}$ in the daytime CBL. As previously discussed, $w^{'}$ and $\theta^{'}$ are positively correlated, which means that $\overline{w'\theta'} > 0$. This occurs because rising thermals are associated with upward motion ($w^{'} > 0$) of positively buoyant ($\theta^{'} > 0$) air and the compensating return flow is associated with the downward motion ($w^{'} < 0$) of negatively buoyant ($\theta^{'} < 0$) air.

The average of the sum of two variables is given by

$$\overline{(A + B)} = \overline{A} + \overline{B} \;. \tag{8}$$

The average of the product of a constant and a variable is given by

$$\overline{cA} = c\overline{A} \;. \tag{9}$$

Here is another common rule applied in Reynolds averaging

$$\overline{(\overline{A}B)} = \overline{A} + \overline{B} \;, \tag{10}$$

where the averaged A essentially acts as a constant based on the previous rule that $\overline{\overline{A}} = \overline{A}$.

It can also be shown that

$$\overline{\frac{\partial A}{\partial t}} = \frac{\partial \overline{A}}{\partial t} \quad \text{and} \quad \overline{\frac{\partial A}{\partial x}} = \frac{\partial \overline{A}}{\partial x} \; . \tag{11}$$

Let us examine the product of two variables by applying the previous rules:

$$\begin{aligned}
\overline{AB} &= \overline{(\overline{A} + A')(\overline{B} + B')} \\
&= \overline{\overline{A}\overline{B} + \overline{A}B' + A'\overline{B} + A'B'} \\
&= \overline{\overline{A}\overline{B}} + \overline{\overline{A}B'} + \overline{A'\overline{B}} + \overline{A'B'} \\
&= \overline{A}\,\overline{B} + \overline{A'B'} \; .
\end{aligned} \tag{12}$$

In short, the average of the product of two variables is the product of the mean components of each variable plus the average of the product of the fluctuating components.

In the above, we see another averaging rule

$$\overline{(A'\overline{B})} = \overline{A'}\,\overline{B} = 0 \cdot \overline{B} = 0 \; . \tag{13}$$

We call AB a *nonlinear* term since both are time or space dependent variables. $A^{'}B^{'}$ is also a nonlinear term whose average is not necessarily zero.

$\overline{A'B'}$ is called the *covariance* of A and B. If $A^{'} = B^{'}$, the term $\overline{A'^2}$ is called the *variance*. Terms like $\overline{A'B'}$, $\overline{A'^2}$, and $\overline{B'^2}$ are called *second-order moments*.

These second moments are extremely important for boundary layer studies. They illustrate how the Reynolds averaging procedure statistically accounts for the effect of turbulent eddies on the mean field.

Below is a summary of the rules for Reynolds averaging

- $\overline{\overline{A}} = \overline{A}$
- $\overline{cA} = c\overline{A}$
- $\overline{A'} = 0$
- $\overline{A'\overline{B}} = \overline{\overline{A}B'} = 0$
- $\overline{A + B} = \overline{A} + \overline{B}$
- $\overline{AB} = \overline{A}\,\overline{B} + \overline{A'B'}$
- $\overline{\dfrac{\partial A}{\partial t}} = \dfrac{\partial \overline{A}}{\partial t}$
- $\overline{\dfrac{\partial A}{\partial x}} = \dfrac{\partial \overline{A}}{\partial x}$