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An Introduction to

Boundary
Layer
Meteorology

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Preface

Part of the excitement of boundary-layer meteorology is the challenge in studying and understanding turbulent flow — one of the unsolved problems of classical physics. Additional excitement stems from the rich diversity of topics and research methods that we collect under the umbrella of boundary-layer meteorology. That we live our lives within the boundary layer makes it a subject that touches us, and allows us to touch it. I've tried to capture some of the excitement, challenges and diversity within this book.

I wrote this book with a variety of goals in mind. First and foremost, this book is designed as a *textbook*. Fundamental concepts and mathematics are presented prior to their use, physical interpretations of equation terms are given, sample data is shown, examples are solved, and exercises are included. Second, the book is organized as a *reference*, with tables of parameterizations, procedures, field experiments, useful constants, and graphs of various phenomena in a variety of conditions. Third, the last several chapters are presented as a *literature review* of the current ideas and methods in boundary layer meteorology.

It is assumed that this book will be used at the beginning graduate level for students with an undergraduate background in meteorology. However, a diversity in the background of the readers is anticipated. Those with a strong mathematical background can skip portions of Chapter 2 on statistics, and those with experience with time series can skip Chapter 8. These two mathematics chapters were separated to offer the reader a chance to apply the first dose of statistics to boundary layer applications before delving back into more math. Some students might have had a course on geophysical turbulence or statistical fluid mechanics, and can skim through the first 5 chapters to get to the boundary-layer applications. By excluding a few chapters, instructors can easily fit the remaining material into a one-semester course. With supplemental readings, the book can serve as a two-semester sequence in atmospheric turbulence and boundary-layer meteorology.

Notational diversity proved to be the greatest difficulty. Each subdiscipline appeared to have its own set of notation, which often conflicted with the notation of other subdisciplines. To use the published notation would have lead to confusion. I was therefore forced to select a consistent set of notation to use throughout the book. In most cases I've tried to retain the notation most frequently used in the literature, or to add subscripts or make logical extensions to existing notation. In other cases, I had to depart from previously published notation. Readers are referred to Appendix B for a comprehensive list of notation.

I certainly cannot claim to be an expert in all the myriad subdisciplines of boundary-layer meteorology, yet I knew the book should be comprehensive to be useful. My interest and enthusiasm in writing this book motivated many trips to the library, and stimulated my analysis of many research papers to learn the underlying themes and common tools used in the diverse areas of boundary layer meteorology. Unfortunately, the limited space within this book necessitated some difficult decisions regarding the amount and level of material to include. Hopefully I've presented sufficient background to lay the building blocks upon which more advanced concepts can be built by other instructors and researchers. Certain topics such as atmospheric dispersion and agricultural micrometeorology are not covered here, because there are other excellent books on these subjects.

Many colleagues and friends helped with this book and contributed significantly to its final form and quality. Michelle Vandall deserves special recognition and thanks for drawing most of the figures, designing the page and chapter headers, and for her overall dedication to this project. Colleagues Steve Stage, Larry Mahrt, George Young, Jacq Schols, Chandran Kaimal, Steve Silberberg, Beth Ebert, and Bruce Albrecht reviewed various chapters and provided valuable suggestions. Information regarding field experiments was provided by Anton Beljaars, Ad Driedonks, Jean-Claude André, Jean-Paul Goutorbe, Anne Jochum, Steve Nelson, Bob Murphy, Ruwim Berkowicz, Peter Hildebrand, Don Lenschow, and others. Eric Nelson proofread the manuscript, and helped with the list of notation. Sam Zhang compiled the index. Some of the equations were typeset by Camille Riner and Michelle Vandall (her name keeps reappearing). Four years of students in my micrometeorology courses at Wisconsin graciously tolerated various unfinished drafts of the book, and caught many mistakes. The patient editors of Kluwer Academic Publishers (formerly D. Reidel) provided constant encouragement, and are to be congratulated for their foresight and advice. The American Meteorological Society is acknowledged for permission to reproduce figures 10.7, 12.10 and 12.19. To all of these people and the many more to whom I apologize for not naming here, I thank you.

To my wife, Linda, I give my gratitude for her devotion during this exciting and exhausting episode of my life.

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(a)

Z (m)

1000

500

0

1000

500

0

— 5

Range from lidar (km)

— 10

(b)

Z (m)

500

250

0

500

250

0

— 5

Range from lidar (km)

10
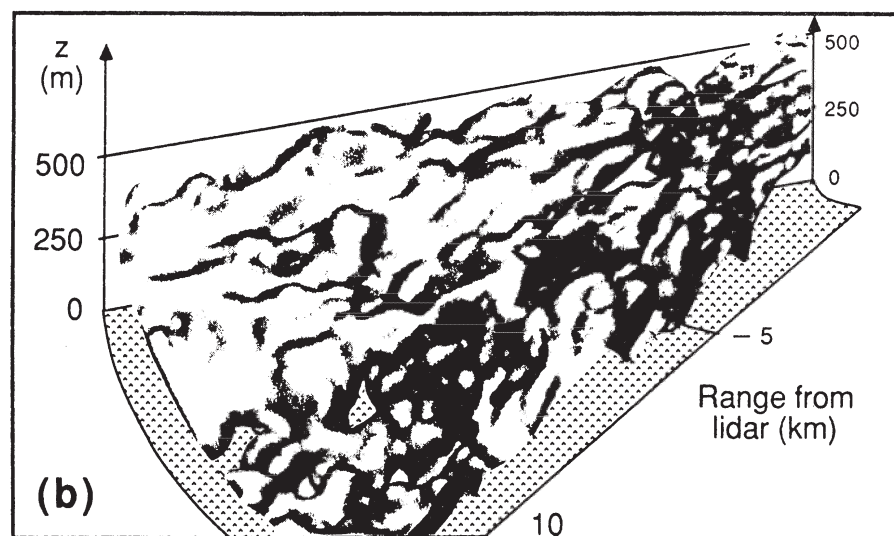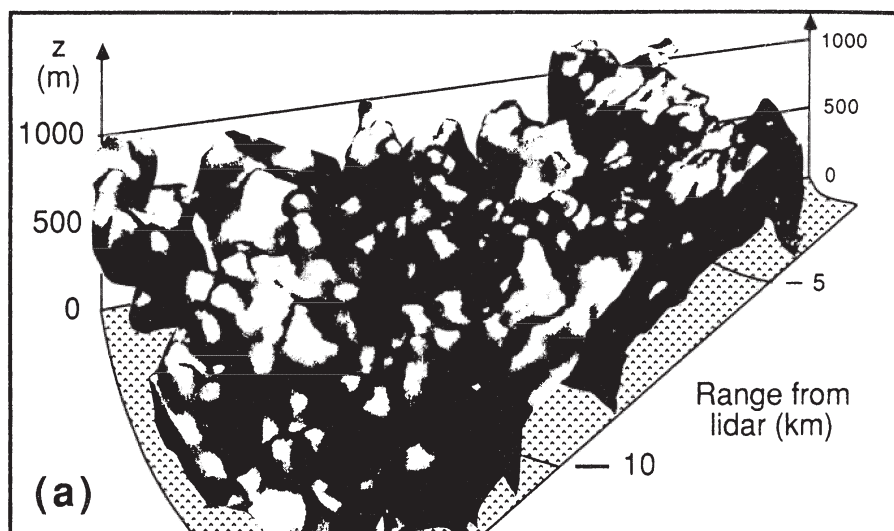
Frontispiece Lidar images of the aerosol-laden boundary layer, obtained during the FIFE field experiment in Kansas. (a) Convective mixed layer observed at 1030 local time on 1 July 1987, when winds were generally less than 2 m/s. (b) Slightly-stable boundary layer with shear-generated turbulence, observed at 530 local time on 7 July 1987. Winds ranged from 5 m/s near the surface to 15 m/s near the top of the boundary layer. Photographs from the Univ. of Wisconsin lidar are courtesy of E. Eloranta, Boundary Layer Research Team.

1 Mean Boundary Layer Characteristics

From our first breath, we spend most of our lives near the earth's surface. We feel the warmth of the daytime sun and the chill of the nighttime air. It is here where our crops are grown, our dwellings are constructed, and much of our commerce takes place. We grow familiar with our local breezes and microclimates, and we sense the contrasts when we travel to other places.

Such near-earth characteristics, however, are not typical of what we observe in the rest of the atmosphere. One reason for this difference is the dominating influence of the earth on the lowest layers of air.

The earth's surface is a boundary on the domain of the atmosphere. Transport processes at this boundary modify the lowest 100 to 3000 m of the atmosphere, creating what is called the **boundary layer** (Fig 1.1). The remainder of the air in the troposphere is loosely called the *free atmosphere*. The nature of the atmosphere as perceived by most individuals is thus based on the rather peculiar characteristics found in a relatively shallow portion of the air.
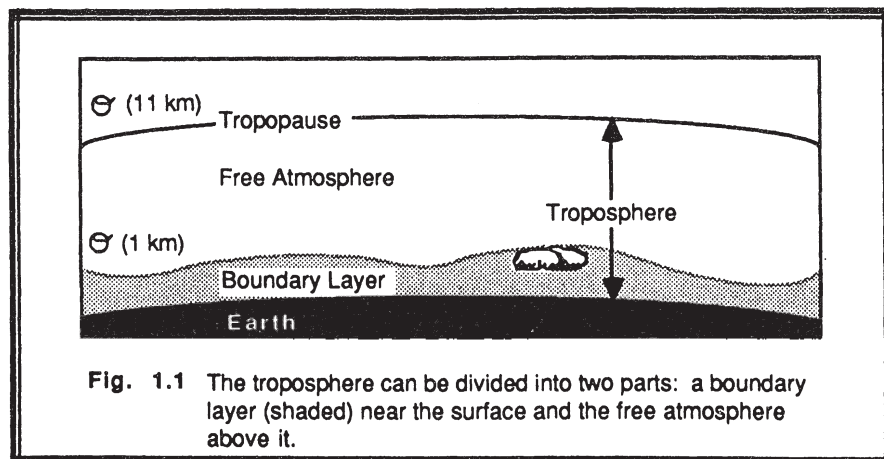
Fig. 1.1 The troposphere can be divided into two parts: a boundary layer (shaded) near the surface and the free atmosphere above it.

1

This chapter is meant to provide a descriptive overview of the boundary layer. Emphasis is placed on mid-latitude boundary layers over land, because that is where much of the world's population resides and where many boundary layer measurements have been made. In this region the diurnal cycle and the passage of cyclones are the dominant forcing mechanisms. Tropical and maritime boundary layers will also be briefly reviewed. Theories and equations presented in later chapters will make more sense when put into the context of the descriptive nature of the boundary layer as surveyed here.

1.1 A Boundary-Layer Definition

The troposphere extends from the ground up to an average altitude of 11 km, but often only the lowest couple kilometers are directly modified by the underlying surface. We can *define the boundary layer as that part of the troposphere that is directly influenced by the presence of the earth's surface, and responds to surface forcings with a timescale of about an hour or less.* These forcings include frictional drag, evaporation and transpiration, heat transfer, pollutant emission, and terrain induced flow modification. The boundary layer thickness is quite variable in time and space, ranging from hundreds of meters to a few kilometers.

An example of temperature variations in the lower troposphere is shown in Fig 1.2. These time-histories were constructed from rawinsonde soundings made every several hours near Lawton, Oklahoma. They show a *diurnal variation* of temperature near the ground that is not evident at greater altitudes. Such diurnal variation is one of the key characteristics of the boundary layer over land. The free atmosphere shows little diurnal variation.

Fig. 1.2
Evolution of temperatures measured near the ground (97.5 kPa) and at a height of roughly 1100 m above ground (85 kPa). Based on rawinsonde lauches from Ft. Sill, OK.
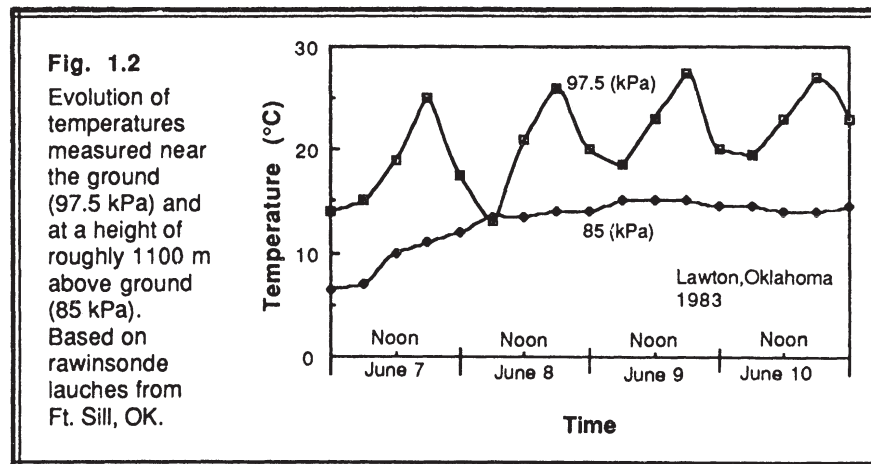
This diurnal variation is not caused by direct forcing of solar radiation on the boundary layer. Little solar radiation is absorbed in the boundary layer; most is transmitted to the ground where typical absorptivities on the order of 90% result in absorption of much of

the solar energy. It is the ground that warms and cools in response to the radiation, which in turn forces changes in the boundary layer via transport processes. *Turbulence* is one of the important transport processes, and is sometimes also used to define the boundary layer.

Indirectly, the whole troposphere can change in response to surface characteristics, but this response is relatively slow outside of the boundary layer. Hence, our definition of the boundary layer includes a statement about one-hour time scales. This does not imply that the boundary layer reaches an equilibrium in that time, just that alterations have at least begun.

Two types of clouds are often included in boundary-layer studies. One is the *fair-weather cumulus cloud*. It is so closely tied to thermals in the boundary layer that it is difficult to study the dynamics of this cloud type without focusing on the triggering boundary-layer mechanisms. The other type is the *stratocumulus cloud*. It fills the upper portion of a well-mixed, humid boundary layer where cooler temperatures allow condensation of water vapor. *Fog*, a stratocumulus cloud that touches the ground, is also a boundary-layer phenomenon.

Thunderstorms, while not a surface forcing, can modify the boundary layer in a matter of minutes by drawing up boundary-layer air into the cloud, or by laying down a carpet of cold downdraft air. Although thunderstorms are rarely considered to be boundary layer phenomena, their interaction with the boundary layer will be reviewed in this book.

1.2 Wind and Flow

Air flow, or wind, can be divided into three broad categories: *mean wind*, *turbulence*, and *waves* (Fig 1.3). Each can exist separately, or in the presence of any of the others. Each can exist in the boundary layer, where transport of quantities such as moisture, heat, momentum, and pollutants is dominated in the horizontal by the mean wind, and in the vertical by turbulence.

Fig. 1.3 Idealization of (a) Mean wind alone, (b) waves alone, and (c) turbulence alone. In reality waves or turbulence are often super-imposed on a mean wind. U is the component of wind in the x-direction.
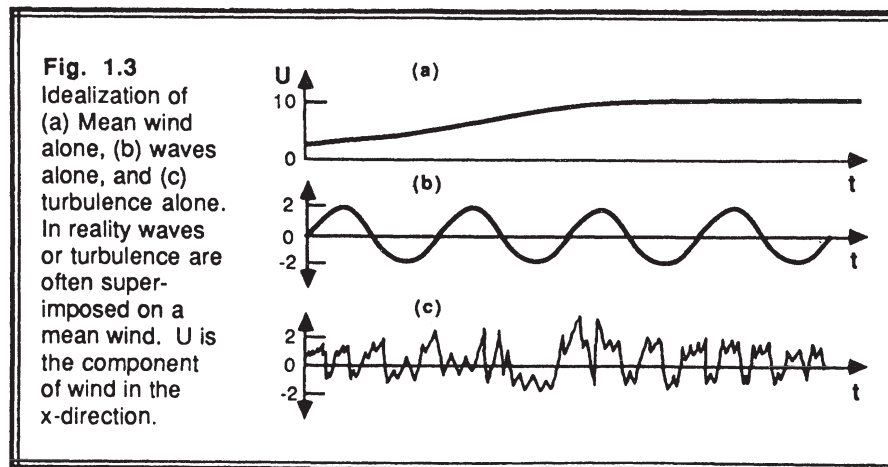
Mean wind is responsible for very rapid horizontal transport, or *advection*. Horizontal winds on the order of 2 to 10 m/s are common in the boundary layer. Friction causes the mean wind speed to be slowest near the ground. Vertical mean winds are much smaller, usually on the order of millimeters to centimeters per second.

Waves, which are frequently observed in the nighttime boundary layer, transport little heat, humidity, and other scalars such as pollutants. They are, however, effective at transporting momentum and energy. These waves can be generated locally by mean-wind shears and by mean flow over obstacles. Waves can also propagate from some distant source, such as a thunderstorm or an explosion.

The relatively high frequency of occurrence of turbulence near the ground is one of the characteristics that makes the boundary layer different from the rest of the atmosphere. Outside of the boundary layer, turbulence is primarily found in convective clouds, and near the jet stream where strong wind shears can create clear air turbulence (CAT).

Sometimes atmospheric waves may enhance the wind shears in localized regions, causing turbulence to form. Thus, wave phenomena can be associated with the turbulent transport of heat and pollutants, although waves without turbulence would not be as effective.

A common approach for studying either turbulence or waves is to split variables such as temperature and wind into a *mean part* and a *perturbation part*. The mean part represents the effects of the mean temperature and mean wind, while the perturbation part can represent either the wave effect or the turbulence effect that is superimposed on the mean wind.

As will be seen in later chapters of this book, such a splitting technique can be applied to the equations of motion, creating a number of new terms. Some of these terms, consisting of products of perturbation variables, describe *nonlinear* interactions between variables and are associated with turbulence. These terms are usually neglected when wave motions are of primary interest. Other terms, containing only one perturbation variable, describe *linear* motions that are associated with waves. These terms are neglected when turbulence is emphasized.

1.3 Turbulent Transport

Turbulence, the gustiness superimposed on the mean wind, can be visualized as consisting of irregular swirls of motion called *eddies*. Usually turbulence consists of many different size eddies superimposed on each other. The relative strengths of these different scale eddies define the *turbulence spectrum*.

Much of the boundary layer turbulence is generated by forcings from the ground. For example, *solar heating* of the ground during sunny days causes *thermals* of warmer air to rise. These thermals are just large eddies. *Frictional drag* on the air flowing over the ground causes *wind shears* to develop, which frequently become turbulent. *Obstacles* like trees and buildings deflect the flow, causing *turbulent wakes* adjacent to, and downwind of the obstacle.

The largest boundary layer eddies scale to (ie, have sizes roughly equal to) the depth of the boundary layer; that is, 100 to 3000 m in diameter. These are the most intense

eddies because they are produced directly by the forcings discussed above. "Cats paws" on lake surfaces and looping smoke plumes provide evidence of the larger eddies.

ꞌ Smaller size eddies are apparent in the swirls of leaves and in the wavy motions of the grass. These eddies feed on the larger ones. The smallest eddies, on the order of a few millimeters in size, are very weak because of the dissipating effects of molecular viscosity.

Turbulence is several orders of magnitude more effective at transporting quantities than is molecular diffusivity. It is turbulence that allows the boundary layer to respond to changing surface forcings. The frequent lack of turbulence above the boundary layer means that the rest of the free atmosphere cannot respond to surface changes. Stated more directly, the free atmosphere behaves as if there were no boundary to contend with, except in sense of mean wind flowing over the boundary-layer-top height contours.

1.4 Taylor's Hypothesis

We often need information on the size of eddies and on the scales of motions in the boundary layer. Unfortunately, it is difficult to create a snapshot picture of the boundary layer. Instead of observing a large region of space at an instant in time, we find it easier to make measurements at one point in space over a long time period. For example, meteorological instruments mounted on a tower can give us a time record of the boundary layer as it blows past our sensors.

In 1938, G. I. Taylor suggested that for some special cases, turbulence might be considered to be *frozen* as it advects past a sensor. Thus, the wind speed could be used to translate turbulence measurements as a function of time to their corresponding measurements in space. We must keep in mind that turbulence is not really frozen. Taylor's simplification is thus useful for only those cases where the turbulent eddies evolve with a timescale longer than the time it takes the eddy to be advected past a sensor (Powell and Elderkin, 1974).

Let U and V represent the eastward-moving and northward-moving Cartesian wind components, and let M represent the total wind magnitude (speed) given by $M^2 = U^2 + V^2$. If an eddy of diameter λ is advected at mean wind speed M, then the time period $\mathbb{P}$ for it to pass by a stationary sensor is given by $\mathbb{P} = \lambda / M$.

Suppose that some variable like temperature varies from one side of the eddy to the other. Then the temperature measured at our sensors would vary with time as the eddy advects past. For example, Fig 1.4a shows an initial condition where an eddy of 100 m in diameter is beginning to advect past our sensor tower. The leading side of the eddy, at a temperature of 10°C, is warmer than the trailing side, at only 5°C. This is our spatial description of the eddy at an instant in time. Namely, the temperature gradient across the eddy is $\partial T/\partial x_d = 0.05$ K/m, where x_d is in a direction parallel to the mean wind.

At that initial instant, our sensor on the tower measures a temperature of 10°C. If the wind speed were 10 m/s, then the sensor would measure a temperature of 5°C ten seconds later, assuming that the eddy has not changed as it advected by (Fig 1.4b). The local change of temperature with time measured at our sensor is $\partial T/\partial t = -0.5$ K/s. We see that:
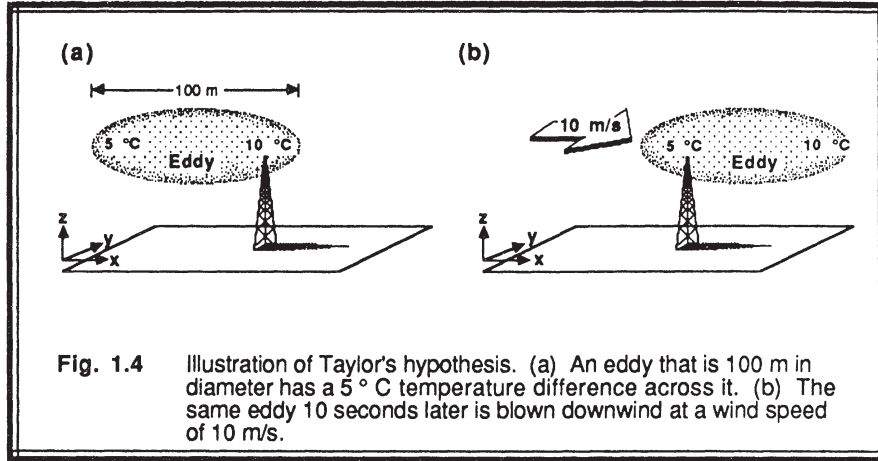
Fig. 1.4   Illustration of Taylor's hypothesis. (a) An eddy that is 100 m in diameter has a 5 ° C temperature difference across it. (b) The same eddy 10 seconds later is blown downwind at a wind speed of 10 m/s.

$$\frac{\partial T}{\partial t} = -M \frac{\partial T}{\partial x_d} \qquad (1.4a)$$

which is an expression of Taylor's hypothesis for temperature in one dimension.

For any variable ξ, Taylor's hypothesis states that turbulence is frozen when $d\xi/dt = 0$. But the total derivative is defined by: $d\xi/dt = \partial\xi/\partial t + U \partial\xi/\partial x + V \partial\xi/\partial y + W \xi/\partial z$. Thus, the general form of Taylor's hypothesis is

$$\frac{\partial \xi}{\partial t} = -U \frac{\partial \xi}{\partial x} - V \frac{\partial \xi}{\partial y} - W \frac{\partial \xi}{\partial z} \qquad (1.4b)$$

This hypothesis can also be stated in terms of a wavenumber, κ, and frequency, f:

$$\kappa = f / M \qquad (1.4c)$$

where $\kappa = 2\pi/\lambda$, and $f = 2\pi/P$, for wavelength λ and wave period P (Wyngaard and Clifford, 1977). The dimensions of κ are radians per unit length, while f has dimensions of radians per unit time.

To satisfy the requirements that the eddy have negligible change as it advects past a sensor, Willis and Deardorff (1976) suggest that

$$\sigma_M < 0.5 M \qquad (1.4d)$$

where σ_M, the standard deviation (see chapter 2 for a review of statistics) of wind speed,

is a measure of the intensity of turbulence. Thus, Taylor's hypothesis should be satisfactory when the turbulence intensity is small relative to the mean wind speed.

1.5 Virtual Potential Temperature

Buoyancy is one of the driving forces for turbulence in the BL. Thermals of warm air rise because they are less dense than the surrounding air, and hence *positively buoyant*. *Virtual temperature* is a popular variable for these studies because it is the temperature that dry air must have to equal the density of moist air at the same pressure. Thus, variations of virtual temperature can be studied in place of variations in density.

Water vapor is less dense than dry air; thus, moist unsaturated air is more buoyant than dry air of the same temperature. The virtual temperature of unsaturated moist air is therefore always greater than the absolute air temperature, T. Liquid water, however, is more dense than dry air, making cloudy air heavier or less buoyant than the corresponding cloud-free air. The suspension of cloud droplets in an air parcel is called *liquid water loading*, and it always reduces the virtual temperature.

Virtual potential temperatures are analogous to potential temperatures in that they remove the temperature variation caused by changes in pressure altitude of an air parcel. Turbulence includes vertical movement of air, making a variable such as virtual potential temperature not just attractive but almost necessary.

1.5.1   Definitions

For saturated (cloudy) air, the virtual potential temperature, θ_v, is defined by:

$$\theta_v = \theta \cdot (1 + 0.61 \cdot r_{sat} - r_L) \tag{1.5.1a}$$

where r_{sat} is the water-vapor saturation mixing ratio of the air parcel, and r_L is the liquid-water mixing ratio. In (1.5.1a) the potential temperatures are in units of K, and the mixing ratios are in units of g/g. For unsaturated air with mixing ratio r, the virtual potential temperature is:

$$\theta_v = \theta \cdot (1 + 0.61 \cdot r) \tag{1.5.1b}$$

A derivation of the virtual temperature is given in Appendix D.

As usual, the potential temperature, θ, is defined as

$$\theta = T \left(\frac{P_o}{P} \right)^{0.286} \tag{1.5.1c}$$

where P is air pressure and P_o is a reference pressure. Usually, P_o is set to 100 kPa

(1000 mb), but sometimes for boundary layer work the surface pressure is used instead.
    To first order, we can approximate the potential temperature by

$$\theta \cong T + (g/C_p) \cdot z \qquad (1.5.1d)$$

where z is the height above the 100 kPa (1000 mb) level, although sometimes height above ground level (agl) is used instead. The quantity $g/C_p = 0.0098$ K/m is just the negative of the dry adiabatic lapse rate (9.8 °C per kilometer), where g is the gravitational acceleration and C_p is the specific heat at constant pressure for air. Sometimes the quantity $C_p \cdot \theta$ is called the *dry static energy*.

Fig. 1.5
Example of the difference between mean potential temperature, $\bar{\theta}$, and mean virtual potential temperature, $\bar{\theta}_v$, given observations of mixing ratio, $\bar{r}$, and absolute temperature, $\bar{T}$. Dew point, $\bar{T}_d$, is also shown.
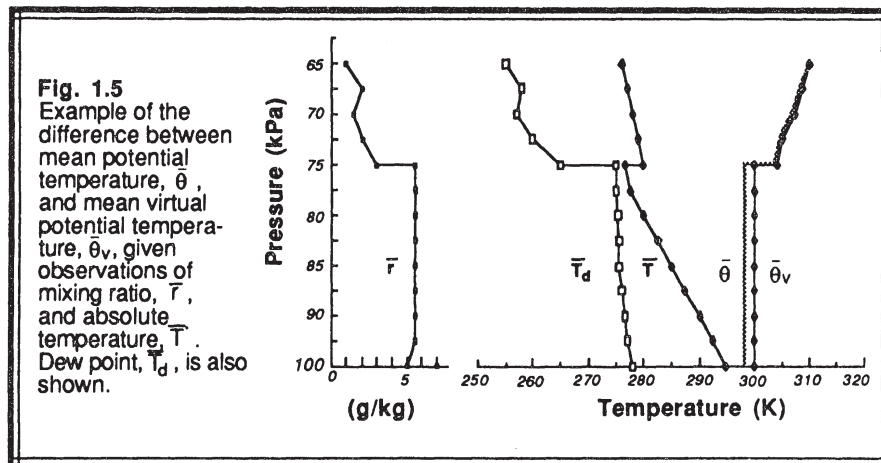
An example of the difference between potential temperature and virtual potential temperature is shown in Fig 1.5 for a case of moist unsaturated air. The difference is small, but not negligible. Only when the air is very dry can we neglect the difference.

1.5.2   Example

Problem. Given a temperature of 25°C and a mixing ratio of 20 g/kg measured at a pressure of 90 kPa (900 mb), find the virtual potential temperature.

Solution. First, we must find the potential temperature:

$$\theta = T (P_o/P)^{0.286} = 298.16 \cdot (100/90)^{0.286} = 307.28 \text{ K}$$

The air is unsaturated. allowing us to find the virtual potential temperature from:

$$\theta_v = \theta \cdot (1 + 0.61 \cdot r) = 307.28 \cdot [1 + 0.61 \cdot (0.020)] = 311.03 \text{ K}$$

Discussion. Even though the virtual potential temperature is only about 4 K warmer than the potential temperature, this difference is on the same order as the difference between the warm air rising in thermals and the surrounding environment. Thus, neglect of the humidity in buoyancy calculations could lead to erroneous conclusions regarding convection and turbulence.

1.6 Boundary Layer Depth and Structure

Over oceans, the boundary layer depth varies relatively slowly in space and time. The sea surface temperature changes little over a diurnal cycle because of the tremendous mixing within the top of the ocean. Also, water has a large heat capacity, meaning that it can absorb large amounts of heat from the sun with relatively little temperature change. Thus, a slowly varying sea surface temperature means a slowly varying forcing into the bottom of the boundary layer.

Most changes in boundary layer depth over oceans are caused by synoptic and mesoscale processes of vertical motion and advection of different air masses over the sea surface. An air mass with a temperature different than that of the ocean will undergo a modification as its temperature equilibrates with that of the sea surface. Once equilibrium is reached, the resulting boundary layer depth might vary by only 10% over a horizontal distance of 1000 km. Exceptions to this gentle variation can occur near the borders between two ocean currents of different temperatures (Stage and Weller, 1976).

Fig. 1.6   Schematic of synoptic - scale variation of boundary layer depth between centers of surface high (H) and low (L) pressure. The dotted line shows the maximum height reached by surface modified air during a one-hour period. The solid line encloses the shaded region, which is most studied by boudary-layer meteorlogists.
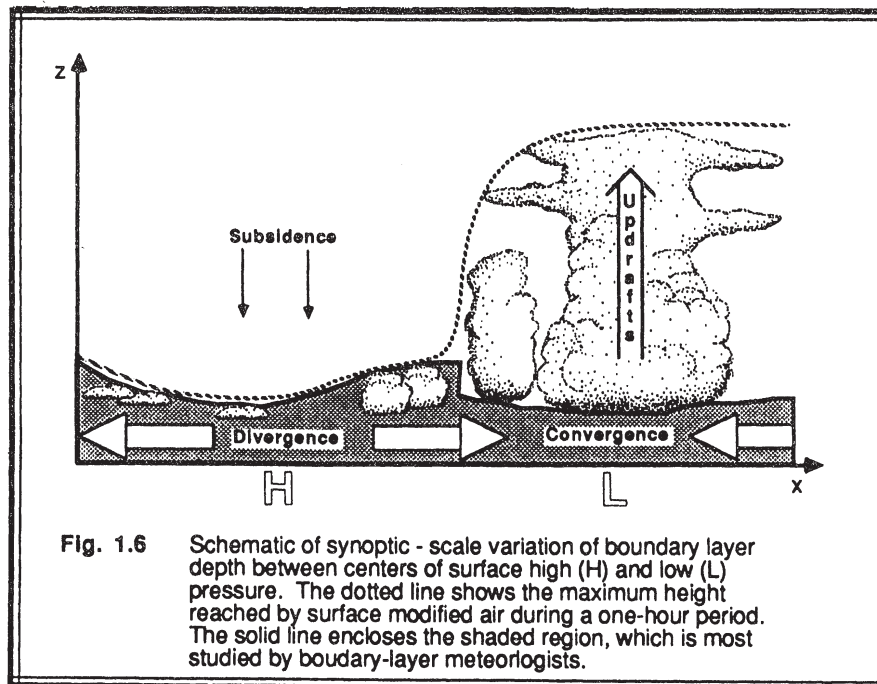
Over both land and oceans, the general nature of the boundary layer is to be thinner in high-pressure regions than in low-pressure regions (Fig 1.6). The subsidence and low-level horizontal divergence associated with synoptic high pressure moves boundary layer air out of the high towards lower pressure regions. The shallower depths are often associated with cloud-free regions. If clouds are present, they are often fair-weather cumulus or stratocumulus clouds.

In low pressure regions the upward motions carry boundary-layer air away from the ground to large altitudes throughout the troposphere. It is difficult to define a boundary-layer top for these situations. Cloud base is often used as an arbitrary cut-off for boundary layer studies in these cases. Thus, the region studied by boundary layer meteorologists may actually be thinner in low-pressure regions than in high pressure ones (see Fig 1.6).

Over land surfaces in high pressure regions the boundary layer has a well defined structure that evolves with the diurnal cycle (Fig 1.7). The three major components of this structure are the *mixed layer*, the *residual layer*, and the *stable boundary layer*. When clouds are present in the mixed layer, it is further subdivided into a *cloud layer* and a *subcloud layer*.

The *surface layer* is the region at the bottom of the boundary layer where turbulent fluxes and stress vary by less than 10% of their magnitude. Thus, the bottom 10% of the boundary layer is called the surface layer, regardless of whether it is part of a mixed layer or stable boundary layer. Finally, a thin layer called a *microlayer* or *interfacial layer* has been identified in the lowest few centimeters of air, where molecular transport dominates over turbulent transport.

The following shorthand notation is often used for the various parts of the boundary layer. For the sake of completeness, some additional terms are listed here that will not be discussed until later:

BL   Boundary layer (also known as the planetary boundary layer, PBL, or the atmospheric boundary layer, ABL)
CL   Cloud layer
FA   Free atmosphere
IBL   Internal boundary layer
ML   Mixed layer (also known as the convective boundary layer, CBL)
RL   Residual layer
SBL   Stable boundary layer (also known as the nocturnal boundary layer, NBL)
SCL   Subcloud layer
SL   Surface layer (the bottom 10% of the boundary layer)

The tops of four of these layers are given the following symbols:
h Top of the stable boundary layer (often defined as the top of the NBL)
z_i Top of the mixed layer (often defined as the average base of the overlying stable layer)
z_r Top of the residual layer (often defined as the average base of the overlying stable layer)
z_b Top of the subcloud layer (this is the height of cloud base, usually near the *lifting condensation level*, LCL)
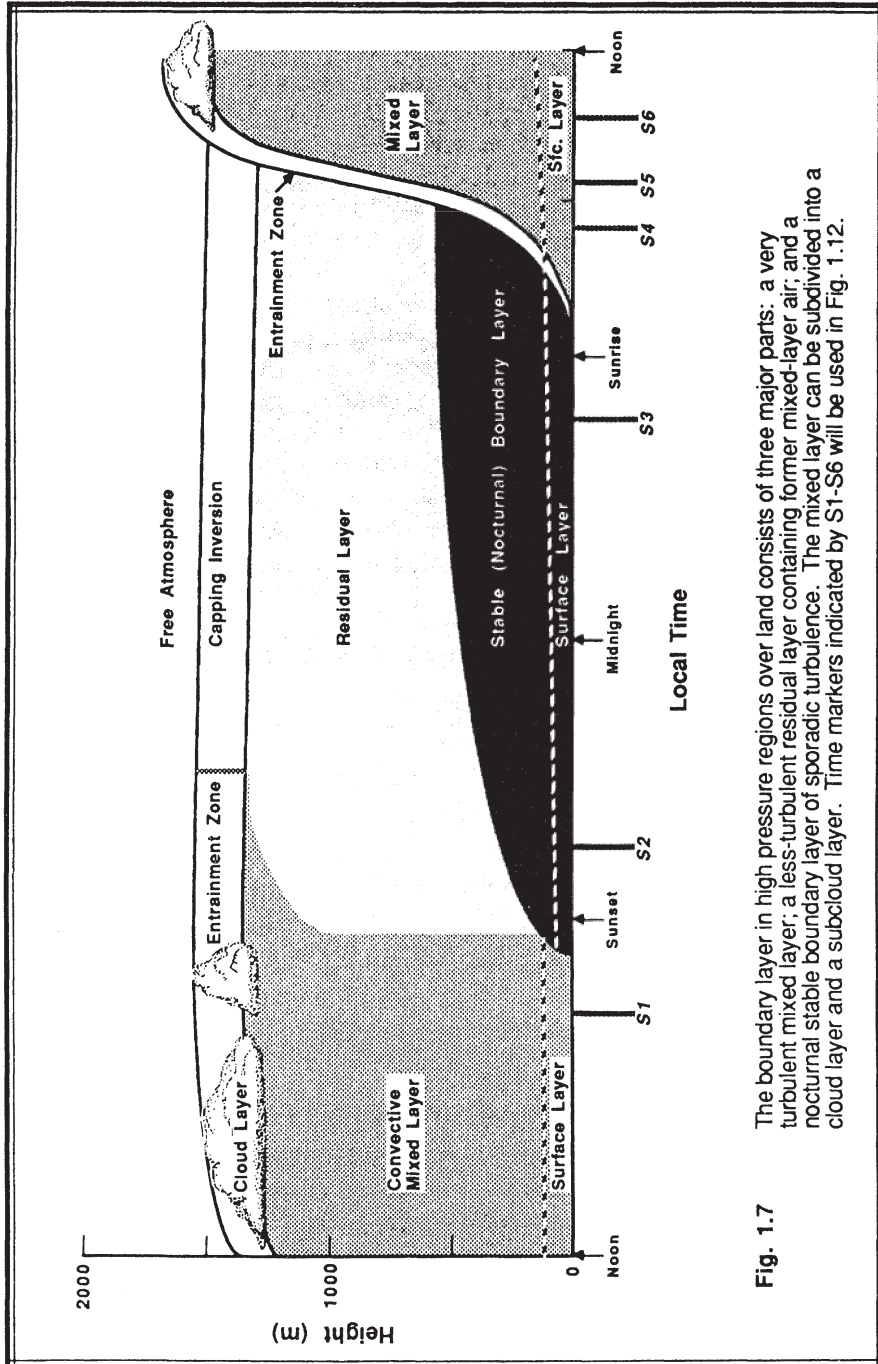
Fig. 1.7   The boundary layer in high pressure regions over land consists of three major parts: a very turbulent mixed layer; a less-turbulent residual layer containing former mixed-layer air; and a nocturnal stable boundary layer of sporadic turbulence. The mixed layer can be subdivided into a cloud layer and a subcloud layer. Time markers indicated by S1-S6 will be used in Fig. 1.12.

1.6.1 Mixed Layer

The turbulence in the mixed layer is usually *convectively driven*, although a nearly well-mixed layer can form in regions of strong winds. Convective sources include heat transfer from a warm ground surface, and radiative cooling from the top of the cloud layer. The first situation creates thermals of warm air rising from the ground, while the second creates thermals of cool air sinking from cloud top. Both can occur simultaneously, particularly when a cool stratocumulus topped mixed layer is being advected over warmer ground.

Even when convection is the dominant mechanism, there is usually *wind shear* across the top of the ML that contributes to the turbulence generation. This free-shear situation is more akin to CAT, and is thought to be associated with the formation and breakdown of waves in the air known as *Kelvin-Helmholtz* waves.

On initially cloud-free days, however, ML growth is tied to solar heating of the ground. Starting about a half hour after sunrise, a turbulent ML begins to grow in depth. This ML is characterized by intense mixing in a statically unstable situation where thermals of warm air rise from the ground (Fig 1.8). The ML reaches its maximum depth in late afternoon. It grows by *entraining*, or mixing down into it, the less turbulent air from above.
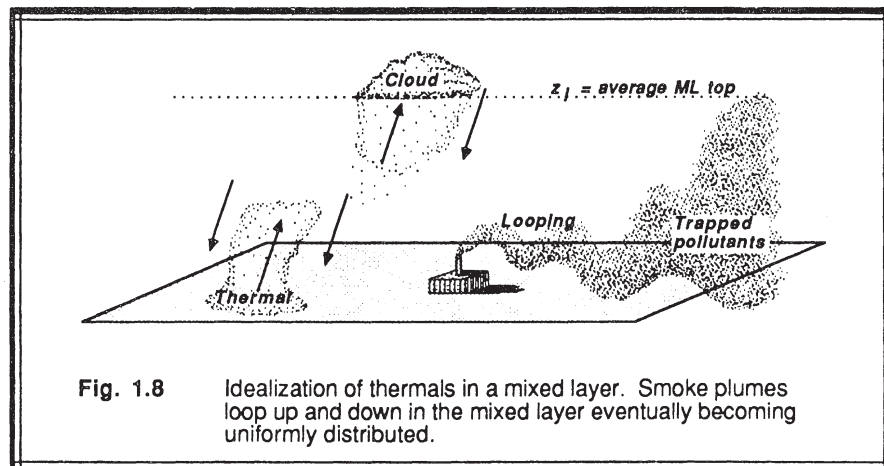
Fig. 1.8   Idealization of thermals in a mixed layer. Smoke plumes loop up and down in the mixed layer eventually becoming uniformly distributed.

The resulting turbulence tends to mix heat, moisture, and momentum uniformly in the vertical. Pollutants emitted from smoke stacks exhibit a characteristic *looping* as those portions of the effluent emitted into warm thermals begin to rise (Fig 1.8). The resulting profiles of virtual potential temperature, mixing ratio, pollutant concentration, and wind speed frequently are as sketched in Figure 1.9.

Virtual potential temperature profiles are nearly *adiabatic* in the middle portion of the ML. In the surface layer one often finds a *superadiabatic* layer adjacent to the ground.

Fig. 1.9   Typical daytime profiles of mean virtual potential temperature $\overline{\theta}_v$, wind speed $\overline{M}$ (where $\overline{M}^2 = \overline{U}^2 + \overline{V}^2$), water vapor mixing ratio $\overline{r}$, and pollutant concentration $\overline{c}$.
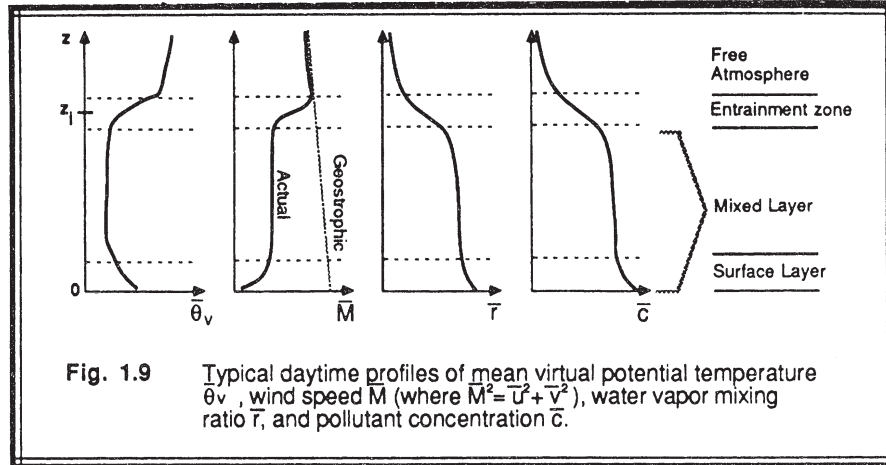
A stable layer at the top of the ML acts as a lid to the rising thermals, thus restraining the domain of turbulence. It is called the **entrainment zone** because entrainment into the ML occurs there.

At times this capping stable layer is strong enough to be classified as a temperature inversion; that is, the absolute temperature increases with height. In fact, it is frequently called an **inversion layer** regardless of the magnitude of the stability. The most common symbol for ML depth is z_i, which represents the average height of the inversion base.

Wind speeds are **subgeostrophic** throughout the ML, with wind directions **crossing the isobars** at a small angle towards low pressure. The middle portion of the ML frequently has nearly constant wind speed and direction. Wind speeds decrease towards zero near the ground, resulting in a wind speed profile that is nearly **logarithmic** with height in the surface layer. Wind directions cross the isobars at increasingly large angles as the ground is approached, with 45 degree angles not uncommon near the surface.

Mixing ratios tend to decrease with height, even within the center portion of the ML. This reflects the evaporation of soil and plant moisture from below, and the entrainment of drier air from above. The moisture decrease across the top of the ML is very pronounced, and is often used together with potential temperature profiles to identify the ML top from rawinsonde soundings.

Most pollutant sources are near the earth's surface. Thus, pollutant concentrations can build up in the ML while FA concentrations remain relatively low. Pollutants are transported by eddies such as thermals; therefore, the inability of thermals to penetrate very far into the stable layer means that the stable layer acts as a lid to the pollutants too. **Trapping** of pollutants below such an "inversion layer" is common in high- pressure regions, and sometimes leads to pollution alerts in large communities.

As the tops of the highest thermals reach greater and greater depths during the course of the day, the highest thermals might reach their *lifting condensation level*, LCL, if sufficient moisture is present. The resulting fair-weather clouds are often targets for soaring birds and glider pilots, who seek the updraft of the thermals.

High or middle overcast can reduce the insolation at ground level. This, in turn, reduces the intensity of thermals. On these days the ML may exhibit slower growth, and may even become nonturbulent or neutrally-stratified if the clouds are thick enough.

1.6.2 Residual Layer

About a half hour before sunset the thermals cease to form (in the absence of cold air advection), allowing turbulence to decay in the formerly well-mixed layer. The resulting layer of air is sometimes called the residual layer because its initial mean state variables and concentration variables are the same as those of the recently-decayed mixed layer.

For example, in the absence of advection, passive tracers dispersed into the daytime mixed layer will remain aloft in the RL during the night. The RL is neutrally stratified, resulting in turbulence that is nearly of equal intensity in all directions. As a result, smoke plumes emitted into the RL tend to disperse at equal rates in the vertical and lateral directions, creating a cone-shaped plume. Figure 1.10 shows a sketch of *coning*.
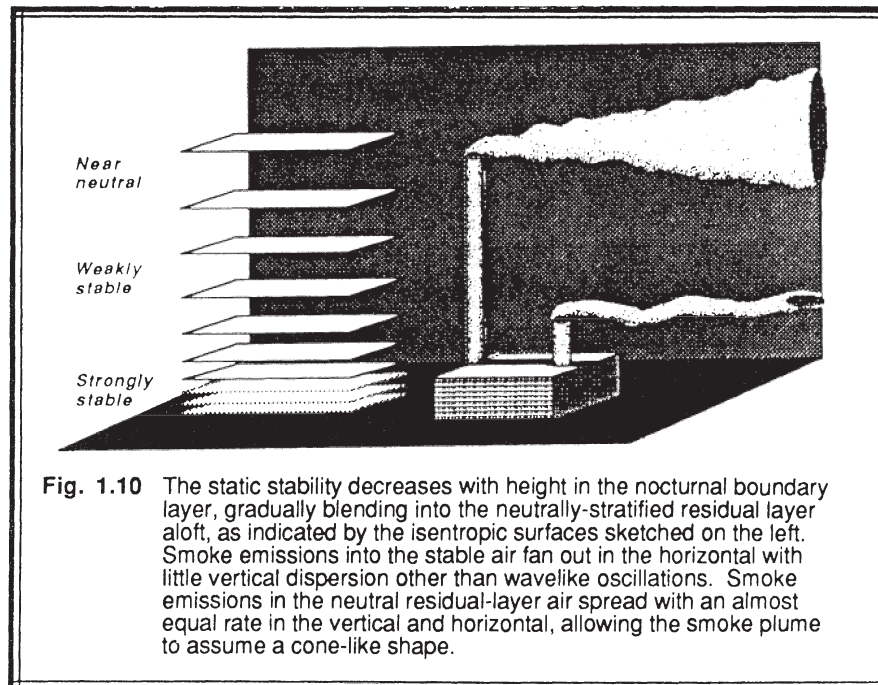
Fig. 1.10 The static stability decreases with height in the nocturnal boundary layer, gradually blending into the neutrally-stratified residual layer aloft, as indicated by the isentropic surfaces sketched on the left. Smoke emissions into the stable air fan out in the horizontal with little vertical dispersion other than wavelike oscillations. Smoke emissions in the neutral residual-layer air spread with an almost equal rate in the vertical and horizontal, allowing the smoke plume to assume a cone-like shape.

Nonpassive pollutants may react with other constituents during the night to create compounds that were not originally emitted from the ground. Sometimes gaseous chemicals may react to form aerosols or particulates which can precipitate out. The RL often exists for a while in the mornings before being entrained into the new ML. During this time solar radiation may trigger photochemical reactions among the constituents in the RL.

Moisture often behaves as a passive tracer. Each day, more moisture may be evaporated into the ML and will be retained in the RL. During succeeding days, the re-entrainment of the moist air into the ML might allow cloud formation to occur where it otherwise might not.

Variables such as virtual potential temperature usually decrease slowly during the night because of radiation divergence. This cooling rate is on the order of 1 °C/d. The cooling rate is more-or-less uniform throughout the depth of the RL, thus allowing the RL virtual potential temperature profile to remain nearly adiabatic. When the top of the next day's ML reaches the base of the RL, the ML growth becomes very rapid.

The RL does not have direct contact with the ground. During the night, the nocturnal stable layer gradually increases in thickness by modifying the bottom of the RL. Thus, the remainder of the RL is not affected by turbulent transport of surface-related properties and hence does not really fall within our definition of a boundary layer. Nevertheless, we will include the RL in our studies as an exception to the rule.

1.6.3 Stable Boundary Layer

As the night progresses, the bottom portion of the residual layer is transformed by its contact with the ground into a stable boundary layer. This is characterized by statically stable air with weaker, sporadic turbulence. Although the wind at ground level frequently becomes lighter or calm at night, the winds aloft may accelerate to *supergeostrophic speeds* in a phenomenon that is called the *low-level jet* or *nocturnal jet*.

The statically stable air tends to suppress turbulence, while the developing nocturnal jet enhances wind shears that tend to generate turbulence. As a result, turbulence sometimes occurs in relatively short bursts that can cause mixing throughout the SBL. During the nonturbulent periods, the flow becomes essentially decoupled from the surface.

As opposed to the daytime ML which has a clearly defined top, the SBL has a poorly-defined top that smoothly blends into the RL above (Fig 1.10 and 1.11). The top of the ML is defined as the base of the stable layer, while the SBL top is defined as the top of the stable layer or the height where turbulence intensity is a small fraction of its surface value.

Pollutants emitted into the stable layer disperse relatively little in the vertical. They disperse more rapidly, or "fan out", in the horizontal. This behavior is called *fanning*, and is sketched as the bottom smoke plume in Fig 1.10. Sometimes at night when winds are lighter, the effluent *meanders* left and right as it drifts downwind.

Winds exhibit a very complex behavior at night. Just above ground level the wind speed often becomes light or even calm. At altitudes on the order of 200 m above ground, the wind may reach 10-30 m/s in the nocturnal jet. Another few hundred meters above

Fig. 1.11   Mean virtual potential temperature, $\overline{\theta}_v$, and wind speed, $\overline{M}$, profiles for an idealized stable boundary layer in a high-pressure region.
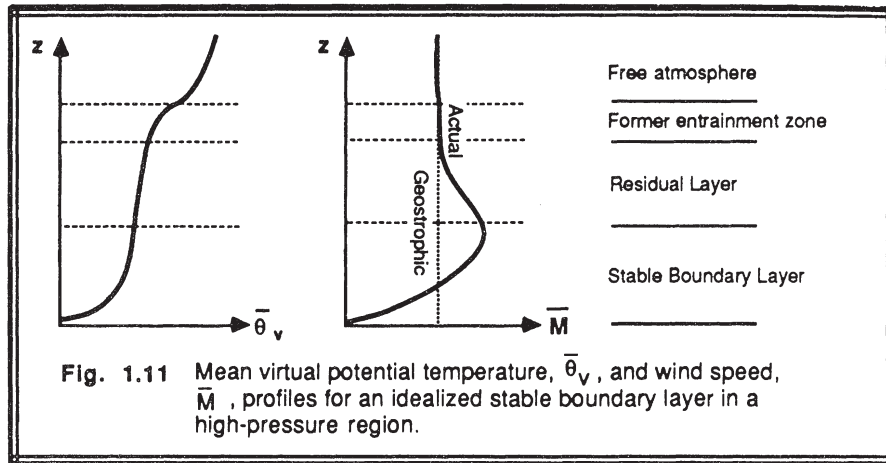
that, the wind speed is smaller and closer to its geostrophic value. The strong shears below the jet are accompanied by a rapid change in wind direction, where the lower level winds are directed across the isobars towards low pressure.

Touching the ground, however, is a thin (order of a few meters) layer of *katabatic* or *drainage* winds. These winds are caused by the colder air, adjacent to the ground, flowing downhill under the influence of gravity. Wind speeds of 1 m/s at a height of 1 m are possible. This cold air collects in the valleys and depressions and stagnates there. Unfortunately, many weather stations are located in or near valleys, where the observed surface winds bear little relationship to the synoptic-scale forcings at night.

Wave motions are a frequent occurrence in the SBL. The strongly stable NBL not only supports gravity waves, but it can trap many of the higher-frequency waves near the ground. Vertical wave displacements of 100 m have been observed, although the associated wind and temperature oscillations are relatively small and difficult to observe without sensitive instruments.

SBLs can also form during the day, as long as the underlying surface is colder than the air. These situations often occur during warm-air advection over a colder surface, such as after a warm frontal passage or near shorelines.

1.6.4 Virtual Potential Temperature Evolution

Given the virtual potential temperature profiles from the previous subsections, it is useful to integrate these profiles into our concept of how the boundary layer evolves. If rawinsonde soundings were made at the times indicated by flags S1 through S6 in Fig 1.7, then Fig 1.12 shows the resulting virtual potential temperature profile evolution.

We see from these soundings that knowledge of the virtual potential temperature profile is usually sufficient to identify the parts of the boundary layer. The structure of the BL is clearly evident.

Fig. 1.12
Profiles of mean virtual potential temperature, $\bar{\theta}_v$, showing the boundary-layer evolution during a diurnal cycle starting at about 1600 local time. S1-S6 identify each sounding with an associated launch time indicated in Fig. 1.7.
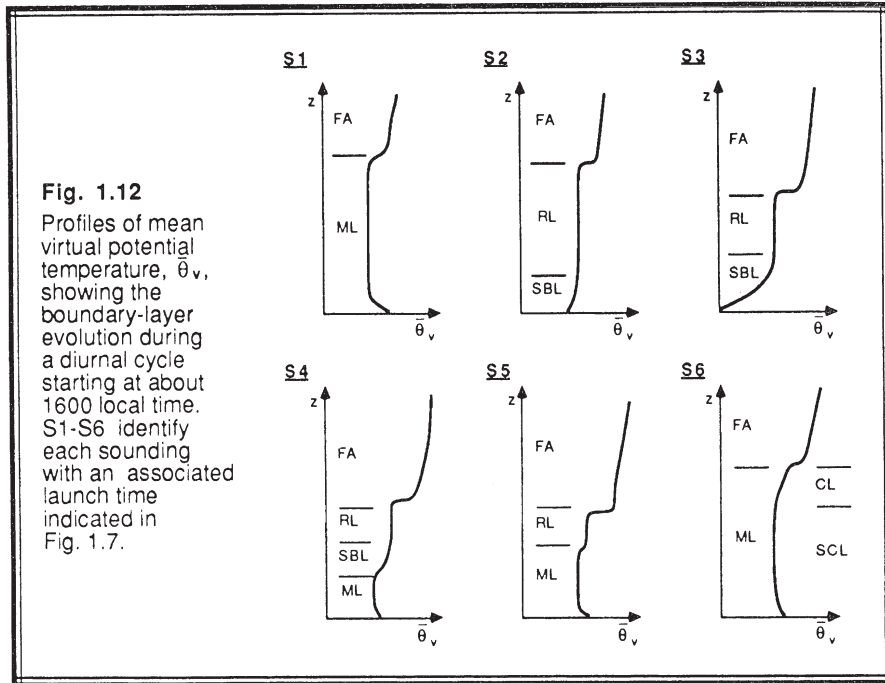
Stated another way, knowledge of the virtual potential temperature lapse rate is usually sufficient for determining the static stability. An exception to this rule is evident by comparing the lapse rate in the middle of the RL with that in the middle of the ML. Both are adiabatic; yet, the ML corresponds to statically unstable air while the RL contains statically neutral air.

One way around this apparent paradox for the classification of adiabatic layers is to note the lapse rate of the air immediately below the adiabatic layer. If the lower air is superadiabatic, then both that superadiabatic layer and the overlying adiabatic layer are statically unstable. Otherwise, the adiabatic layer is statically neutral. A more precise definition of static stability is presented later.

It is obvious that as the virtual potential temperature profile evolves with time, so must the behavior of smoke plumes. For example, smoke emitted into the top of the NBL or into the RL rarely is dispersed down to the ground during the night because of the limited turbulence. These smoke plumes can be advected hundreds of kilometers downwind from their sources during the night.

Smoke plumes in the RL may disperse to the point where the bottom of the plume hits the top of the NBL. The strong static stability and frequent reduction in turbulence reduces the downward mixing into the NBL. The top of the smoke plume sometimes can continue to rise into the neutral air. This is called *lofting* (see Fig 1.13).

Fig. 1.13   Lofting of a smoke plume occurs when the top of the plume grows upward into a neutral layer of air while the bottom is stopped by a stable layer.
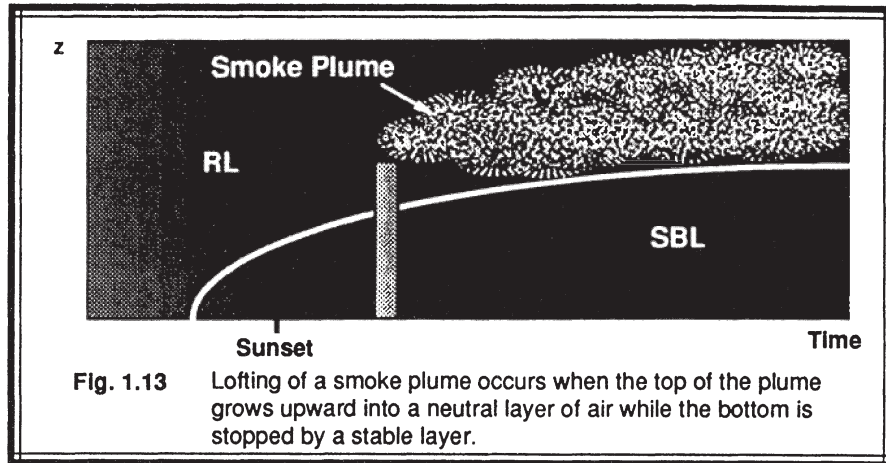
After sunrise a new ML begins to grow, eventually reaching the height of the elevated smoke plume from the previous night. At this time, the elevated pollutants are mixed down to the ground by ML entrainment and turbulence in a process that is called *fumigation*. A sketch of this process is shown in Fig 1.14. An analogous process is

Fig. 1.14   Sketch of the fumigation process, where a growing mixed layer mixes elevated smoke plumes down to the ground. Smoke plume 1 is fumigated at time F1, while plume 2 is fumigated at time F2.
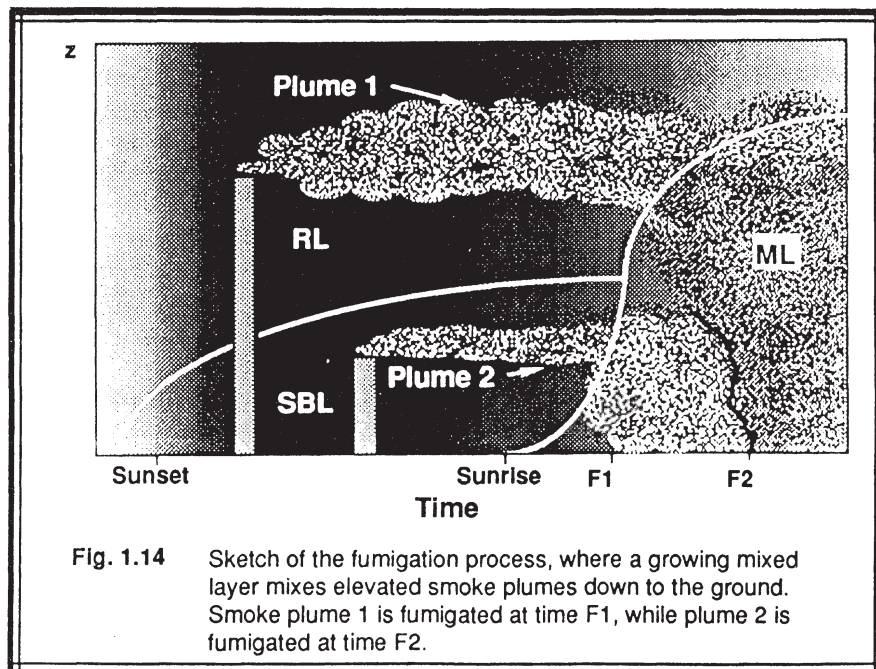
often observed near shorelines, where elevated smoke plumes in stable or neutral air upstream of the shoreline are continuously fumigated downstream of the shoreline after advecting over a warmer bottom boundary that supports ML growth.

1.7 Micrometeorology

Compared to the other scales of meteorological motions, turbulence is on the small end. Figure 1.15 shows a classification scheme for meteorological phenomena as a function of their time and space scales. Phenomena such as turbulence with space scales smaller than about 3 km and with time scales shorter than about 1 h are classified as *microscale*. *Micrometeorology* is the study of such small-scale phenomena.

It is evident that the study of the boundary layer involves the study of microscale processes. For this reason, boundary layer meteorology and micrometeorology are virtually synonymous. Since many of the early micrometeorological measurements were made with sensors on short stands and towers, micrometeorology was often associated with surface-layer phenomena.

Regardless of what you call it, the small-scale phenomena being studied here are so transient in nature that the deterministic description and forecasting of each individual eddy is virtually impossible. As a result, micrometeorologists have developed three primary avenues for exploring their subject:
- stochastic methods
- similarity theory
- phenomenological classifications.

Stochastic methods deal with the average statistical effects of the eddies. *Similarity theory* involves the apparent common-behavior exhibited by many empirically-observed phenomena, when properly scaled. In the *phenomenological* methods, the largest size structures such as thermals are classified and sometimes approached in a partially deterministic manner.

Micrometeorology has always relied heavily on *field experiments* to learn more about the boundary layer. Unfortunately, the large variety of scales involved and the tremendous variability in the vertical require a large array of sensors including airborne platforms and remote sensors. The relatively large costs have limited the scope of many field experiments. Only a few general-purpose, large-scale boundary layer experiments have been conducted.

Alternative studies have used *numerical* and *laboratory simulations*. Much of the turbulence work has been performed in laboratory tanks, usually using liquids such as water as the working medium. Although there have been many successful laboratory studies of small-scale turbulence, there have been only a few simulations of larger phenomena such as thermals. Wind tunnel studies have been used to observe the flow of neutral boundary layers over complex terrain and buildings, although the difficulty of stratifying the air has meant that typical daytime and nighttime boundary layers could not be adequately simulated.
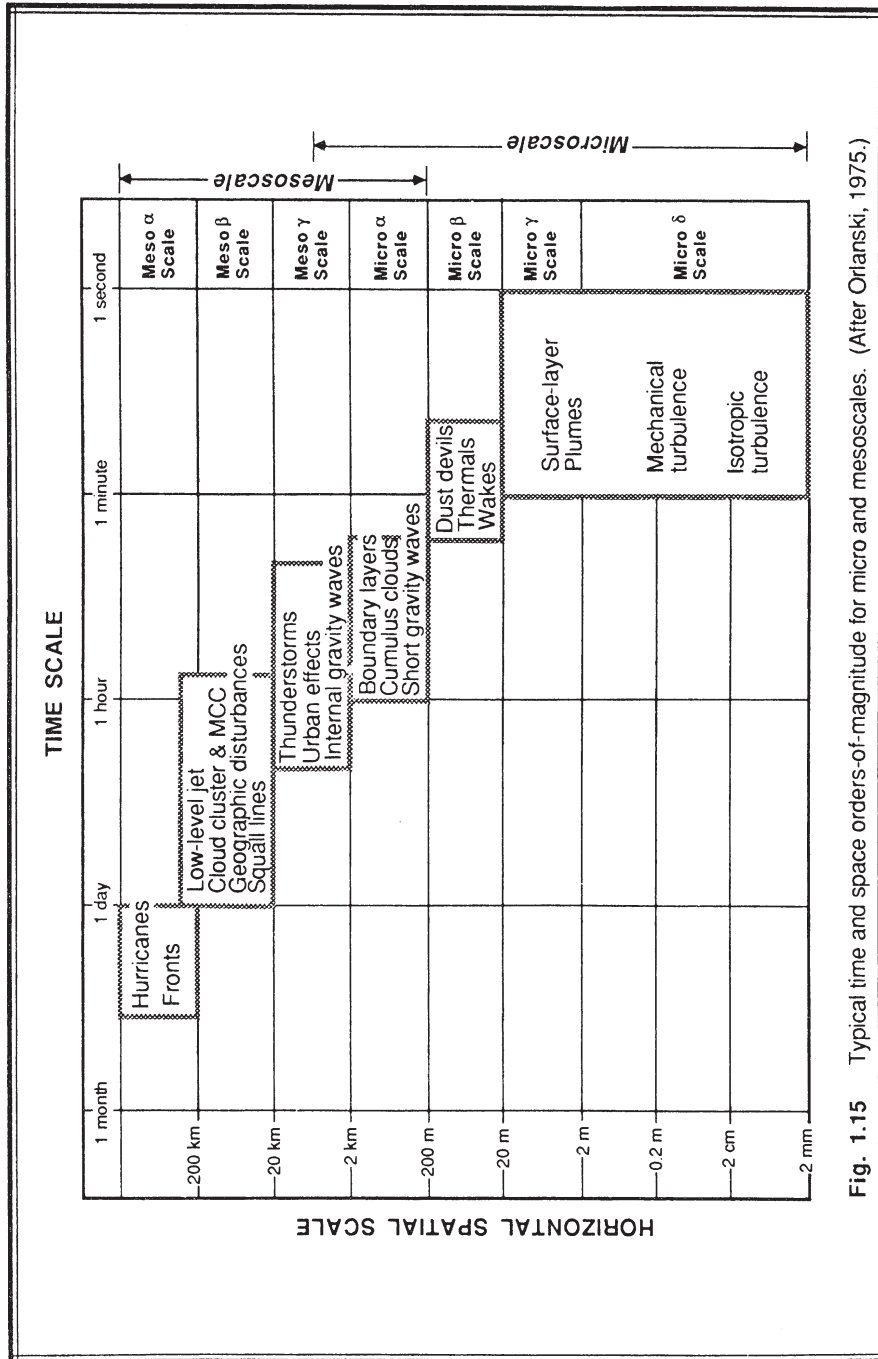
Fig. 1.15   Typical time and space orders-of-magnitude for micro and mesoscales. (After Orlanski, 1975.)

Numerical simulation using digital computers has been very popular since the 1960's, with many discoveries that were verified years later with field studies. Most of these computer simulations employ the stochastic method of modeling fluid flow. Unfortunately, a difficulty known as the *closure problem* has meant that each of these models has at best been able to only approximate the governing equations, with uncertainties introduced via a necessary parameterization of the unknowns.

That same closure problem has limited the avenues for *theoretical studies* involving analytical solutions. For the most part, only highly simplified approximations to the boundary layer have been amenable to direct solution.

If anything, these difficulties have stimulated, rather than stifled, the work of micrometeorologists. There is an underlying assumption in meteorology that subsynoptic-scale phenomena such as turbulence might be responsible, in part, for the difficulty in making quality weather forecasts beyond a few days. Thus, part of the effort in boundary layer meteorology involves the search for adequate turbulence parameterization schemes for larger-scale numerical forecast models.

Additional motivation has come from concern over our environment. Every species of animal and plant modifies its environment; the human species, however, is in a position to recognize the consequences of its pollution and take appropriate action. Since most of the anthropogenic effluents are emitted from near-surface sources, the resulting dispersion of the pollutants is tied to boundary layer processes. As a result, *air-pollution meteorology* is an applied form of micrometeorology.

Other applications include *agricultural meteorology*, where airborne transport of chemicals necessary to plant life is governed by turbulence. Nocturnal processes such as frost formation warrant improved study and forecast methods for crop protection. Fog and low stratocumulus, which inhibit aviation operations, are essentially boundary layer phenomena. Wind-generated power, a popular energy source for centuries, has had a recent increase in interest as wind turbines have been designed to extract energy more efficiently from the boundary layer wind. Other structures such as bridges and buildings must be designed to withstand wind gusts appropriate to their sites.

1.8 Significance of the Boundary Layer

The role of the boundary layer on our lives is put into perspective when we compare the characteristics of the boundary layer and free atmosphere (Table 1-1). A taste of the importance of the BL is given in the following summary:
- People spend most of their lives in the BL.
- Daily weather forecasts of dew, frost, and maximum and minimum temperatures are really BL forecasts.
- Pollution is trapped in the BL.
- Fog occurs within the BL.
- Some aviation, shipping, and other commerce activities conducted within it.
- Air masses are really boundary layers in different parts of the globe that have equilibrated with their underlying surface. Baroclinicity is generated this way.

Table 1-1. Comparison of boundary layer and free atmosphere characteristics.

Property	Boundary Layer	Free Atmosphere
Turbulence	• Almost continuously turbulent over its whole depth.	• Turbulence in convective clouds, and sporadic CAT 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 mean wind.
Winds	• Near logarithmic wind speed profile in the surface layer. Subgeostrophic, cross-isobaric flow 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. 8-18 km. Slow time variations.

- The primary energy source for the whole atmosphere is solar radiation, which for the most part is absorbed at the ground and transmitted to the rest of the atmosphere by BL processes. About 90% of the net radiation absorbed by oceans causes evaporation, amounting to the evaporation of about 1m of water per year over all the earth's ocean area. The latent heat stored in water vapor accounts for 80% of the fuel that drives atmospheric motions.
- Crops are grown in the BL. Pollen distributed by boundary layer circulations.
- Cloud nuclei are stirred into the air from the surface by BL processes.
- Virtually all water vapor that reaches the FA is first transported through the BL by turbulent and advective processes.
- Thunderstorm and hurricane evolution are tied to the inflow of moist BL air.
- Turbulent transport of momentum down through the BL to the surface is the most important momentum sink for the atmosphere.
- About 50% of the atmosphere's kinetic energy is dissipated in the BL.
- Turbulence and gustiness affects architecture in the design of structures.
- Warm and cold fronts separate boundary layers of different temperature.
- Wind turbines extract energy from the BL winds.

• Wind stress on the sea surface is the primary energy source for ocean currents.
• Turbulent transport and advection in the BL move water and oxygen to and from immobile life forms like plants.

Obviously the list could go on, but the main point is BL processes affect our lives directly, and indirectly via its influence on the rest of the weather. In this book we examine some of the processes that occur in the BL, develop some schemes for coping with turbulence, and show how they can be applied to benefit mankind.

For additional information on boundary layers, turbulence and micrometeorology, the General References section below lists books and other secondary sources. A reference section at the end of each chapter lists the specific articles sited in the chapter.

A summary table of frequently-used scaling variables and dimensionless groups is given in Appendix A. Appendix B lists notation, including abbreviations, symbols and acronyms. Appendix C gives values of frequently-used parameters and constants. Additional appendices are specifically referenced in the text.

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1.11 Exercises

1) a) Resketch figure 1.7 without any shading.
b) Next, shade on your figure those regions where the lapse rate is superadiabatic. (A color like red is appropriate.)
c) In a different color (like dark blue), shade those regions that have a strongly stable lapse rate.
d) With a different color (like light blue), shade those regions that have a weakly stable lapse rate.
e) Leave the nearly adiabatic lapse rate regions unshaded. (Hint, look at the virtual potential temperature graphs shown after figure 1.7.)

2) From your answer above, we see that a large portion of the mixed layer and the residual layer have nearly adiabatic lapse rates. Nevertheless, there is a difference between the two. Comment on this difference. How would smoke disperse in these regions?

3) If a very tall smoke stack could inject a smoke plume into the free atmosphere, then what characteristic dispersion pattern would you expect? Options include looping, fanning, coning, lofting, fumigation, and trapping within a layer.

4) Given the boundary layer evolution shown in Fig 1.7 and 1.12, sketch the modified evolution that would occur if the sun was turned off at time S5. Also comment on the processes that could effectively turn off the effects of the sun on the boundary layer.

5) If you observed the following measurement of wind speed on a strip chart, then what diameter (in meters) are the major eddies? (Hint: First find the typical time duration of an eddy, and then employ Taylor's Hypothesis.)
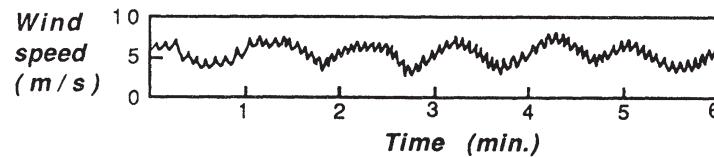
6) Suppose that the boundary layer can be approximated by a 1 km constant thickness layer over the whole continental United States. Above that, assume that a constant thickness (10 km) free atmosphere exists. Assume that each of these layers is well mixed for simplicity. (You may need to consult synoptics textbooks or study a number of weather maps to answer the following questions.)
a) On the average in spring, what fraction of the area of the continental United States is covered by cyclones (low pressure regions)?
b) What is the average vertical velocity magnitude out of the top of the boundary layer in those cyclonic regions?
c) What is the average subsidence velocity acting on the top of the boundary layer in anticyclonic (high pressure) regions? Remember that the boundary layer is of constant

thickness, so the total inflow of air must equal the outflow. (Normally, subsidence would not be able to cause air to penetrate down into the top of the boundary layer, but for simplicity we will assume that the turbulent entrainment rate of air into the top of the boundary layer is exactly equal to the subsidence velocity. Thus, all of the subsiding air would enter the boundary layer.)

d) What are the average residence times, P , of air within the boundary layer and the free atmosphere? Residence time is defined as P=V/F, where F = volumetric flow rate (volume/time) and V = volume. Neglect horizontal advection out of the United States.

e) Compare these residence times with the typical Rossby wave time period (that is, the average time period between cyclone passage over a fixed point). Comment on the significance of these times.

7) Would there be a boundary layer on a planet that had an atmosphere, but that did not experience a diurnal variation of net radiation at the ground?

8) Define the boundary layer in your own terms.

9) Add five items to the list in section 1.8 regarding the significance of the boundary layer. Hint, look for things close to home, school, or work.

10) a) Given air at a pressure height of 90 kPa (900 mb) with a temperature of 30 °C and a mixing ratio of 20 g/kg, find the virtual potential temperature.
b) Given saturated air at 85 kPa (850 mb) with a temperature of 20 °C and a total water mixing ratio (i.e., sum of vapor and liquid mixing ratios) of 20 g/kg find the virtual potential temperature. (Hint, you might want to employ a thermodynamic diagram.)

11) Given the following virtual potential temperature sounding, identify each layer (example, ML, RL, SBL, FA). Estimate what time of day that sounding was made, and then sketch the virtual potential temperature profile that you might expect four hours later.