# Soil 7170 Agricultural Micrometeorology
# Background Information

# Section 1
# The Earth's Atmosphere

## 2.1 The Temperature-Defined Zones of the Atmosphere

The word "atmosphere" is derived from the Greek words "atmos" (meaning vapor) and "sphaira" (meaning sphere). It is now used to denote the gaseous sphere that surrounds a planet and includes not only water vapor but also numerous gases and aerosols. The atmosphere is a mechanical mixture of gases. It is a dynamic fluid that obeys the laws of fluid mechanics or Newtonian physics.

Four major atmospheric zones are defined based on the thermal properties of the atmosphere (Figure 2.1). The zones are not necessarily sharply separated and do not occur at fixed distances. Therefore, the distances given in Figure 2.1 are an approximation. For example, the troposphere varies in depth both spatially and temporally. It is thickest at the equator (18 km) and thinnest at the poles (8 km) and thicker during the summer months.

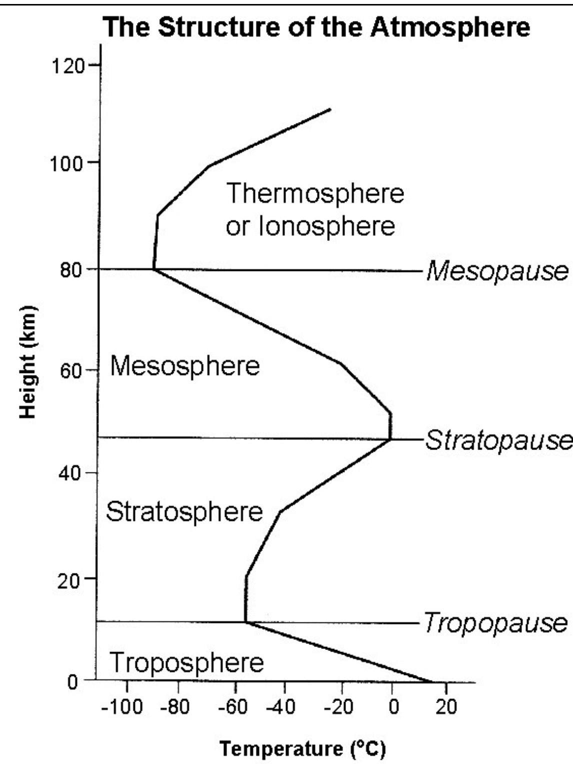


Figure 2.1 The temperature-defined zones of the atmosphere (Thompson, 1998).

The outer limit of the atmosphere is difficult to define. It can be calculated as about 32,000 km, or the distance at which the Earth's gravitational pull approximates the centrifugal force of the Earth's rotation. Even though the atmosphere reaches these altitudes, about 99% of the atmosphere is within just 32 km of the Earth's surface and 90% of the water vapor and virtually all the weather systems occur in the lowest 16 km.

The **troposphere** is the zone in which virtually all the major weather phenomena occur. The air in the troposphere generally cools progressively with altitude. The main source of heat for the troposphere is the longwave radiation emitted from the Earth's surface and the air that is further away from this heat source tends to be cooler. The **tropopause** is the zone where the temperature no longer continues to cool with altitude. The temperature decrease stops and isothermal conditions exist into the lower stratosphere.

The middle and upper **stratosphere** becomes increasingly warmer with elevation. This inversion of temperature is associated with the zone of maximum ozone concentration. Stratospheric ozone is formed by the reaction of diatomic oxygen ($O_2$) with ultraviolet (UV) solar radiation. The solar energy breaks the diatomic oxygen bond and some free oxygen atoms (O) bond with diatomic oxygen to form triatomic oxygen ($O_3$) or ozone. This formation is catalyzed by a neutral molecule (M), usually nitrogen. Since M also takes up the kinetic energy in this reaction, it accelerates and becomes hotter, which warms up the stratosphere.

(2.1) $$O_2 + UV \longrightarrow O + O$$

(2.2) $$O + O_2 + M \longrightarrow O_3 + M^*$$

where M* is a more energetic version of M that carries away the heat of ozone production.

Ozone is concentrated in the stratosphere, especially at elevations between 16 and 25 km. The reason for the abundance of ozone in the stratosphere is related to the decrease in the availability of molecular oxygen with altitude. Gravity keeps most oxygen molecules ($O_2$) near Earth's surface. However, the availability of atomic oxygen (O) increases with altitude. The creation of ozone requires both oxygen molecules and oxygen atoms. In general, the collision of three particles ($O_2$, O, M) is a rare event at great altitudes where the density of air is very low but becomes more common with decreasing altitude (increasing atmospheric density). The elevation at which ozone can be created most efficiently is therefore a compromise between low elevations where the density of air is high enough that three particles collide frequently, and great heights where atomic oxygen is abundant. This compromise is reached in the stratosphere, which therefore is the location of the ozone layer.

The **stratopause** marks the point at the top of the stratosphere where the temperature no longer warms with elevation. Above the stratopause, the **mesosphere** is again a zone of decreasing temperature with elevation. At a height of approximately 80 km, the **mesopause** marks the point where conditions again become isothermal.

The uppermost zone of the atmosphere, termed the **thermosphere** or **ionosphere** is another layer of increasing temperature with elevation. At elevations of 100 km above the North Pole, rockets have recorded temperatures as high as 1480 $^oC$. The density of the atmosphere at this altitude is extremely low. The lower portion of the thermosphere is composed mainly of nitrogen ($N_2$), and oxygen in molecular ($O_2$) and atomic (O) forms. Above 200 km altitude, atomic oxygen predominates. Temperatures rise with elevation in the lower thermosphere through the absorption of extreme ultraviolet radiation by molecular and atomic oxygen. Gases become ionized or dissociated into electrically-charged ions. These ionized particles reflect electromagnetic waves and act as a "mirror" for radio waves. At greater heights, the temperature of the thermosphere declines towards 3$^o$K, which is the background temperature of deep space.

The most important thing to remember about the typical temperature profile of the earth's atmosphere is that it occurs as a result of three distinct and very important sources of heat energy. These sources are longwave terrestrial radiation from the earth's surface (discussed more thoroughly in Section 3), the heat produced by ozone production, which reaches a maximum near the stratopause and the heat produced by ionization of oxygen in the thermosphere.

The 1976 U.S. Standard Atmosphere is an idealized, dry, steady-state approximation of the atmosphere with altitude. It has been adopted as an engineering reference (Stull, 2000). It uses geopotential height, rather than altitude, to compensate for the decrease of gravitational acceleration above the earth's surface. Below a geopotential height of 51 km (up to the top of the stratopause), equations 2.3 through 2.7 can be used to approximate the standard atmospheric temperature as a function of altitude.

(2.3) $T = 288.15\ K - (6.5\ K / km) \times H$ for $H \leq 11$ km

(2.4) $T = 216.65\ K$ $11 \leq H \leq 20$ km

(2.5) $T = 216.65\ K + (1\ K / km) \times (H - 20\ km)$ $20 \leq H \leq 32$ km

(2.6) $T = 228.65\ K + (2.8\ K / km) \times (H - 32\ km)$ $32 \leq H \leq 47$ km

(2.7) $T = 270.65\ K$ $47 \leq H \leq 51$ km

where $T$ = absolute temperature of the atmosphere in degrees Kelvin (K)
$H$ = geopotential height $= r_e \times z / (r_e + z)$
$r_e$ = average radius of the earth (6356.766 km)
$z$ = altitude

## 2.2 Atmospheric Pressure

### 2.2.1 Pressure Fundamentals

Pressure is force per unit area. Force is a mass multiplied by its acceleration. The SI unit of force is the Newton (1 Newton = 1 kg of mass x 1 m $s^{-2}$ of acceleration) and a square meter is the unit of area. Thus, pressure has units of N $m^{-2}$. One N $m^{-2}$ equals a Pascal (Pa). Static pressure (i.e. pressure with calm wind) is caused by randomly moving molecules of gas in the atmosphere that bounce off each other and off the surfaces they hit. In a vacuum, atmospheric pressure is zero.

Atmospheric pressure that you measure at any altitude is caused by the weight of all the air molecules above you. As you travel higher in the atmosphere there are fewer molecules still above you; hence, atmospheric pressure decreases with altitude (Fig 2.2).

Mean sea level pressure can be estimated by using the total mass of the atmosphere, the acceleration due to gravity along with the mean radius of the Earth to estimate the total surface area over which the pressure of the atmosphere is applied.

(2.8) $P_o = g_o (M_a / 4 \pi r_e^2)$

where $P_o$ = mean sea level pressure
$g_o$ = acceleration due to gravity = 9.8 m $s^{-2}$
$M_a$ = total mass of the atmosphere = 5.14 x $10^{18}$ kg
$r_e$ = average radius of the earth (6356.766 km)

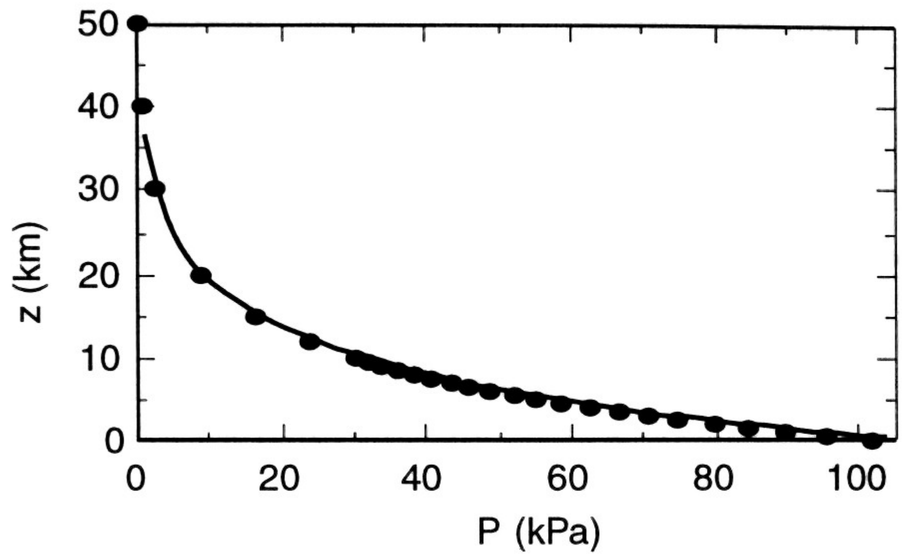


Figure 2.2 Altitude (z) versus pressure (P) in the atmosphere (Stull, 2000).

Equation 2.8 yields a value for $P_o$ of slightly less than 100,000 Pa. Measurements of global mean sea level pressure average 101,325 Pa, which is very similar to the calculated value. Mean sea level pressure is equivalent to the pressure exerted by a column of water 11 m high or a column of mercury 0.76 m high. There are a number of different pressure units utilized in various texts. One standard atmosphere (atm) equals:

i) 101,325 pascals (Pa)
ii) 1013.25 hectopascals (hPa)
iii) 101.325 kilopascals (kPa)
iv) 1.01325 bar
v) 1013.25 millibars (mb)
vi) 14.69595 pounds per square inch (psi)
vii) 760 mm Hg
viii) 760 Torr

Two simple laws specify the main factors governing changes in atmospheric pressure. Boyle's Law states that, at a constant temperature, the volume (V) of a gas varies inversely with pressure (P).

(2.9) $V = c_1 / P$

Charles's Law states that, at a constant pressure, the volume of a gas varies directly with absolute temperature (T), which is degrees Kelvin (K).

(2.10) $V = c_2\,T$

These laws imply that the three atmospheric qualities of pressure, temperature and volume are completely interdependent for any gas. A change in one parameter will cause a compensating change to occur in one or both of the remaining two. The gas laws can be combined to form the ideal gas law.

(2.11) $PV = Rm_aT$

The value, $m_a$, is the mass of air and R is a gas constant with a value of 287 J $kg^{-1}$ $K^{-1}$ for dry air. If $m_a$ and T are held fixed, then the right side of equation 2.11 is constant and we obtain Boyle's Law (equation 2.9). If $m_a$ and P are held fixed, we obtain Charles's Law (equation 2.10).

Below a geopotential height of 51 km (up to the top of the stratopause), equations 2.12 through 2.16 can be used to approximate atmospheric pressure for a standard atmosphere. These equations take into account the fact that temperature is not uniform with height.

(2.12) $P = (101.325\text{ kPa}) \times (288.15\text{ K} / T)^{-5.255877}$ for $H \leq 11$ km

(2.13) $P = (22.632\text{ kPa}) \times \exp[-0.1577 \times (H - 11\text{ km})]$ $11 \leq H \leq 20$

(2.14) $P = (5.4749\text{ kPa}) \times (216.65\text{ K} / T)^{34.16319}$ $20 \leq H \leq 32$

(2.15) $P = (0.868\text{ kPa}) \times (228.65\text{ K} / T)^{12.201}$ $32 \leq H \leq 47$

(2.16) $P = (0.1109\text{ kPa}) \times \exp[-0.1262 \times (H-47\text{ km})]$ $47 \leq H \leq 51$

where P = atmospheric pressure (kPa)
T = absolute temperature of the atmosphere in degrees Kelvin (K)
H = geopotential height $= r_e \times z / (r_e + z)$
$r_e$ = average radius of the earth (6356.766 km)
z = altitude

Note: Geopotential height, rather than altitude, is used to compensate for the decrease in gravitational acceleration with increasing altitude above the earth's surface.

The relationship between pressure and altitude is so significant that meteorologists usually express elevations in pressure units such as mb or kPa. When upper atmosphere readings were taken with the original radiosondes, the instruments measured pressure, not altitude, so it became a convention to plot atmospheric conditions as a function of pressure instead of altitude. A reference to 1000 mb elevation refers to sea level. A 500 mb elevation refers to an altitude about 5.5 km above sea level and 300 mb refers to an altitude about 9 km above sea level.

### 2.2.2 Measurement of Atmospheric Pressure

Atmospheric pressure is equal to the weight of a vertical column of air of unit area above a site extending to the outer limit of the atmosphere. The sea level pressure is the weight of a vertical column of air of unit area above sea level extending to the outer limit of the atmosphere. Weather stations situated at sea level will measure sea level pressure directly while other stations must calculate it by adding to the local pressure the equivalent weight of an air column extending from the station elevation down to sea level.

Manned climatological stations use mercury or aneroid barometers to measure atmospheric pressure. Both mercury and aneroid barometers must be read and recorded manually at scheduled reporting times. The basic principle of the mercury barometer is that the pressure of the atmosphere is balanced against the weight of a column of mercury the length of which is measured on a scale graduated in units of pressure (Figure 2.3). The mercury barometer is the most accurate of all barometers, however, it requires an altitude correction and a temperature correction because of thermal expansion of the mercury. In addition, it is highly inconvenient to move this fragile instrument with a tube about a meter long and a few kilograms of mercury.

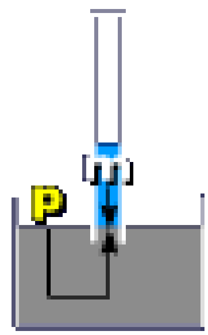


Figure 2.3 Schematic of a mercury barometer.

Aneroid barometers consist mainly of a closed metal chamber, completely or partly evacuated, and a strong spring system, which prevents the chamber from collapsing due to the external atmospheric pressure (Figure 2.4). An aneroid barometer is a mechanical barometer, so it has the inaccuracies and imprecisions inherent in any mechanical design. At any given atmospheric pressure there will be an equilibrium between the force of the spring and the external pressure. This equilibrium position is coupled to an indicator against a scale calibrated in units of pressure. If this expansion/contraction moves a pen across a chart, a recording barometer, known as a barograph, is created. The main advantages of an aneroid barometer are that it is somewhat more rugged than the mercury barometer and altitude and temperature corrections are not required.

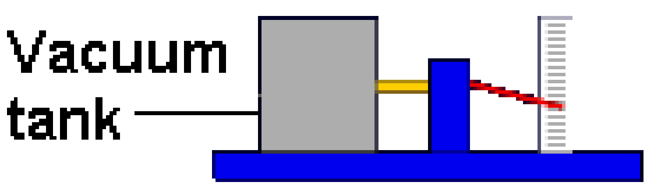


Figure 2.4 Schematic of an aneroid barometer

Climatological autostations use sensors to measure atmospheric pressure based on aneroid pressure transducers, which provide a proportional electrical signal. Electronic barometers have been around for some time but are now more prevalent due to the decreasing cost of materials. A piezoelectric crystal produces different amounts of current for different pressures on its faces. The data logger can measure and convert these electrical outputs into pressure units.

## 2.3 Atmospheric Density

The air offers so little resistance to our everyday activities that it is often difficult to appreciate that it has a density that averages 1.225 kg $m^{-3}$ at sea level. Density increases as the number and molecular weight of molecules in a volume increase. Gases are compressible. Therefore, as the pressure on a gas increases, the volume of the gas decreases (equation 2.9). Atmospheric density can vary over a wide range.

Density decreases roughly exponentially with height in an atmosphere of uniform temperature (Figure 2.5). The lower layers of the atmosphere are much denser than those above because the weight of the overlying atmosphere compresses the lowest layers. Roughly 80% of the atmosphere's mass is contained within the troposphere (Figure 2.6).

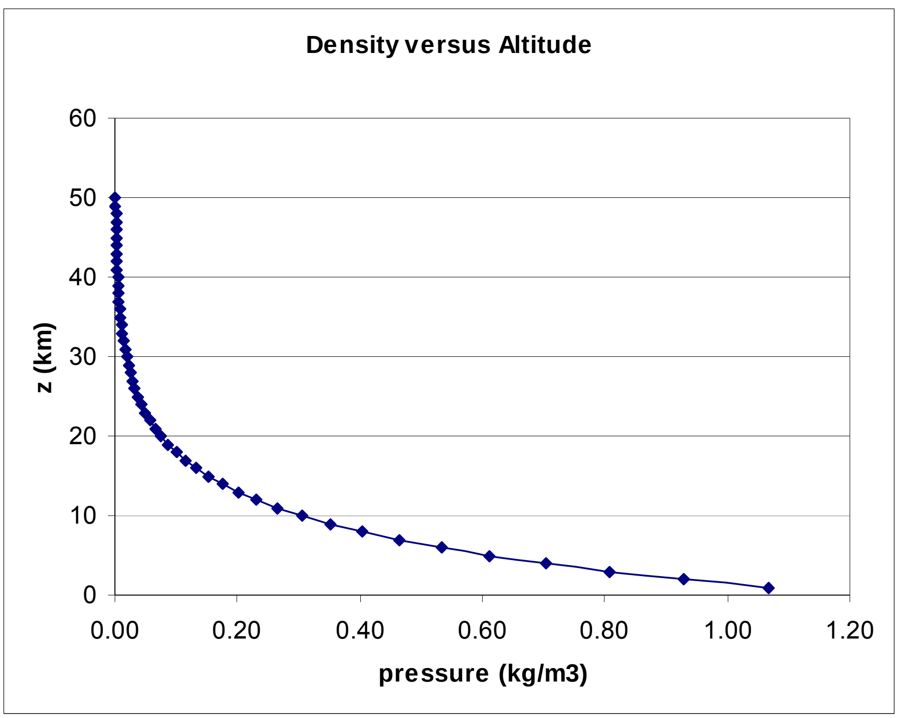


Figure 2.5 Atmospheric density versus altitude assuming air temperature is a constant 288 K with altitude (from Stull, 2000).

## 2.4 The Atmosphere in Motion

### 2.4.1 The Planetary Boundary Layer

The vast majority of land-dwelling organisms on the planet spend their entire existence surrounded by the lowest layer of the atmosphere. The conditions in this layer are what affect our everyday lives. Our basic existence requires that photosynthesizing plants can flourish in abundance in their environment, which is largely a function of the atmospheric conditions that exist in their location.

The planetary boundary layer, a depiction of which is shown in Figure 2.8, is 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 (Stull, 1988). The presence of the solid earth, with its significant fluctuation in surface temperature from night to day, can be observed in the atmosphere immediately adjacent to the earth. At progressively higher altitudes the diurnal fluctuations diminish and eventually disappear (Figure 2.9).

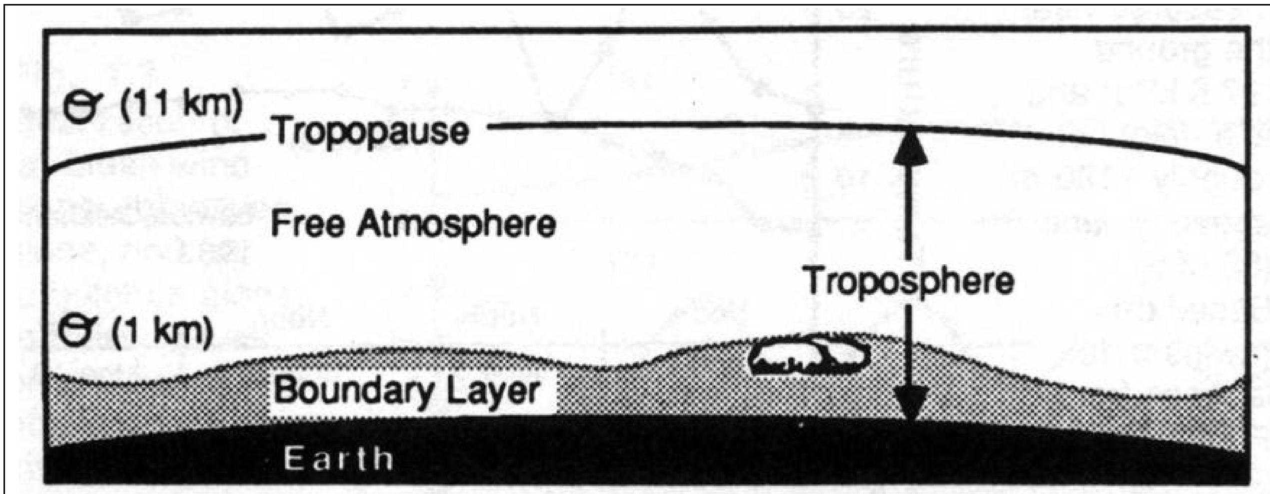


Figure 2.8 A depiction of the planetary boundary layer within the troposphere (Stull, 1988).

The planetary boundary layer is extremely dynamic. Its thickness fluctuates over the course of a day according to the strength of mixing that is occurring. By midday, the layer can be 2000 m thick or more as a result of vigorous mixing caused by rising pockets of warm air or thermals. Rising warm air is described as "buoyant". Buoyant air creates turbulence which mixes the air and everything it contains.

At night, the earth's surface cools more rapidly than the atmosphere and there is no longer an upward transfer of heat. The air in the atmosphere right next to the earth's surface becomes cooler than the overlying air. Since cold air is denser than warm air, a layer of cold air next to the earth's surface is very stable. This suppresses buoyancy and mixing of the atmosphere. The planetary boundary layer can shrink to a 100 m thickness or less. This can be observed near sunrise on some mornings and is termed an "inversion".

Horizontal movement of wind can also create turbulence and mixing of air. As the fluid atmosphere moves across the solid earth, the horizontal movement creates a shear force on the atmosphere. In response to this force, the air next to the earth begins to roll or tumble creating turbulence. The stronger the wind and the rougher the surface, the more turbulence that is created.

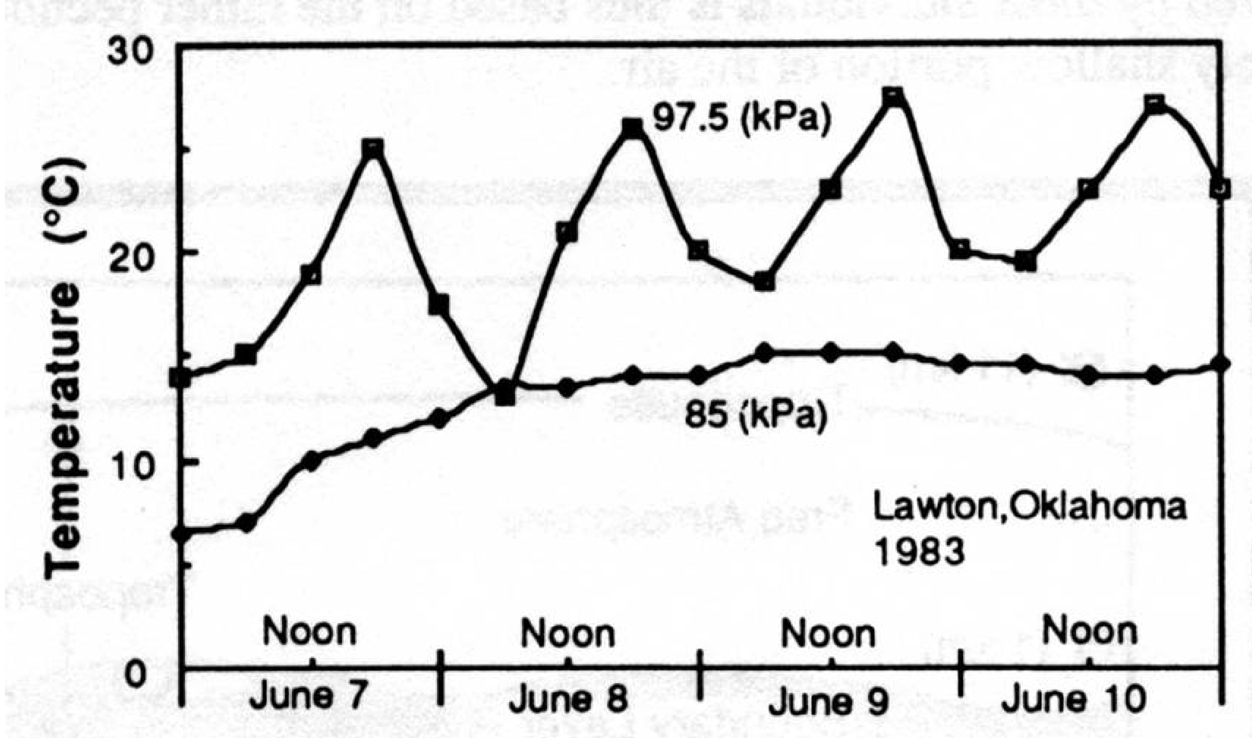


Figure 2.9 Air temperatures near the ground (97.5 kPa) and at a height of approximately 1100 m (85 kPa) (Stull, 1988).

The conditions of the surface (dry, wet, vegetated, bare, snow-covered, etc.) play a significant role in determining whether there is a deep, well-mixed or a shallow, stable planetary boundary layer. There is a cycle of turbulent and stable conditions driven by the diurnal cycle of radiative heating on the surface as well as changes in wind speed over time (Figure 2.10). Therefore, the planetary boundary layer is dynamic and changes rapidly in response to forcing by either radiant energy or shear stress.

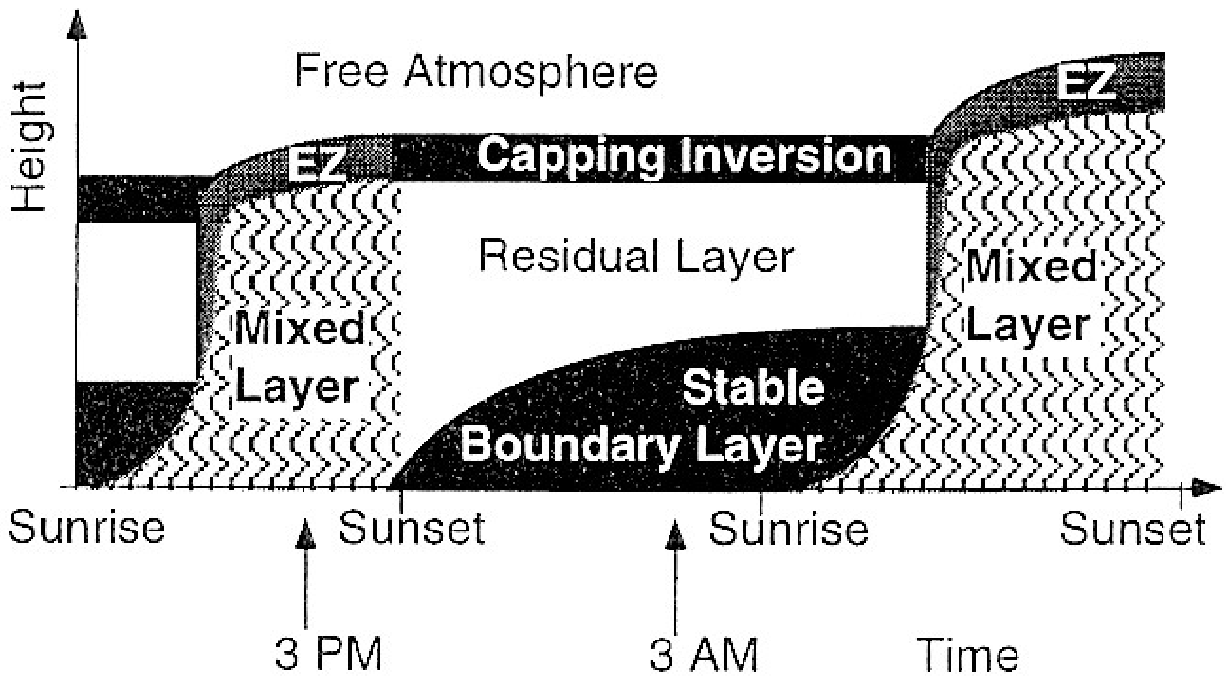


Figure 2.10 Illustration of the planetary boundary layer diurnal cycle (Stull, 2000).

### 2.4.2 Measuring Surface Wind

Surface wind, measured at the international standard height of 10 m above the ground, is usually treated as a two dimensional vector quantity specified by its horizontal direction and speed. Wind direction by convention, is the direction from which the wind is blowing and is referenced to true north. Canadian reports on current weather conditions from principal weather stations contain the windspeed in km $h^{-1}$ and the direction from which the wind is blowing. The wind direction is reported to the closest of 16 compass points (Figure 2.11), which will be one of four cardinal directions (N, S, E, W) or one of four primary intercardinal directions (NW, NE, SW, SE) or one of eight secondary intercardinal directions (NNE, ENE, ESE, SSE, SSW, WSW, WNW, NNW).

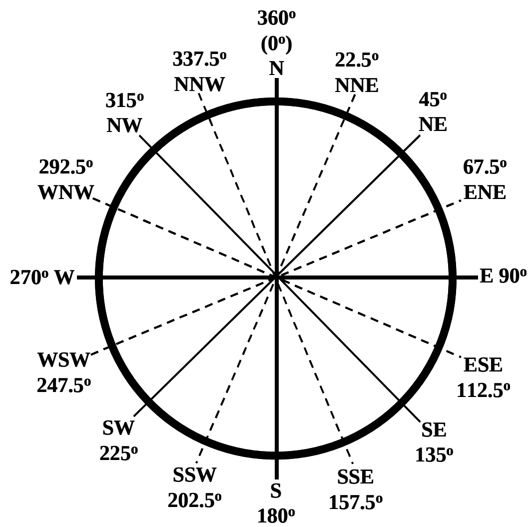


Figure 2.11 Compass bearings of the four cardinal, four primary intercardinal and eight secondary intercardinal directions.

Manned climatological stations use wind sensors usually consisting of an anemometer for speed and a wind vane for direction (Figure 2.12). These sensors provide electrical outputs to drive chart recorders and/or visual indicators from which the observer takes the measurement. An anemometer spins faster as the wind speed increases. The cup anemometer has either a generator in the base that puts out an electric current proportional to the wind speed or sends an electric pulse for every revolution, so the frequency of the pulses is proportional to wind speed. Depending on the type of station and anemometer used, the wind direction and speed for a given hour may be based on the wind run for the entire hour or it may be the 1, 2 or 10 minute average just prior to the report time. Climatological autostations use similar wind sensors as those used in manned stations.

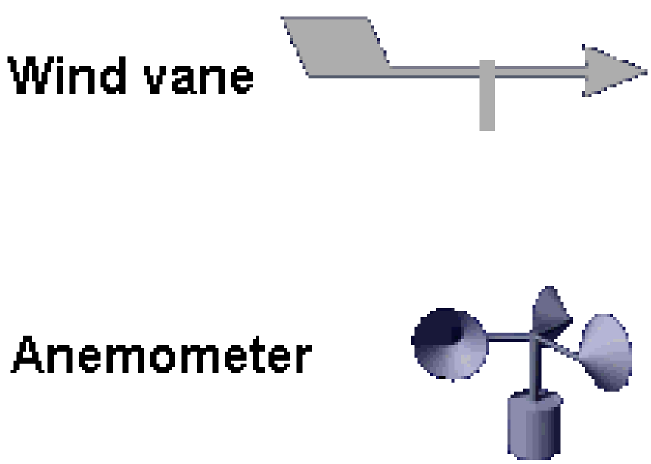


Figure 2.12 Schematics of a wind vane and a cup anemometer.