SPECIAL COURSE

Phy 331

3rd year Atmospheric physics and Astronomy program

Prepared by

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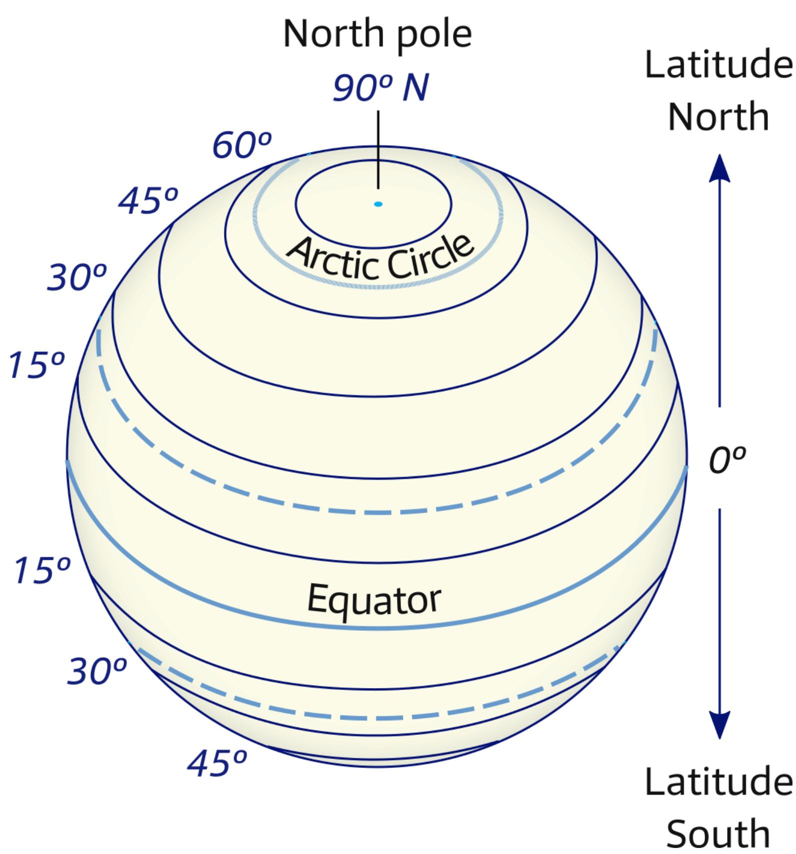
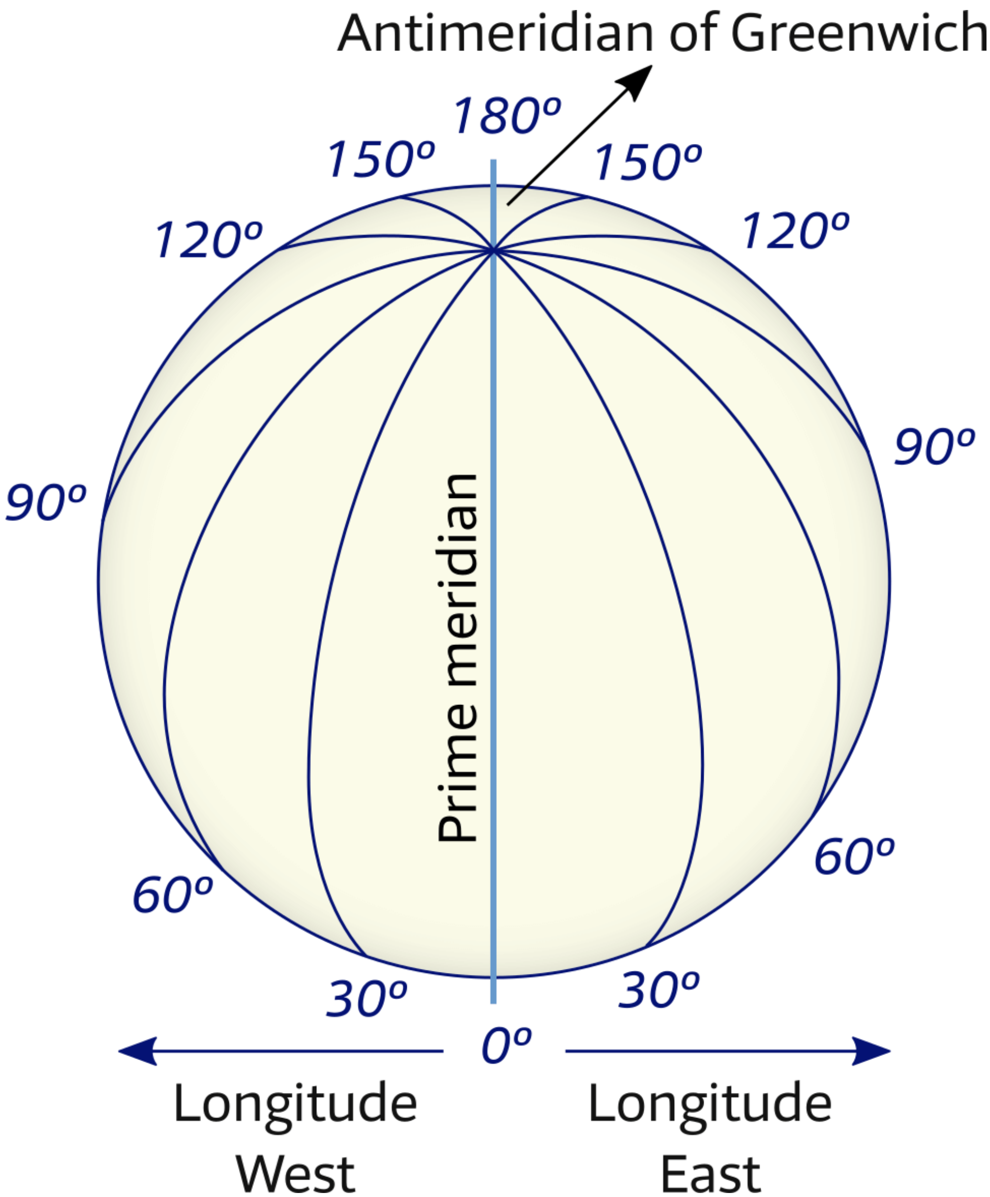
Chapter 1
Great and small circles

Latitude and longitude

The earth can be regarded as a spherical object, and since we're dealing with a 3-dimensional shape we need coordinates of a different form than the usual x- and y-axes. We will use another system of coordinates, meridians and parallels, see figures below.



↑ Parallels of Latitude
The dotted lines are the Tropic of Cancer (North) and Tropic of Capricorn (South)

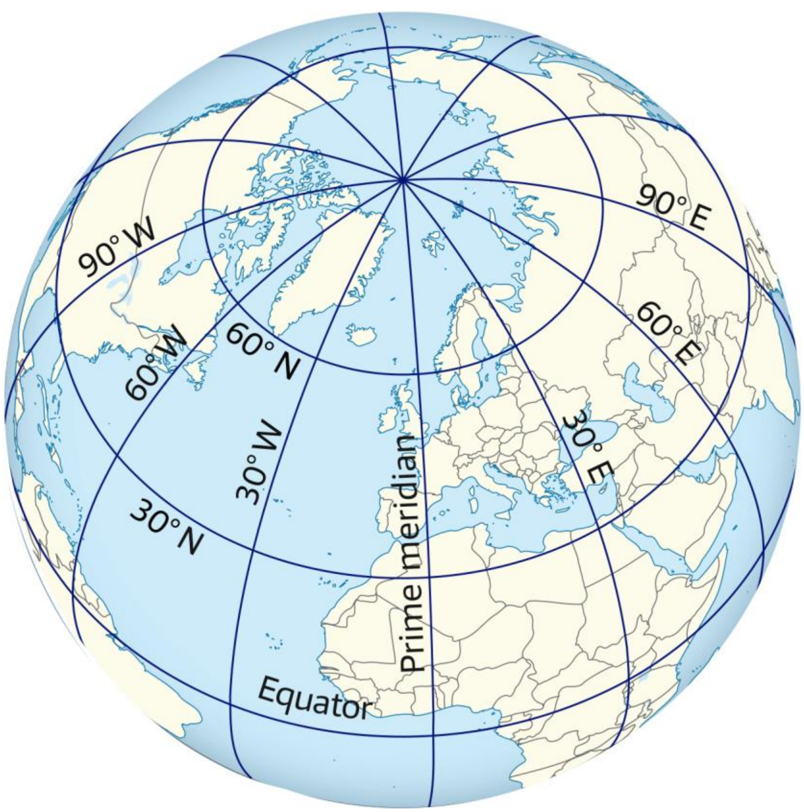


↑ Meridians of Longitude
The Prime Meridian together with the Antimeridian form a great circle.

All these lines together provide the grid which enables us to describe any position in latitudes N↑↓S together with longitudes W←→E.

The obvious place to divide the Northern and Southern Hemispheres was the equator.

It takes the earth 24 hours for a full rotation of 360°. Thus, every hour we rotate 15° longitude. The two world maps beside show the grid and therefore the meridians in 30° intervals: 2 hours.



↑ Northern hemisphere

Northern hemisphere, centered on Greenwich and the Prime Meridian

All these lines together provide the grid that enables us to describe any position in latitudes N↑↓S together with longitudes W←→E.
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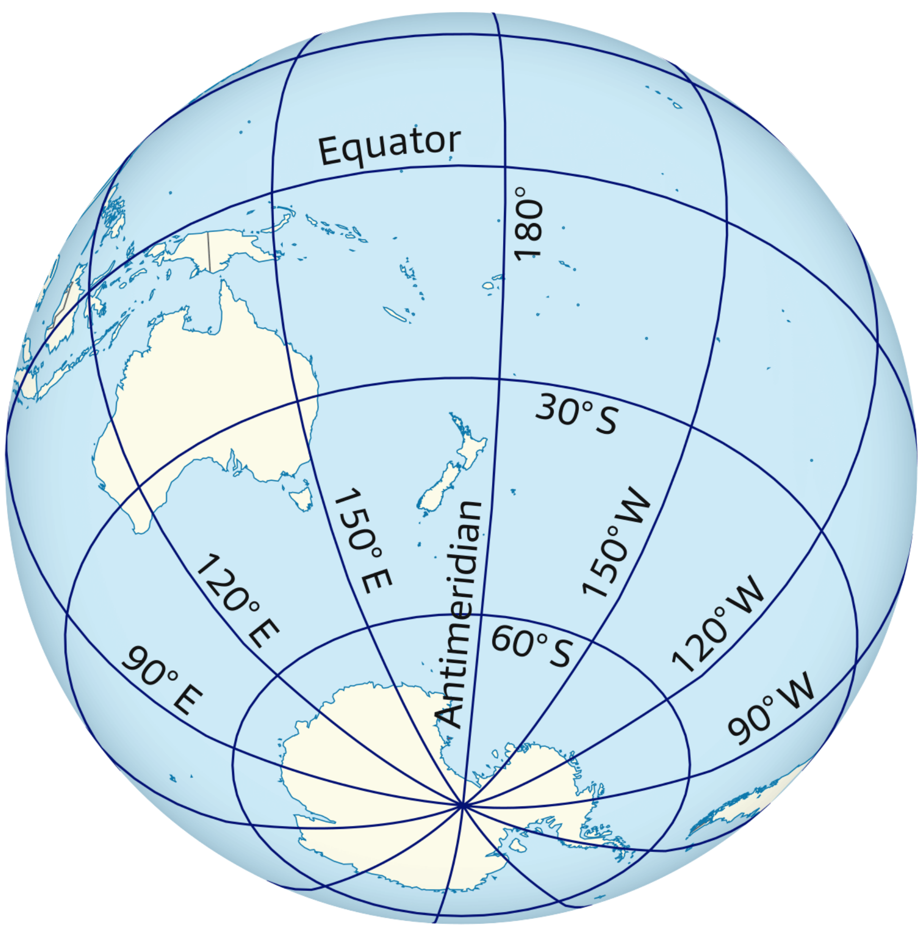
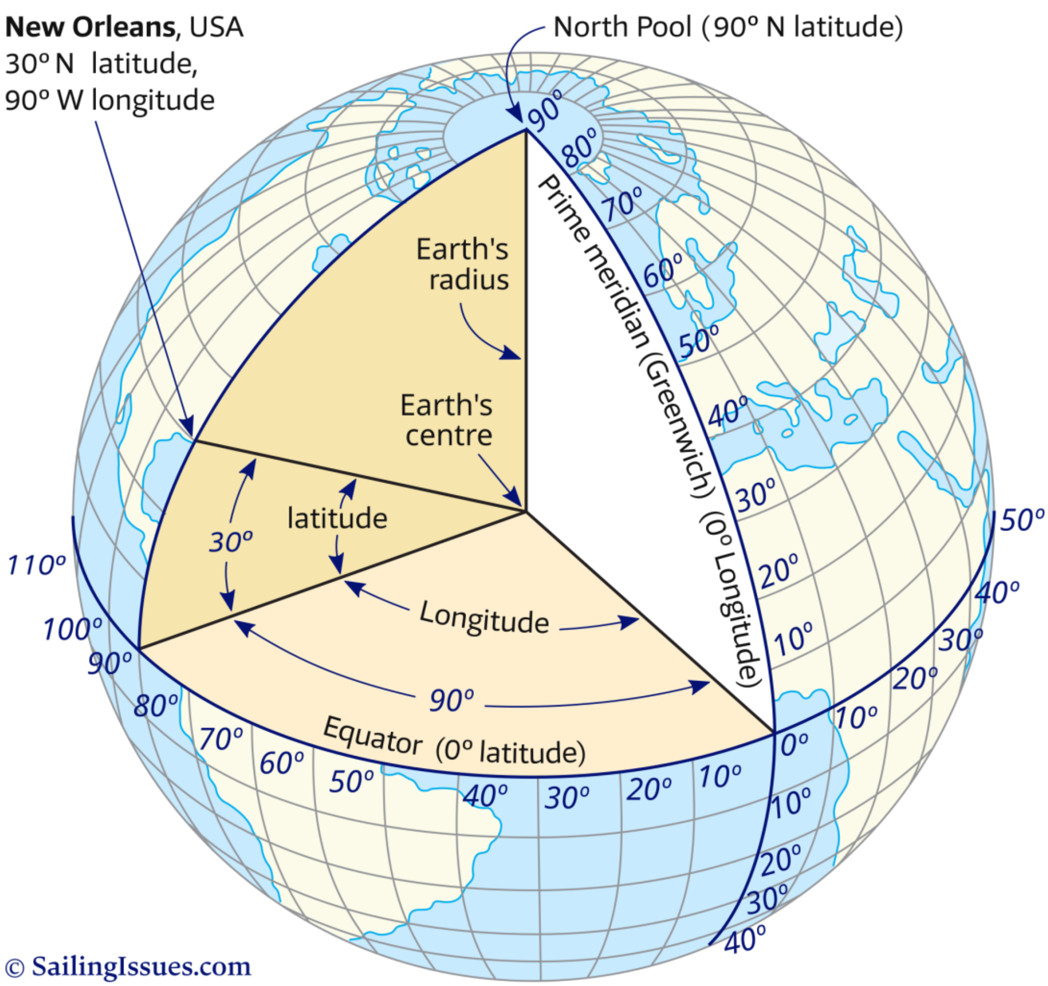


↑ Southern hemisphere
centered on New Zealand and the Antimeridian

For clarity, in the figure below, the position of New Orleans is shown using latitude and longitude only in degrees.

Often more precision is needed with coordinates notated in degrees °, minutes ' (one-sixtieth of a degree) and, optionally, seconds " (one-sixtieth of a minute).

The detailed coordinates of New Orleans are: 29° 57' 20" N 90° 04' 30" W



When it is 12:00 UTC (International Standard Time) – anywhere in the world – it is 12:00 Local Time in Greenwich and 24:00 Local Time

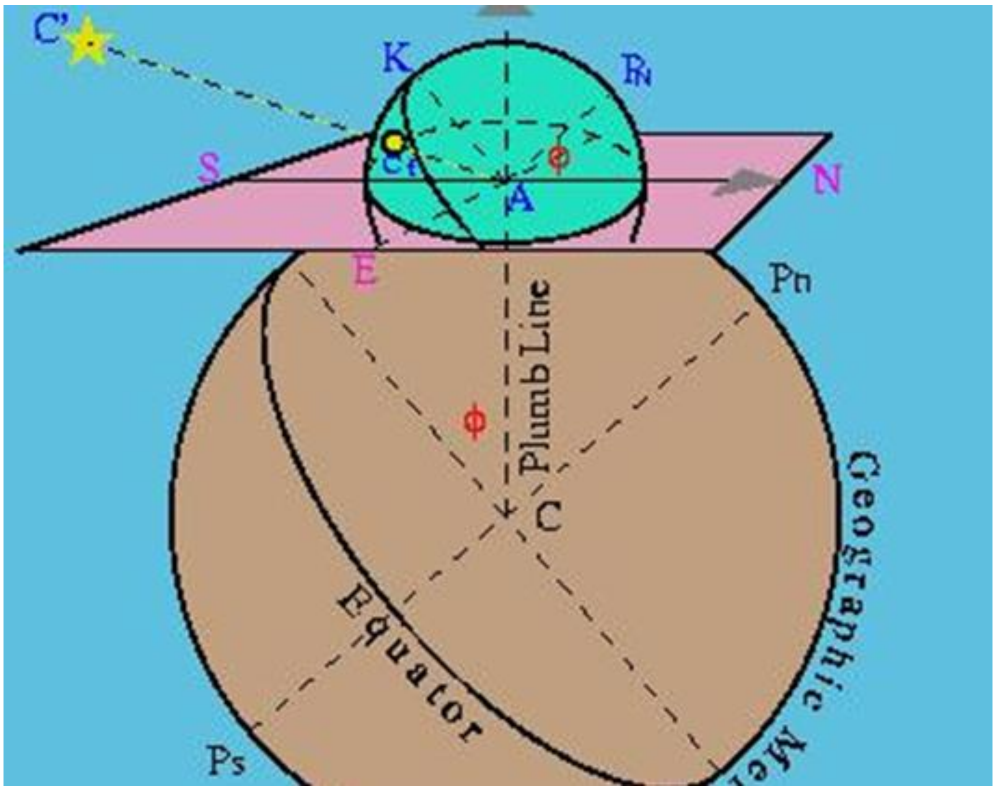
at the other side of the planet: 180° E or 180° W: roughly the International Date Line. Crossing this special meridian changes not only the hour but also the date.
The North Pole has a latitude of 90° N and the South Pole 90° S. The meridians cover twice this angle up to 180° W or E.
Meridians converge at the poles, whereas parallels run parallel to each other and never meet. All meridians and the equator – which is the biggest parallel – form great circles, and the remaining parallels form so-called small circles. A great circle divides the earth into two exact halves.

Glossary

- Parallels: circles parallel to the equator, ranging from 0° to 90° N or S. Only the equator is a great circle.
- Meridians: half-circles converging at the poles, ranging from 0° to 180° E or W. Each pair of opposing meridians forms a great circle.
- Prime meridian: 0° or the Greenwich meridian which – together with the date line meridian – divides the Western and Eastern hemispheres.
- Great circle: the intersection of a sphere and a plane that passes through the sphere's centre.
- Small circle: the intersection of a sphere and a plane that doesn't pass though the sphere's centre.

- Time zones: there are by convention 24 zones, each 15° longitude wide. Hence, noon at Greenwich gives midnight at 180° E.
- GMT, UTC (Coordinated Universal Time), Zulu: the outdated acronym GMT (Greenwich Mean Time) is roughly the same as UTC or Zulu, and is also the local time at Greenwich when daylight saving isn't used. Note that UTC is an atomic time scale which only approximates GMT, so best to use the modern term "UTC".
- Date line: the 180° meridian which extends from or is opposite to the prime meridian. Here, not only the hour changes when crossing the meridian, but also the date. The international date line is affected by the borders of countries.
- Latitude: position property defined by the number of degrees north or south of the equator, varies from 0° to 90°.
- Longitude: position property defined by the number of degrees east or west of the prime meridian, varies from 0° to 180°.
- Position: latitude first and longitude second. For example: Athens in Greece 37° 58' N 23° 43' E.
- Nautical mile: one NM is one minute (') on the vertical scale on the chart. 1' equals 1852 metres. Nautical miles are divided into 10 cables.
- Knots: nautical miles per hour.

The Celestial Sphere



Observer's Horizon: The plane tangent to the earth's surface at the point A and perpendicular to the zenith-Nadir line is the plane of the horizon of the observer.

Celestial Projection: The straight-line AC' represents the direction to a celestial body C'. cf is the projection of C' on the celestial sphere.

Celestial Equator: The plane passing through the celestial sphere that is parallel to Earth's equator is the Celestial Equator (K is on the celestial equator in Figure 1).

Celestial axis and poles: The axis through the center of the celestial sphere perpendicular to the celestial equator is the celestial axis. The poles are at either end of this axis, corresponding to the Earth's poles (PN is the celestial north pole, Pn is the actual north pole).

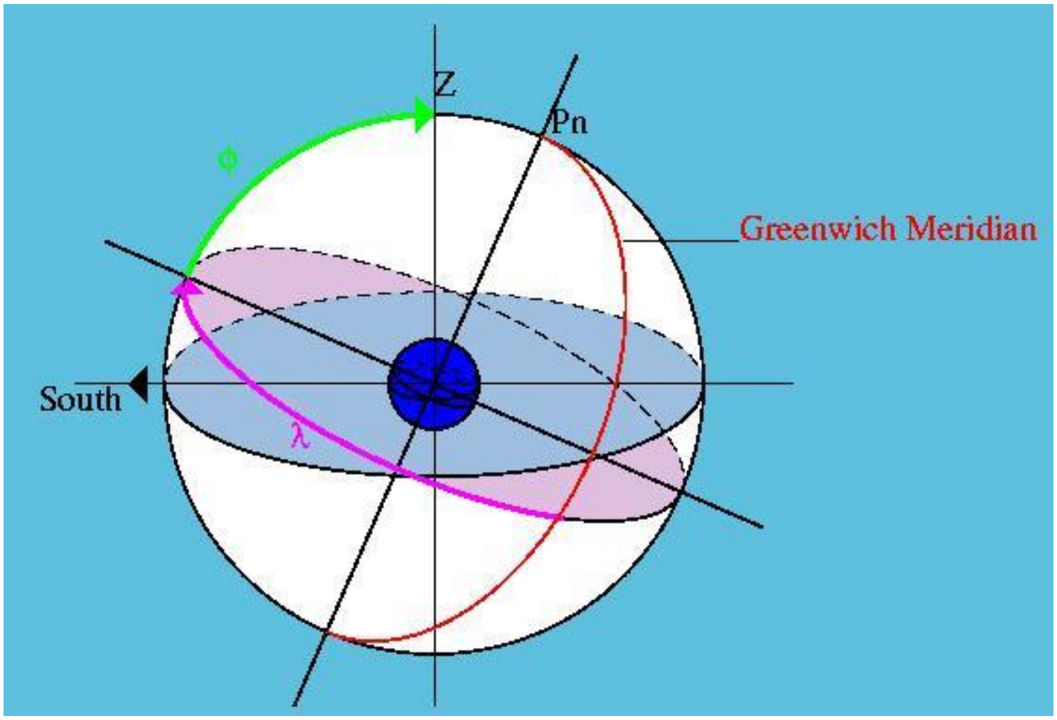
Elevated Pole: The celestial pole that corresponds to the earth pole that is in the same hemisphere (N or S) as the observer. In Figure 1, PN is the elevated pole.

Vertical Circles: Circles passing through the zenith and nadir points are called Vertical Circles.

Meridians: Circles passing through both celestial poles are called Celestial Meridians, they correspond to meridians on the Earth (lines of longitude) which pass through both the north and south pole. The observer's meridian on the celestial sphere is the circle passing through KZPN.

Hour Angles: Hour angles are the spherical angle or the arc of the celestial equator between two meridians. The local hour angle of the body C' above is the arc between the observer's meridian and the meridian passing through cf. Local hour angles are denoted t_{loc}

If the celestial sphere is drawn to be centered at the center of the Earth then the latitude (ϕ) and longitude (λ) of the observer can be found by using the celestial sphere. Latitude is the arc or angle between the closest point on the equator and the observer's zenith. The latitude is followed with a letter N or S depending which hemisphere the observer's meridian is located in.

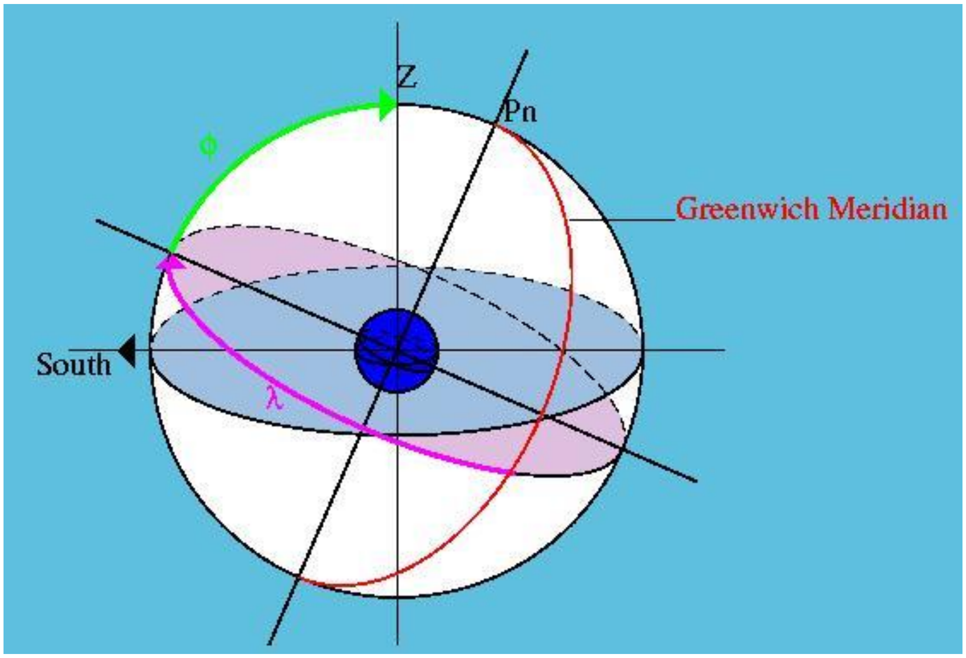


$0 < \phi < 90$

Longitude is the arc or angle between the Greenwich Meridian (GM) and the observer's meridian.

$0 < \lambda < 180$.

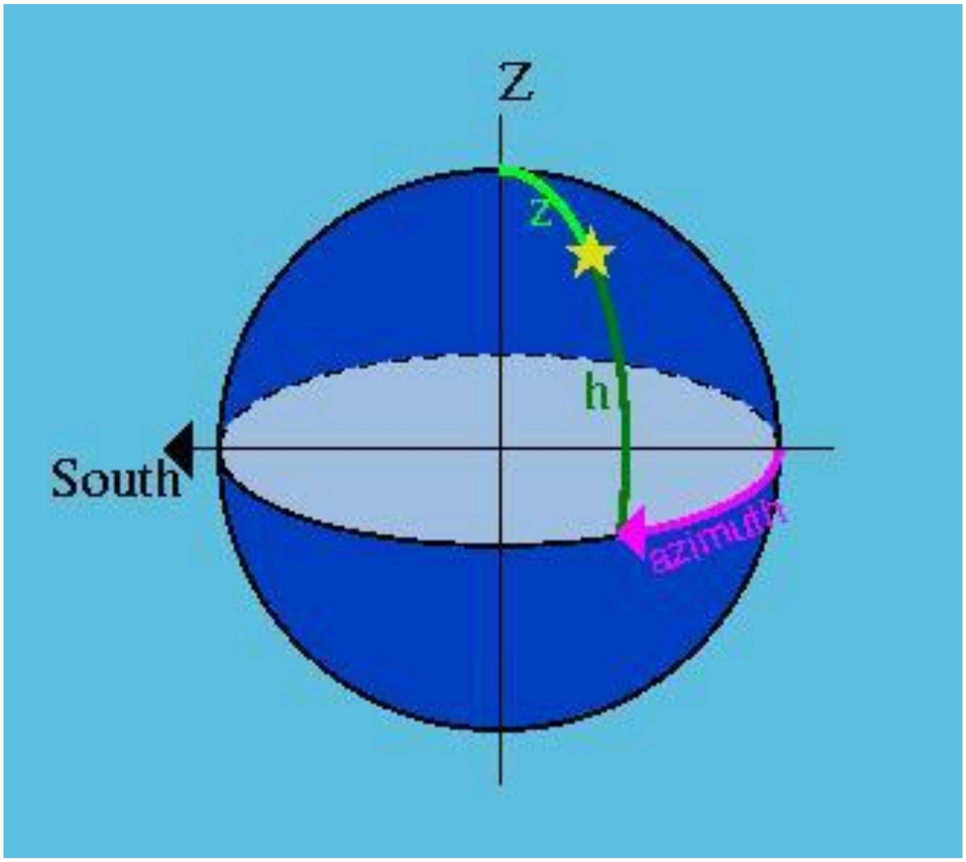
The GM is the meridian dividing the earth into two hemispheres, east and west - much the same as the equator divides the earth into a north and south hemisphere. The longitude is followed with a letter E or W depending on which hemisphere the observer's meridian is located in.



The Azimuth (A) and the Altitude (h) of a celestial body

Two useful coordinates of a particular location are the Azimuth (A) and the Altitude (h). If we have a celestial body (denoted by the yellow star in Figure 3) on the celestial sphere, then its azimuth is the arc of the celestial horizon which lies between the observer's meridian and the vertical circle of the body. A should be measured from the north or south points towards east or west, whichever gives

- $0 \leq A \leq 90$

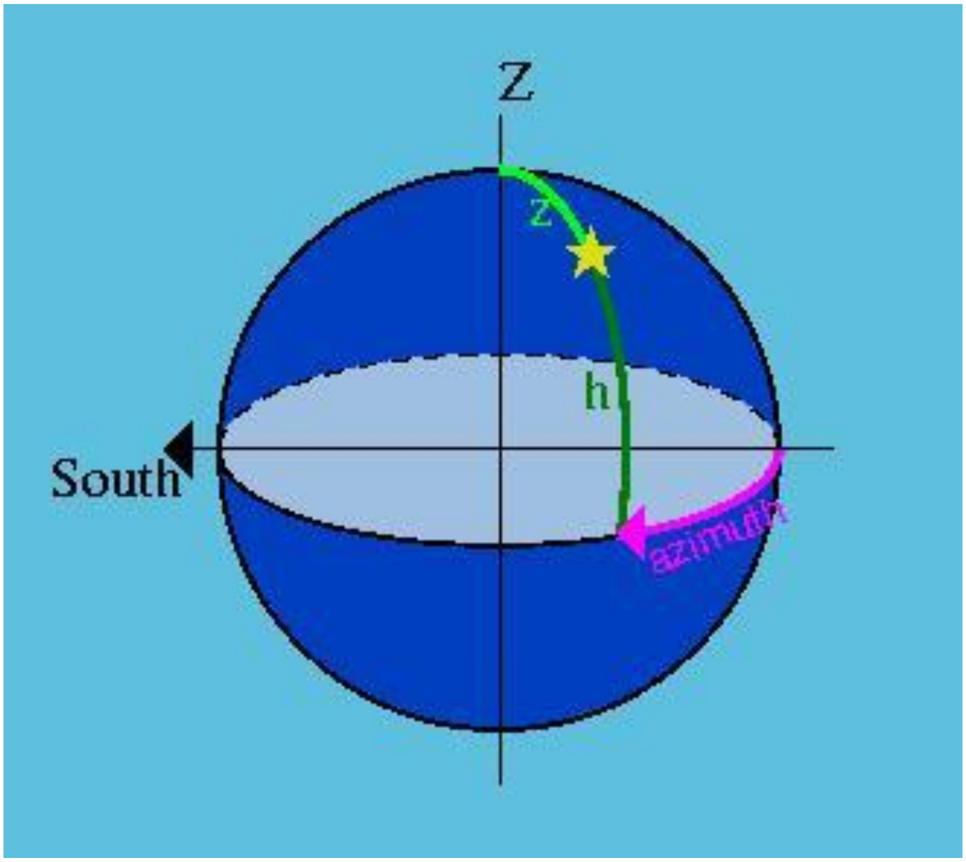


For example, in Figure 3 the azimuth is 65oNE. The altitude of the body is the arc of its vertical circle from the celestial horizon to the place of the body on the sphere.

$$0 \leq h \leq 90$$

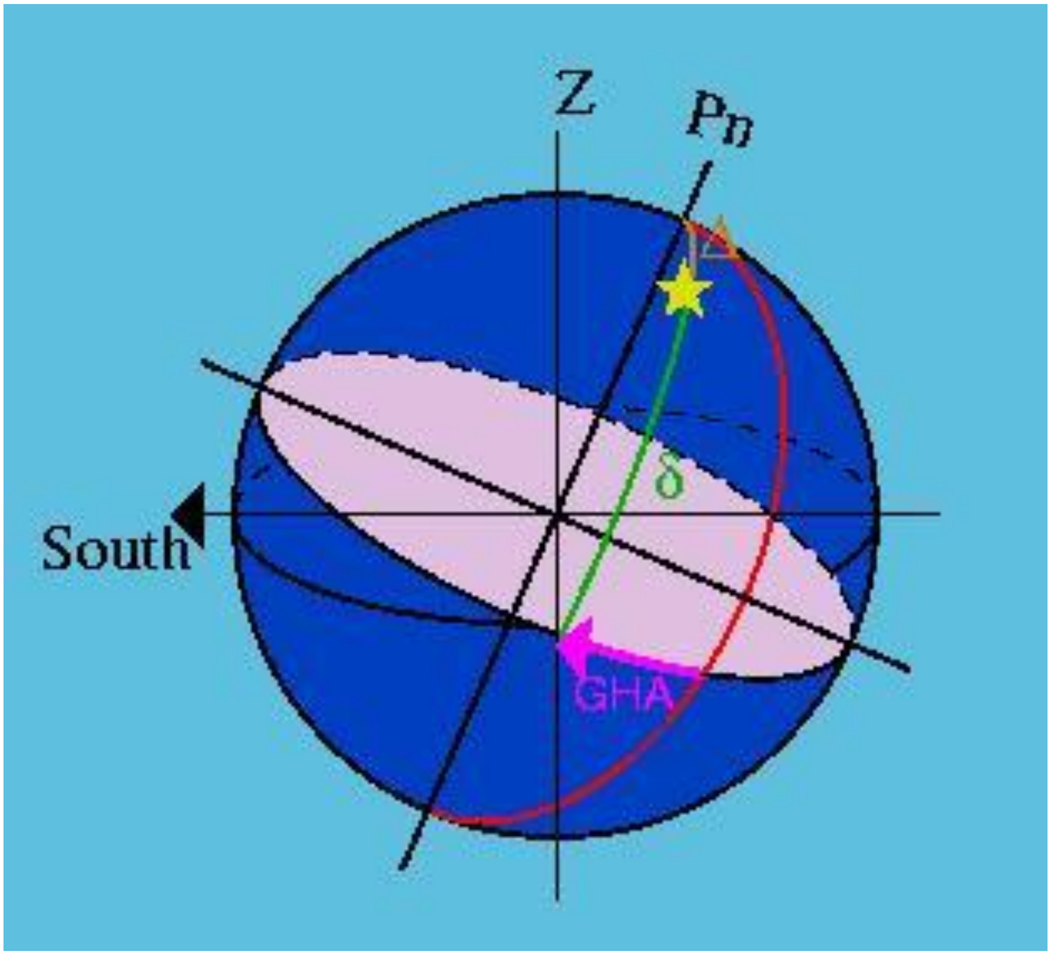
As a coordinate, h can be replaced by zenith distance (z).

z = 90 - h.



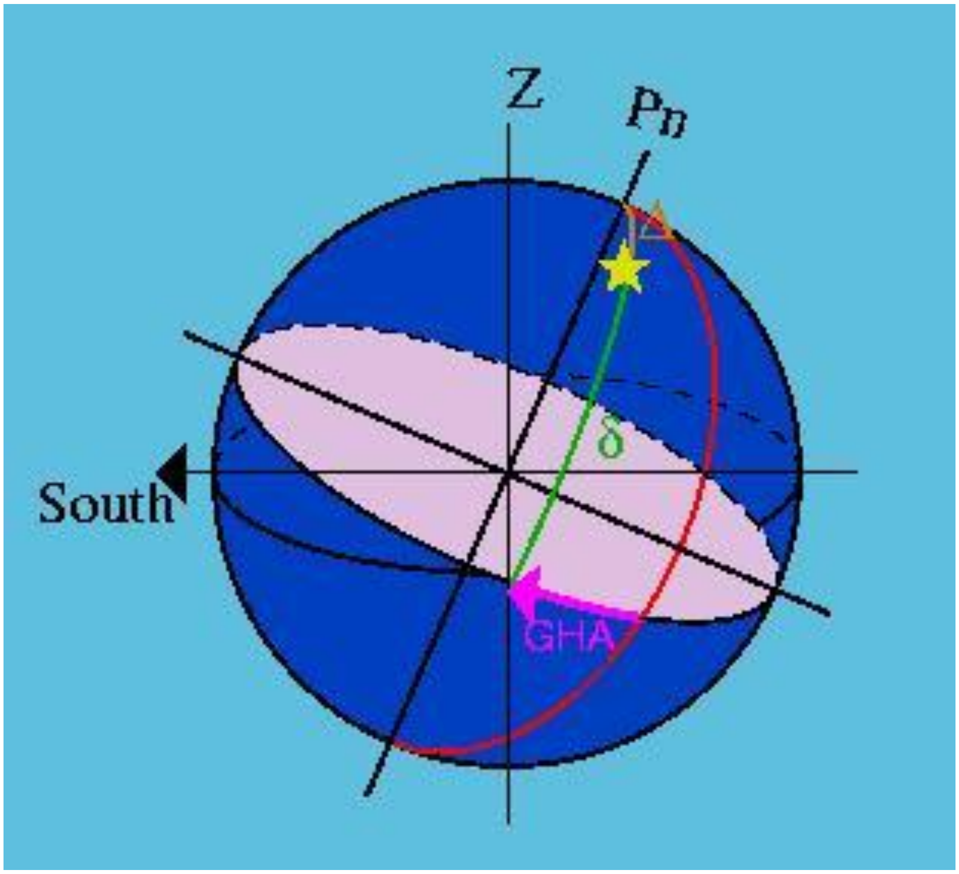
Two other important coordinates are declination (d) and the Greenwich Hour Angle (GHA). The declination of the celestial body (the body is denoted by a star on the Figure) is the arc of the body's meridian from the celestial equator to the place of the body on the sphere. In this respect the declination and latitude of the body are similar. However:

- $-90 \leq d \leq 90$



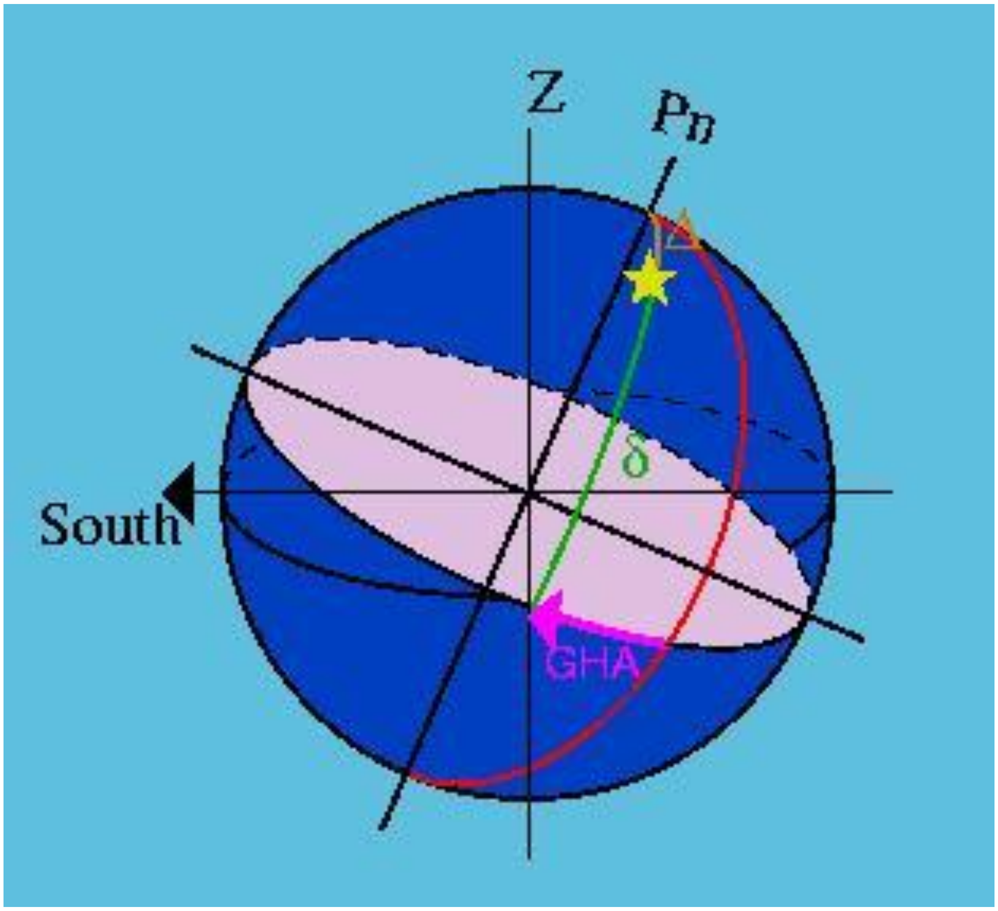
If the body is in the same hemisphere (N or S) as the elevated pole then d is positive, otherwise d is negative. The polar distance (D) is the complementary to the declination:

$$\Delta = 90 - \delta$$
$$\delta = 90 - \Delta$$



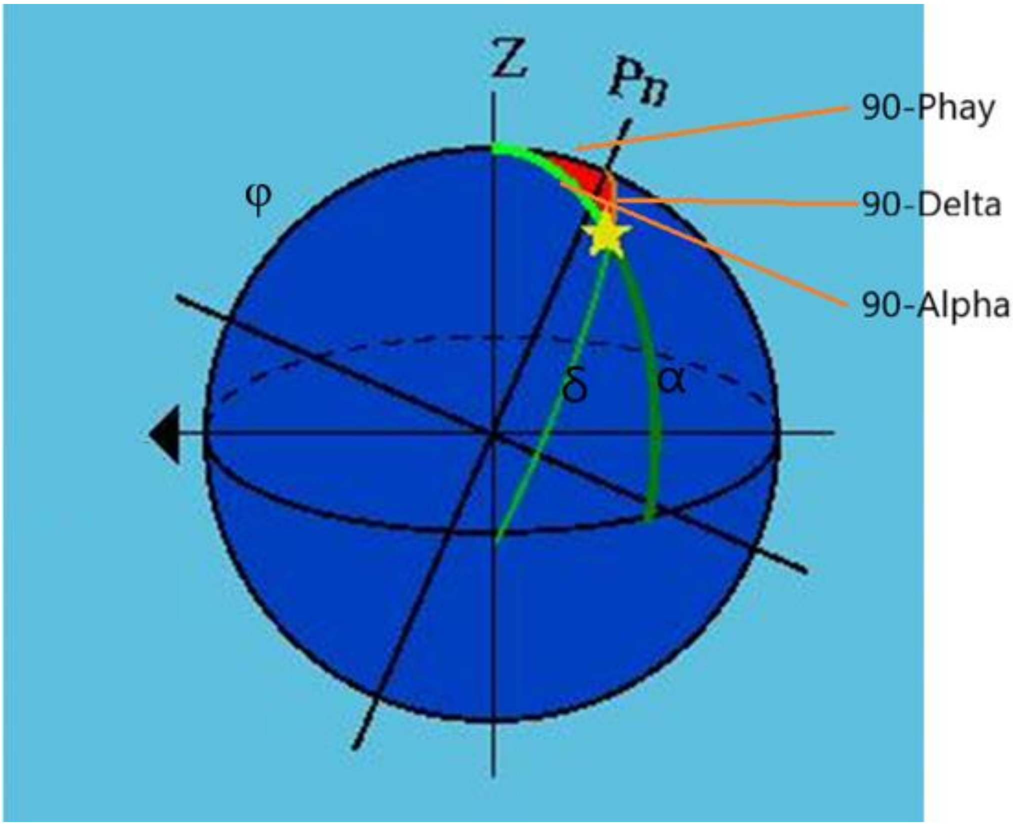
The Greenwich Hour Angle of the body is the arc on the celestial equator between the Greenwich Meridian and the celestial meridian of the body. It is called the hour angle because due to the earth's rotation, 15 degrees of longitude corresponds to 1 hour. i.e.: if the celestial sphere remained stationary and centered on Earth while the earth rotated, the GHA of the celestial body would increase by 15 degrees every hour.

$0 < GHA < 360$



Spherical triangle

When drawn on the same sphere the vertical circle of the body and the celestial meridian of the body form a spherical triangle (shown in red in Figure 5a above). The three corners are the zenith (Z), the elevated pole, and the location of the body on the celestial sphere.



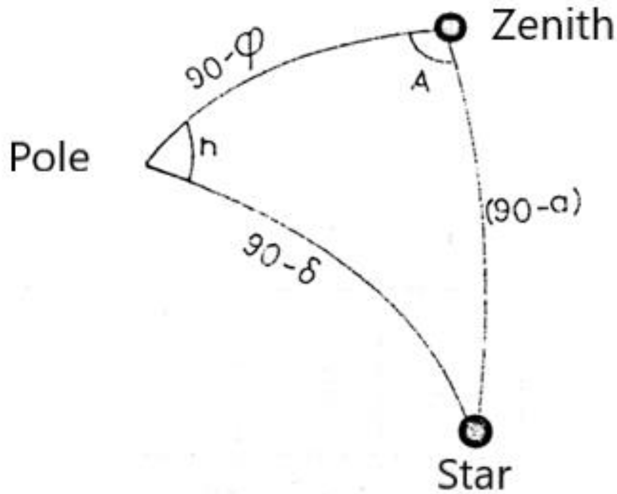
Laws of the spherical triangle

The first law can be used for calculating any arc length as long as the rest two arc lengths and the angle between them are known. The second law is used for calculating any angle or arc length in the spherical triangle when we know the value of two arcs and an angle or two angles and an arc.

1) $sin \propto = \sin\varphi \sin\delta + \cos\varphi \cos\delta \cos h$

2) $\frac{\cos\propto}{\sin h} = \frac{\cos\delta}{\sin A}$

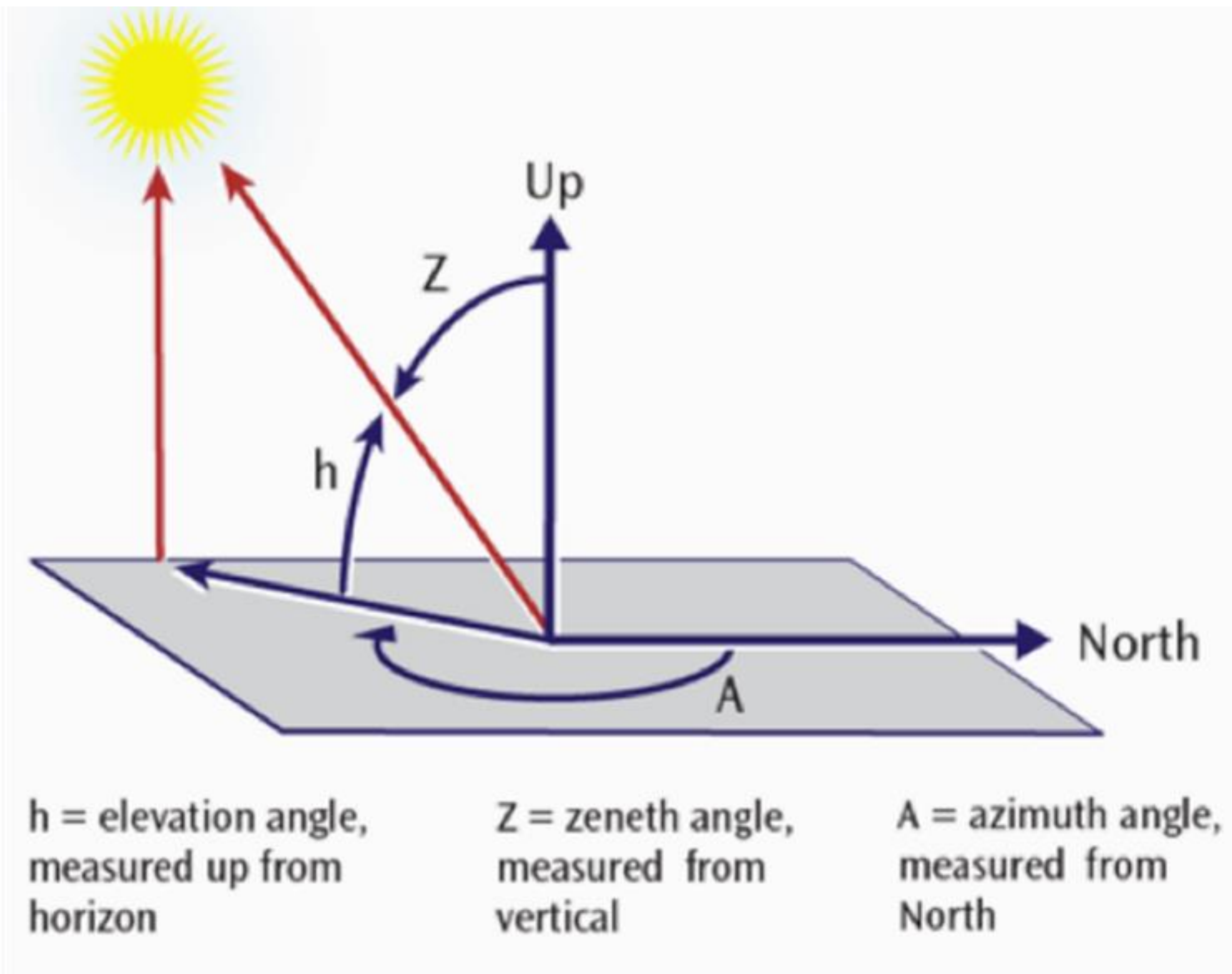
3) $\sin\varphi \sin\delta = \cos\varphi \tan\delta - \sin h \tan A$

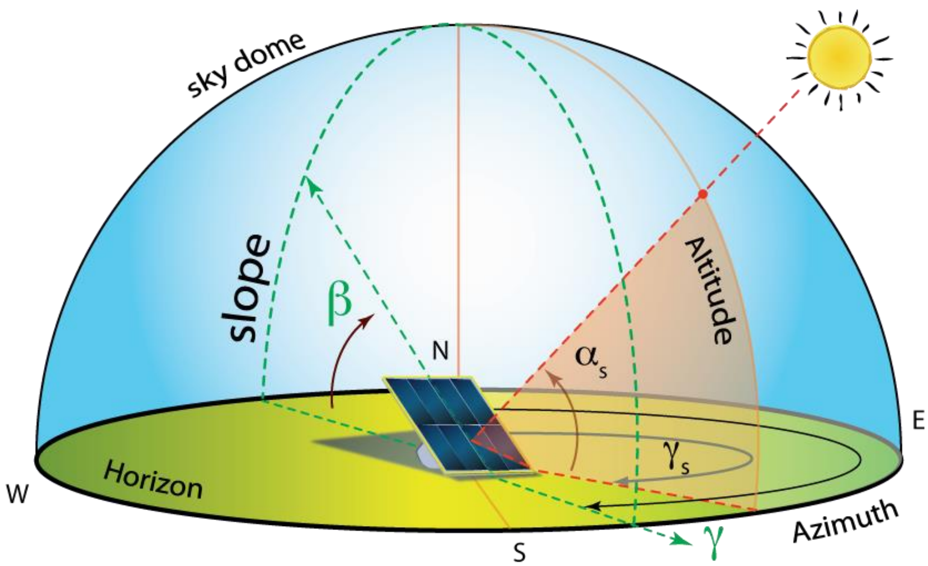

Zenith
90-φ
A
Pole
h
(90-a)
90-δ
Star

Altitude and azimuth angles

Altitude angle

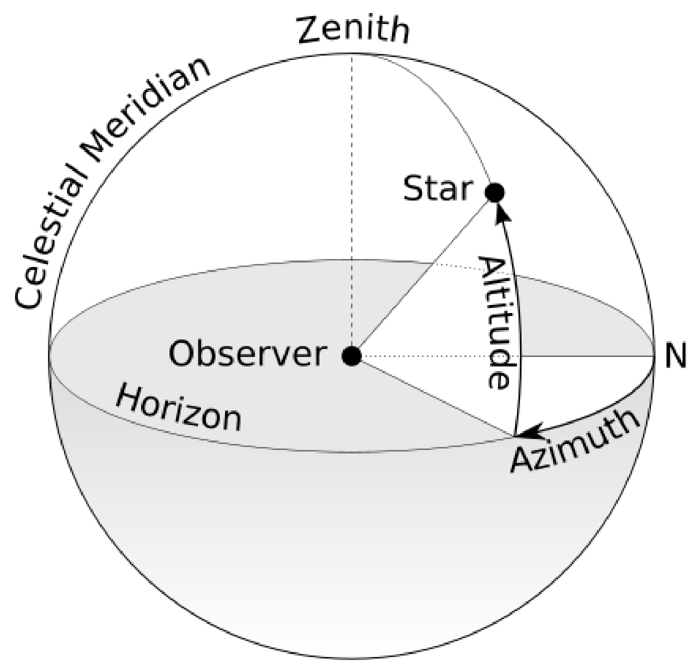
- It is the angle between the horizontal plane at the measurement location and the line connecting the viewer and the sun's disk. It is symbolized in the following figure by the symbol h.





Azimuth angle

- Azimuth varies from 0° to 360°. It starts with North at 0°. As you turn to your right (in a clockwise direction) you'll face East (which is 90°), then South (which is 180°), then West (which is 270°), and then return to North (which is 360° and also 0°). So, if the Azimuth for your satellite is, say, 45°, that means your satellite is northeast of you.

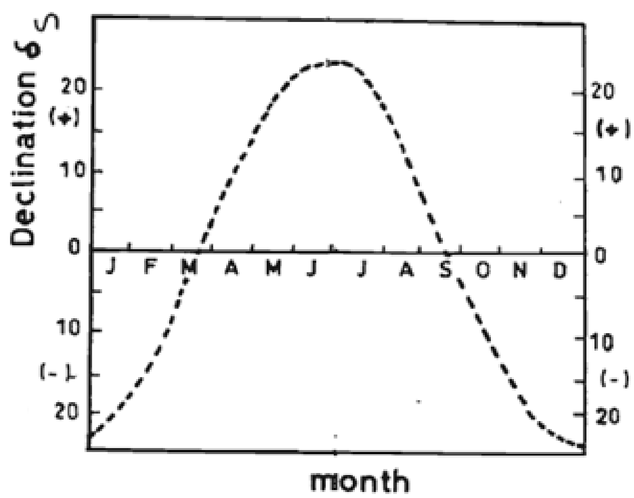


- The sun’s elevation angle and azimuth angle depend on:
- 1-Solar declination angle.
- 2- hour angle.
- 3- Latitude line angle.
- We will explain how to calculate these angles

Solar declination angle

- The declination (δ) is the angular distance of an object perpendicular to the celestial equator, positive to the north, negative to the south.
- Its value ranges from +23.5 north to -23.5 south.

- It is numerically equal to the latitude to which the star is perpendicular.



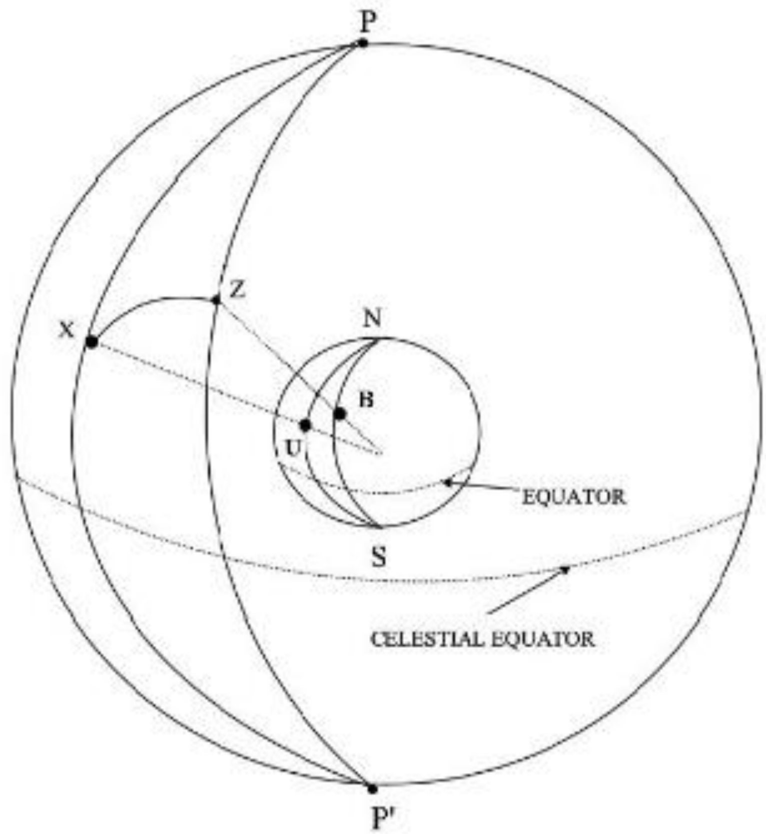
- From the graph, it is clear that the declination angle changes during the year, reaching its maximum on June 21, and its value is 23.5 degrees, when the sun is perpendicular to the Tropic of Cancer.
- The lowest possible is -23.5 degrees on December 21 when the sun is perpendicular to the Tropic of Capricorn.
- It is equal to zero on both March 21 and September 21 when the sun is perpendicular to the equator.
- The solar declination is calculated from the following relationship:

$$\boxed{\delta_s = 23.45 \sin\left[360 \frac{n-82}{365}\right]}$$

- Where n is the day number in the year starting from January 1 = 1

Local Hour Angle

- Local Hour Angle (LHA). In astro navigation, we need to know the position of a celestial body relative to our own position.
- LHA is the angle BNU on the Earth's surface which corresponds to the angle ZPX in the Celestial sphere. In other words, it is the angle between the meridian of the observer and the meridian of the geographical position of the celestial body (GP).
- Due to the Earth's rotation, the Sun moves through 15° of longitude in 1 hour
- So the angle ZPX can be measured in terms of time and for this reason, it is known as the Local Hour Angle.



Hour angle calculation

- To calculate the hour angle at a certain time, T hour:
- The number of longitudinal lines crossed by the sun in one-hour 360/24 is multiplied by the result of subtracting the noon time 12 from the time at which the hour angle is to be calculated.

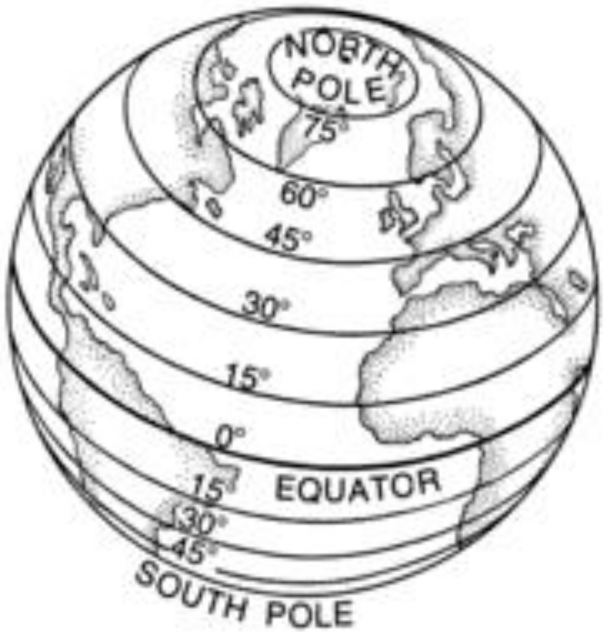
$$\boxed{h_s = 15(T - 12)}$$

Equation of time

- The equation of time describes the discrepancy between two kinds of Solar time. The two times that differ are the apparent solar time, which directly tracks the diurnal motion of the Sun, and mean solar time, which tracks a theoretical mean Sun with Noons 24 hours apart.
- Apparent solar time can be obtained by measurement of the current position (hour angle) of the Sun, as indicated (with limited accuracy) by a sundial.
- Mean solar time, for the same place, would be the time indicated by a steady clock set so that its differences from apparent solar time would resolve to zero over the year.

Latitude angle

- Definition of angle of latitude:
- It is the angle of latitude of the place of measurement that can be known from the geographic atlas.
- The circles of latitude are imaginary parallel circles, numbering 180: 90 of which are north of the equator and 90 south of the equator. The equator represents zero degrees (0), between one circle and the other is one degree, or the equivalent of 111 km on the surface of the Earth.



The azimuth angle calculation

$$\boxed{\sin\alpha_s = \frac{\cos(\delta_s)\sin(h_s)}{\cos(\alpha)}}$$

- $\boxed{\delta_s}$ is the angle of inclination of the sun, hs is the hour angle, α is the angle of solar elevation above the horizon, it is possible for this angle to be higher than 90 degrees on days when the daylight is more than 12 hours.

Example 1:

Find the sun's elevation angle and azimuth angle at solar time 9, 10, 11, 12 hr for a place with latitude L = 40 N on Decembe

Solution:

The order of December 25th during the year is 360, so we can calculate the angle of declination of the sun from the relationship:

Its value will be -23.3914

The hour angle is calculated from the relationship hs=15(12-T)

At nine o'clock solar time, it will be

hs=45^0.

The angle of solar elevation is then calculated from the relationship

$$\boxed{\sin\alpha = \sin(L)\sin(\delta_s) + \cos(L)\cos(\delta_s)\cos(h_s)}$$

So its value is 14.0 degrees

The azimuth angle is then calculated from the relationship

$$\boxed{\sin\alpha_s = \frac{\cos(\delta_s)\sin(h_s)}{\cos(\alpha)}}$$

So its value is 42^0

Example:

Calculate the solar elevation and azimuth angles for a location at latitude 37N at the third hour of solar noon on the twentieth of February.

Solution:

The angle of declination of the sun is calculated from the relationship:

$$\boxed{\delta_s = 23.45 \sin\left[360 \frac{n-82}{365}\right]}$$

Where the day's number is compensated

n = 31+20=51

So the declination angle = -11.57 degrees.

Then the hour angle is calculated from the relationship:

hs=15(T-12)

At the third hour of the solar noon (15 solar time), hs = 45 degrees.

Then the solar elevation of the sun is calculated from the relationship:

$$\boxed{\sin\alpha = \sin(L)\sin(\delta_s) + \cos(L)\cos(\delta_s)\cos(h_s)}$$

So its value = 25.6 degrees.

From the following equation, we calculate the azimuth angle

$$\sin \alpha_s = \frac{\cos(\delta_s)\sin(h_s)}{\cos(\alpha)}$$

$$\sin(\alpha_s = \frac{\cos(-11.57)\sin(45)}{\cos(25.6)}$$

So its value is 50.2 degrees

The hour of sunset

Definition of sunset hour:

It is the number of hours from solar noon to the time when the sun sets behind the horizon when the angle of elevation of the sun is equal to zero.

It can be calculated from the equation:

$$Hourset = \frac{\cos^{-1}(-\tan(\delta_s)\tan(L))}{15}$$

According to the previous example, the hour of sunset is:

$$Hourset = \frac{\cos^{-1}(-\tan(-11.57)\tan(37))}{15}$$

That is, equal to 5.5 h

That is, sunset time is five hours and thirty minutes after noon.

The length of the day, i.e. between sunrise and sunset, can be calculated as follows: 5.5 x 2 = 11 hours

That is, sunset is at 5 hours and thirty minutes and sunrise is at 6 hours and thirty minutes in the morning at latitude 37 degrees north.in

Example:

Calculate the local times of sunrise and sunset for a city located at latitude and longitude (40.78N, 73.97W) on the twenty-first of July.

Solution:

On the twenty-first of July, the day number is 202.

The angle of inclination of the sun is calculated from the relationship:

$$\boxed{\delta_s = 23.45 \sin\left[360 \frac{n-82}{365}\right]}$$

So the angle of inclination of the sun is equal to 20.44 degrees.

The sunset hour is calculated from the equation:

$$\boxed{Hourset = \frac{\cos^{-1}(-\tan(\delta_s)\tan(L))}{15}}$$

So it is equal to 7.25 h

Therefore, the sunset hour is 7.25 solar time and the sunrise hour is 4.75 solar time.

To convert to local time (Standard time), a value is added or subtracted, i. e. correction must be made:

Correction in hours for solar time = (Local meridian -standard meridian)/15

That is (73.97-75)/15=-0.06 h

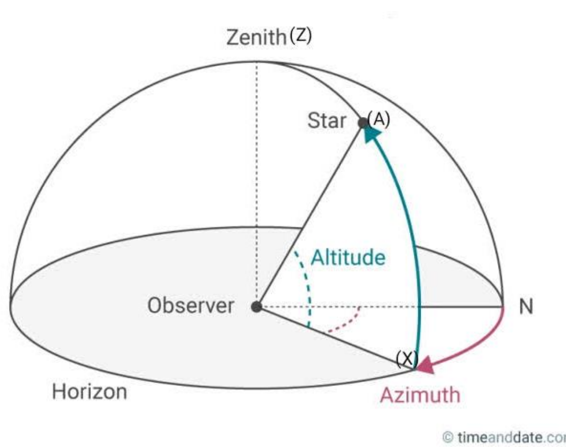
Then the sunset hour becomes = 7.19 pm, and the sunrise hour becomes 4.69 am.

Celestial Coordinate Systems

With the basic information about the celestial sphere in hand, we are now in a position to understand the three basic celestial coordinate systems. So, let's proceed, beginning with the Horizontal system as follows:

The Horizontal Coordinate System (Altitude and Azimuth

Let Z denote the Zenith and A denote the position of any star. If we draw a draw circle ZAX, then the position of a star can be defined on the celestial sphere either by the arc NX, the arc AX, or by the arc ZA and the angle NZA. The arc AX measured along the great circle ZAX represents the angular distance of the star from the horizon. This is known as its Altitude. The complementary arc ZA is called the Zenith distance of the star.



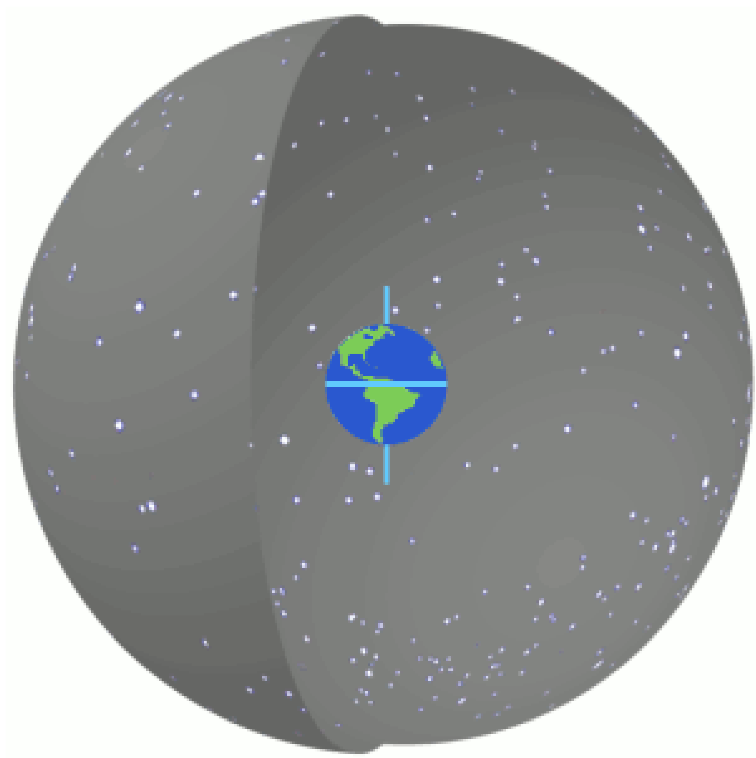
The arc NX of the horizon between the North point and the foot of the great circle through the star or the angle NZA between the meridian and great circle is called the azimuth of the star. It is measured either eastwards or westwards, from the north point. In this way, the positions of stars are defined in the Horizontal celestial coordinate system.

The most basic coordinate system is the Alt-Az system (Altitude and Azimuth) and is useful if you are doing naked-eye or binocular observing. First, you need to know in which direction to stand (North, East, SSW, 3 degrees clockwise from North, etc.). This is the Azimuth. Then you need to know how many degrees above the horizon you should look (the Altitude).

The problem with this system is that not only is it dependent on where in the world you are standing, but the sky is also moving with respect to you. This means that the coordinates that you have been given have to be calculated knowing where and when the observation is about to take place.

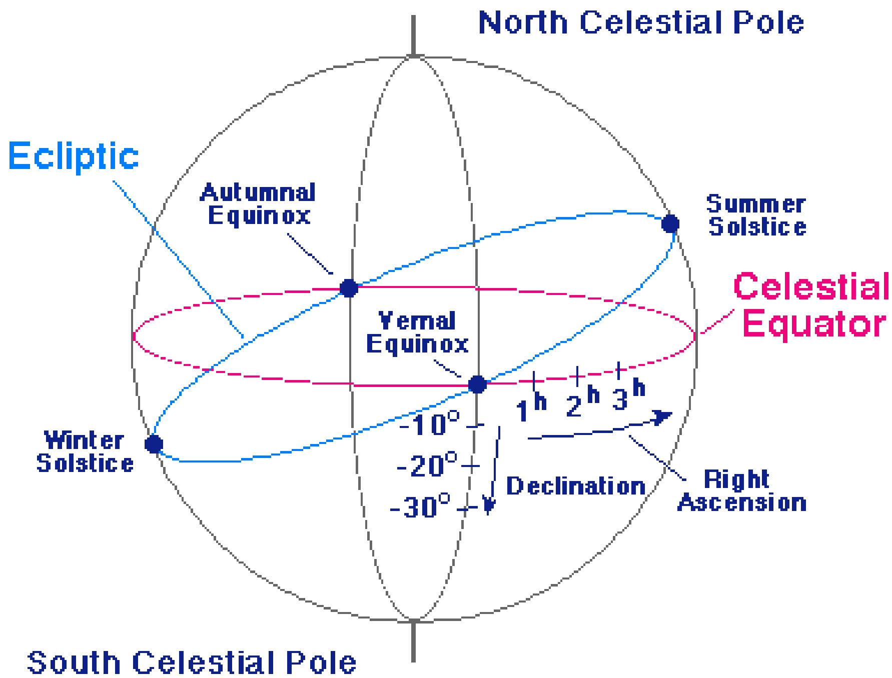
The Equatorial Coordinate System (Right Ascension and Declination)

The equatorial coordinate system uses spherical coordinates. The fundamental plane is formed by the projection of Earth's equator onto the celestial sphere, forming the celestial equator. The primary direction is established by projecting Earth's orbit onto the celestial sphere, forming the ecliptic, and setting up the ascending node of the ecliptic on the celestial equator, forming the vernal equinox. Right ascension is measured eastward along the celestial equator from the equinox, and declination is measured positive northward from the celestial equator. (Two such coordinate pairs are shown here.) Projections of the Earth's north and south geographic poles form the north and south celestial poles, respectively.



Equinoxes and Solstices

The zero point for celestial longitude (that is, for right ascension) is the Vernal Equinox, which is that intersection of the ecliptic and the celestial equator near where the Sun is located in the Northern Hemisphere Spring. The other intersection of the Celestial Equator and the Ecliptic is termed the Autumnal Equinox. When the Sun is at one of the equinoxes the lengths of day and night are equivalent (equinox derives from a root meaning "equal night"). The time of the Vernal Equinox is typically about March 21 and of the Autumnal Equinox about September 22.



The point on the ecliptic where the Sun is most north of the celestial equator is termed the Summer Solstice and the point where it is most south of the celestial equator is termed the Winter Solstice.

In the Northern Hemisphere the hours of daylight are longest when the Sun is near the Summer Solstice (around June 22) and shortest when the Sun is near the Winter Solstice (around December 22). The opposite is true in the Southern Hemisphere.

The term solstice derives from a root that means to "stand still"; at the solstices the Sun reaches its most northern or most southern position in the sky and begins to move back toward the celestial equator. Thus, it "stands still" with respect to its apparent North-South drift on the celestial sphere at that time.

A more useful system for astronomers is the Equatorial system which is based (sometimes loosely) on the celestial poles, where the sky appears to be stationary.

This system is fixed on the sky so is useful for longer-term descriptions of sky position.

In the Equatorial system declination is measured from the celestial equator, with the North Celestial Pole therefore being at +90 degrees. The other coordinate needed is Right Ascension which is measured from the vernal equinox. Traditionally, Right Ascension is measured in hours with a full circle being 24 hours.

Since then, the IAU have decided to move away from using the Earth's pole to define the coordinate system and adopt a non-rotating system that is based on very distant objects. This is the International Celestial Reference System (ICRS) and, to a good approximation, has been defined to have the same pole as the J2000 system. The IAU adopted the ICRS in 2000. Since this system is fixed in space and is not linked to the Earth's pole, it does not have an equinox associated with it. In order to define the ICRS, the ICRF (International Celestial Reference Frame) was published and is the realization of the ICRS and consists of 212 defining extragalactic radio sources. This means that the system is effectively defined by the positions given for these sources in this publication. An update of the realization of the ICRS occurred in 2009 and is called ICRF2 and consists of 3414 compact radio sources. See The Gaia Celestial Reference Frame (Gaia-CRF2) for infortmation about the Gaia Celestial Reference Frame (Gaia-CRF2), defined by means of the Gaia Data Release 2 measurements for 556,869 quasars.

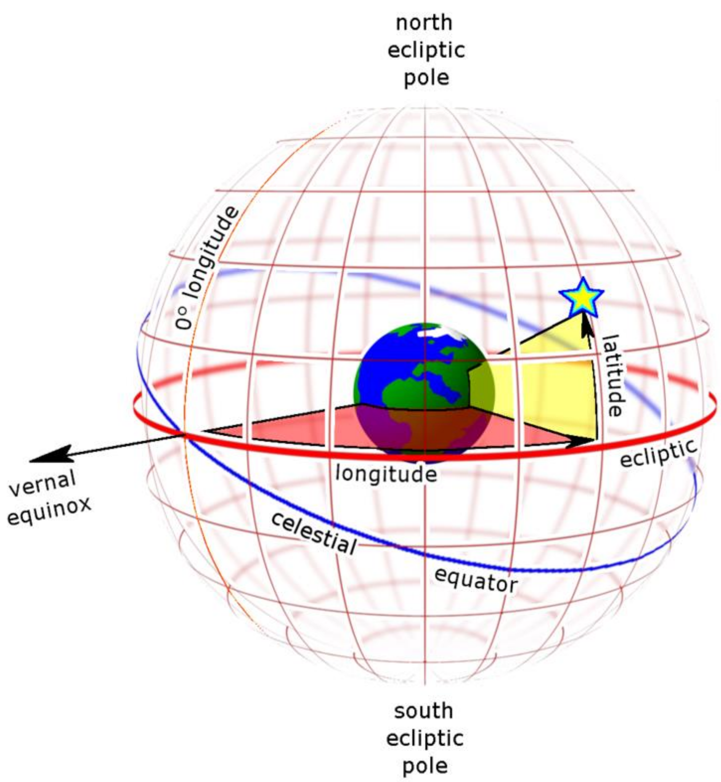
Additionally, there are two other coordinate systems that are used that can be helpful when trying to understand the context of a particular source. These are the ecliptic and Galactic coordinate systems. Both are described by longitudes and latitudes, but have different planes defining zero latitude. In the former case, the plane is the ecliptic, where the orbits of the Earth and the planets approximately lie. The latter case uses the Galactic plane. The longitude zeropoints for these systems are the vernal equinox and the Galactic centre respectively. An example of the use of these systems is if you want to avoid the crowding that occurs in the Galactic plane, you might exclude sources with an absolute Galactic latitude smaller than 10 degrees. Note that, in general, these two coordinate systems are not calculated to any great accuracy since only an approximate idea of where on the sky with respect to these planes and poles is usually needed.

Vernal equinox: The point where the Sun crosses the equator from south to north is called the vernal (or spring) equinox.

TT: Terrestrial Time (TT) is a modern astronomical time standard defined by the International Astronomical Union, primarily for time-measurements of astronomical observations made from the surface of Earth.

The Ecliptic Coordinate System (Celestial Latitude and Longitude)

The third one is the Ecliptic System. If a great circle is drawn through the pole of the ecliptic and the star, the angular distance of the star from the ecliptic measured along this great circle is called the celestial latitude of the star. The arc of the ecliptic intercepted between the First point of Aries and the foot of above great circle measures the celestial longitude of the star.



During its diurnal motion, the celestial latitude and longitude of a star remain unchanged in the same way as the right ascension n and declination. While the latitude is measured positive or negative depending upon whether the star lies to the north or south of the ecliptic, the longitude is measured eastwards from 0 to 360 degrees.

Chapter 2

Composition and Vertical Structure of the Atmosphere

The first session is a basic introduction to the composition and vertical structure of the Earth's atmosphere. It discusses the classification and origin of the many atmospheric components, including the growing concentrations of atmospheric pollutants and the threats they pose to us and our environment. This session also provides a first look at the vertical structure of the atmosphere and introduces basic

concepts such as temperature, pressure, and density. These concepts are the foundation for understanding the forces responsible for atmospheric motion and mixing which govern the transport, dispersion, and removal of atmospheric pollutants described in later sessions.

Part 1: Composition of the Atmosphere

- The Permanent Atmospheric Gases
- The Variable Gases
  - Water Vapor in the Atmosphere
- Air Quality in the Atmosphere
  - Pollutants
    - Nitrogen Oxides
    - Sulfur Dioxide
    - Volatile Organic Compounds
    - Ozone
    - Aerosols
- Greenhouse Gases

Part 2: Vertical Structure

- Change of Density, Pressure, and Temperature in the Vertical
- Sea Level
- Layers and their Characteristics
  - Troposphere
  - Stratosphere
  - Mesosphere
  - Thermosphere
  - Exosphere

Part 1: Composition and Vertical Structure of the Atmosphere

The composition of the Earth's atmosphere is gradually changing over time due to the natural processes that occur on Earth, as well as the many anthropogenic processes that have been introduced by humans. Natural biological processes such as the respiration of animals, including humans, removes

Picture of Mount St. Helens by US Geological Survey scientist Austin Post on May 18, 1980 oxygen from the atmosphere and in turn produces carbon dioxide (CO_2). The photosynthesis of plants uses the carbon dioxide to produce oxygen which is released back into the atmosphere. Other natural processes include geologic events such as volcanic eruptions. Not only do these violent eruptions eject massive amounts of particle debris into the atmosphere, but they also spew out large amounts of carbon dioxide and water vapor. While much of the atmosphere's composition is maintained by natural processes occurring on and within the Earth, how human activities interact with the atmosphere has profound implications for the quality and continuance of life on Earth. Although the percent volume of man-made substances is only a minute fraction of the total volume of the atmosphere, this amount is significant because it is this portion of the atmosphere that comprises most of what we term as pollutants.

The Permanent Atmospheric Gases

Although the composition of the atmosphere is always changing, an analysis of a "snapshot" of the atmosphere can provide us with a fairly good representation of the average concentrations of the "permanent" gases. "Permanent" here means the concentration is virtually constant near the earth's surface. Our snapshot reveals that nitrogen gas (N_2) makes up 78% of the concentration of the dry atmosphere (by volume). Oxygen gas (O_2), the second most abundant, makes up about 21% of the volume of the dry atmosphere. "Dry atmosphere" simply refers to a theoretical atmosphere that contains no water vapor. Together, nitrogen and oxygen make up about 99% of the air we breathe. Natural biological processes within the earth-atmosphere system maintain the amounts of these two gases to near constant proportions from the earth's surface up to about 50 miles (80 kilometers) above the earth.

Nitrogen and oxygen are the only gases that exist in the dry atmosphere that have concentrations above one percent near the earth's surface. The third most abundant gas in the dry atmosphere is argon (Ar). Argon is also considered a permanent gas. The "permanent" concentration of Argon is just less than one percent (about 0.93% to be more exact).

The Permanent Gases
(Near the Earth's Surface)

Gas Name	Chemical Composition	Percent (by volume of dry air)
Nitrogen	N2	78.08
Oxygen	O2	20.95
Argon	Ar	0.93
Neon	Ne	0.0018
Helium	He	0.0005
Hydrogen	H2	0.00005
Xenon	Xe	0.000009

The Variable Atmosphere

Unlike the few "permanent" gases, categorized as such because their proportions are nearly constant near the earth's surface, the concentrations of the numerous other substances found in the earth's atmosphere are variable. With the exception of water vapor, each one of these variable substances exists in the atmosphere in concentrations far less than one percent by volume. Because these constituents exist in such small amounts, their proportions are often recorded in parts per million (ppm) and parts per billion (ppb) by volume.

The figures in the table below are representative of the average concentrations of many of the variable substances found in our atmosphere. Note that in contrast with the permanent gases, the variable substances include gases as well as particulate matter. The air we breathe is not simply a composite of a few isolated molecules, but rather, a complex mixture of gaseous, liquid, and solid substances.

The Variable Substances
(Near the Earth's Surface)

Substance	Chemical Composition	Percent (by volume of dry air)	ppm (parts per million)
Water vapor	H_2O	0 to 4	-------
Carbon dioxide	CO_2	0.035	350
Methane	CH_4	0.00017	1.7
Nitrous oxide	N_2O	0.00003	0.3
Ozone	O_3	0.000004	0.04

Particulate matter	**-------**	**0.000001**	**0.01**
Chlorofluorocarbons (CFCs)	**-------**	**0.00000001**	**0.0001** **Water Vapor**

Water Vapor in the Atmosphere

Of the variable substances in the atmosphere, water vapor (H2O) is the most variable with concentrations ranging from 0-4% by volume. Most water vapor enters the atmosphere via evaporation and transpiration. Evaporation occurs when a single water molecule on a liquid water surface gains enough kinetic energy (often by solar radiation) to break the bond which holds the molecules together. If the energized molecule happens to be heading in the right direction, it will escape into the atmosphere as a single water vapor molecule. Transpiration is better explained in terms of vapor pressure. During the day, plant leaf pores called stomata open as a response to the sunlight. If the water vapor pressure inside the cell of the leaves exceeds the vapor pressure in the atmosphere, the water vapor molecules will travel from areas of high pressure (inside the leaf) to areas of lower pressure (outside the leaf). This process is referred to as transpiration. Evaporation and transpiration are collectively known as evapotranspiration.

Nearly all of the water vapor in the atmosphere resides in the lower portion of the atmosphere known as the troposphere. The cold temperatures at the top of the troposphere prevent all but a few water vapor molecules from escaping into the stratosphere where temperatures are warmer. Water vapor molecules that do manage their way into the higher regions of the atmosphere become disassociated by energy from the sun and can then participate in other chemical reactions. The troposphere actually conserves water on earth.

Water content in the atmosphere is often expressed as relative humidity (RH). The relative humidity is the ratio (usually expressed as a percent) of the actual amount of water in a sample of air to the amount of water in the same volume of saturated air at the same temperature. Relative humidity is only one of many expressions of moisture content. Others will be covered in depth in a later session.

Though the concentrations of all other variable substances might seem so minute as to be insignificant, many of these numbers are on the increase. For instance, air quality professionals have long been concerned with the amount of CFCs already present in the atmosphere because of the destruction of ozone in the stratosphere that occurs because of these chemicals. Stratospheric ozone protects us from the sun's ultraviolet rays which can cause skin cancer. Note in the table above that the concentration of CFCs is but a mere 0.0001 part per million. That tiny fraction is already at work doing damage to the atmosphere which in turn has long-term effects for us. In the next section, we will be looking more closely at some of the substances with which air quality professionals are particularly concerned.

Air Quality and the Atmosphere

Up to this point we have viewed the atmosphere in very general terms by naming and categorizing some of the various substances contained in the atmosphere. However, the atmosphere is not simply a large reservoir which merely stores these substances. Rather, the atmosphere is more like a large laboratory beaker filled with many different reactive chemicals. While it is not in the scope of this course to analyze in depth the many chemicals and reactions we find there, we will discuss a few of those substances that are central to the air quality issues we are currently facing in our communities, nation, and world.

Two terms to be aware of before we talk specifically about some of the substances in our atmosphere are aerosol and pollutant. Aerosol refers only to those liquid and solid particles (with the exception of water vapor and ice) that are suspended in the air such as pollen, dust, and smoke particles. The term pollutant, however, refers to any substance (solid, liquid, or gas) that contaminates the atmosphere and has the
potential to produce adverse health effects on humans and other animals, damage plant life, or cause damage to physical structures.

Some examples of pollutants include nitrogen oxides and sulfur dioxide.

As was stated in the first section of this session, the bulk of the atmosphere originated and is maintained by natural sources. Those naturally produced components are considered to be the pure atmosphere. While not all, many of the air quality issues we are currently faced with have arisen from increased concentrations of those gases and aerosols that have an anthropogenic source. Those by-products of human activities contaminate the pure atmosphere. By monitoring the concentrations of impurities in the atmosphere, knowing how they react with other gases and particles present, and modeling those emissions and reactions to predict future concentrations, legislation can be introduced to limit what we release into the air.

Pollutants

Nitrogen Oxides (NO_x)

The term NOx refers to nitric oxide (NO) and nitrogen dioxide (NO2). Nitrogen oxides have both natural and anthropogenic sources. The major anthropogenic source of nitric oxide is the high temperature combustion of fuel in automobile engines and power plants. Larger quantities of nitric oxide are released along with much smaller quantities of nitrogen dioxide as a result of a reaction between nitrogen and oxygen. This reaction is caused by the high temperatures. NO and NO_2 also occur naturally in the atmosphere as a result of bacterial action. Concentrations in urban areas range from between 10 -100 times the concentrations in non-urban areas. High concentrations of nitrogen oxides can result in respiratory problems, lowering the body's resistance to infections, as well as participating in the production of photochemical smog (Ahrens, 1991).

Sulfur Dioxide (SO_2)

Sulfur dioxide is another pollutant with both natural and anthropogenic origins. Sulfur dioxide is emitted into the atmosphere when coal, oil, and other sulfur-containing fuels are burned. Major sources of sulfur dioxide are petroleum refineries, power plants, paper mills, and smelters. Volcanic activity also releases around 109 kg of sulfur per year in the form of SO_2. Sulfur dioxide can oxidize to form sulfur trioxide (SO_3) and can react with moisture in the air to form sulfuric acid (H_2SO_4). Sulfuric acid in the atmosphere can also cause respiratory problems such as bronchitis and emphysema. Sulfuric acid is also deposited in the form of acid rain which can damage plant life, destroy man-made monuments and structures, and devalue personal property.

Volatile Organic Compounds (VOCs)

VOC is a general term for a class of organic compounds primarily made up of hydrocarbons. Hydrocarbons contain not only carbon, as all organic compounds do, but also, hydrogen. Some examples of VOCs are benzene, formaldehyde, chlorofluorocarbons (CFC's), and methane. Methane, as a pollutant, is becoming more important due to increased concentrations in the Southeastern United States. The major source of methane in that region is livestock. Methane is one of several greenhouse gases. We will discuss the effects of greenhouse gases later in this section. With the exception of methane, the primary sources of VOCs are industry, vehicle emissions, refrigerants, and cleansers. Many VOCs are not primary pollutants and are not harmful in and of themselves. However, many VOCs do react with other chemicals to produce secondary pollutants such as ozone.

Ozone (O_3)

Ozone can be found in both the stratosphere and lower troposphere. In the stratosphere, ozone occurs naturally. In the lower troposphere, however, ozone does not occur naturally and is a pollutant. We do not release ozone into the air directly, rather, in the lower atmosphere ozone is a product of various chemical reactions involving several of the pollutants that we do emit directly into the air such as nitrogen oxides and VOCs. Ozone is an example of a secondary pollutant.

In the lower atmosphere ozone is the primary component of photochemical smog. Smog greatly reduces visibility which creates a hazard for pilots and hinders the enjoyment of viewing some of the natural beauty in our national and state parks. Not only does smog reduce visibility, which is reason enough for concern but ozone, even in small quantities can cause health problems in people and animals as well as pose a threat to vegetation. Exposure to small concentrations of ozone can result in nausea, coughing, discomfort in breathing, and pulmonary congestion. It also retards growth in vegetation and even causes serious and extensive damage to crops (Ahrens, 1994).

Aerosols

As stated above, aerosols, also referred to as particulate matter, include any solid or liquid particles (with the exception of water and ice) suspended in the air such as metals, dust, smoke, and tiny droplets of sulfuric acid. One major concern in terms of air quality is what air quality professionals refer to as PM- 10. PM-10 stands for particulate matter with a diameter of 10 micrometers or less. Aerosols this small can easily travel deep into the lungs causing respiratory problems. A classification of aerosols even smaller than PM-10 is PM-2.5. Many aerosols that form from gases fit into this category. The immediate most obvious effect of increased concentrations of aerosols is reduced visibility. Because the term aerosol is so general, their sources are numerous and can range from automobile and industry emissions, to wind-blown dirt and dust stirred up from agricultural activities, to sea spray, as well as many others.

Part 2:
Vertical Structure of the Atmosphere

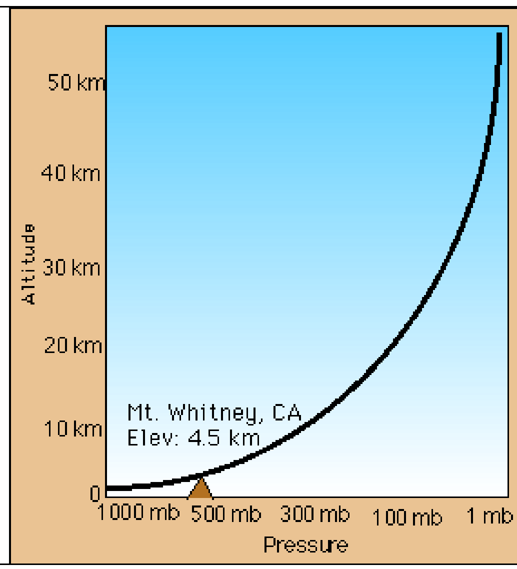
Imagine yourself in a balloon, traveling from the Earth's surface upwards through the atmosphere. As you rise, you will notice decreases in air density and air pressure. You may be surprised to discover that you will also feel both decreases and increases in air temperature. These changes are associated with distinct layers in the atmosphere, each with individual properties and characteristics. We will address each of these changes in turn, as well as the distinctions of each layer.

Density

Density is more or less a measure of concentration. It is expressed in terms of mass per unit volume. In a comparison of solids, liquids, and gases to one another, solids are the most dense while gases are the least dense. Solids have many more molecules (thus, more mass) per unit volume than either liquids or gases. Molecules in gases move very freely. In our atmosphere, density decreases rapidly with height (i.e., number of molecules which make up the air decreases with height). This is due to the Earth's gravitational pull. Molecules which are in the atmosphere are pulled towards the center of the earth. Therefore, there are higher concentrations of molecules near the surface of the earth than 10 miles (16 km) up. In fact, over 90 percent of all molecules in the atmosphere are within the first 10 miles (16 km). So why do all the molecules not form a layer of uniform concentration and density near the earth's surface? Well, there is a lot of mixing and vertical motion that keeps all those molecules moving around, which we will talk about in later lessons. The dramatic decrease in density as you go up affects the air pressure, causing it to decrease at a similar rate.

Pressure

The molecules in the atmosphere are in constant motion. While moving, they bump into each other and other objects on the order of 10 billion times a second. Each collision exerts a small force on the other object, creating a pressure. As



the concentration of molecules decreases, so do the number of collisions and therefore, the associated pressure. Because the density of the atmosphere decreases as you go up, air pressure decreases in the same proportion. There's another important part to the concept of pressure. As said above, molecules in the atmosphere are being pulled toward the Earth by gravity. This pulling is a force also, and so pressure is exerted at the surface of the earth. This force can be thought of as the weight of all the air above any point on the earth's surface. Air is pretty heavy -- 14.7 pounds per square inch (1013.25 millibars or 101.325 kilo-pascals) at sea level. (Meteorologists normally refer to pressure in units of either millibars (mb) or kilo-pascals (kPa). One millibar is equal to 100 Pascals (Pa) or 0.1 kilo-pascals (kPa). The Pascal is the international unit of pressure.) Ultimately, air pressure is a measure of all the air above any point, influenced by the molecules' forces of motion and gravity. It is also important to understand that the pressure at any given point changes with time because the air molecules do not stay in the same location. This changing pressure is key to our concept of weather. Generally, as pressure decreases, the weather becomes more stormy.

Temperature

Air temperature also changes with height. Since the number of molecules decreases with height, it is sometimes assumed that temperature also decreases with height. This, however, is not always the case. Each layer in the atmosphere has its own temperature profile. For our purposes, we will talk about temperature within the descriptions of each layer. But before we go into layers, we should talk about the importance of sea level.

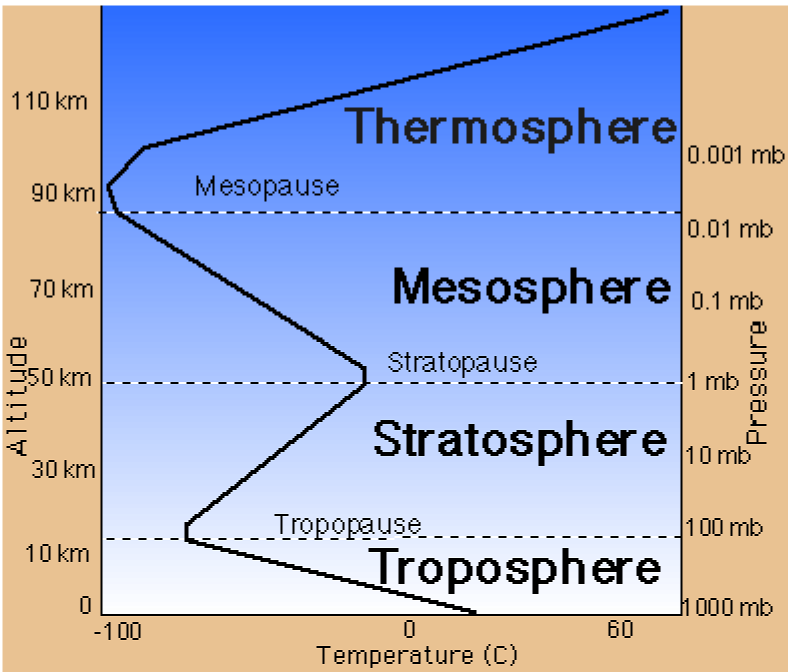
Sea Level

The earth's surface is rather, well . . . bumpy. In order for there to be some kind of standardization for meteorological data collection, all data is related to sea level, which is basically constant. Through various equations (which we'll discuss in later sessions), data, such as pressure, is converted to an equivalent sea level pressure. Therefore, it is important to know standard measures of parameters at sea level in order to make a comparison.

The standard air density at sea level is about 1.2 kilograms (kg) per cubic meter. The standard air pressure at sea level (known as sea level pressure) is 1013.25 mb (101.325 kPa). The standard temperature for sea level is 15 °C (59 °F). By using these values as standard for the atmosphere, we can compare all other measured values at various locations around the world and make a valued comparison. It means nothing to see on the news that the pressure outside is 940 mb (94.0 kPa) if you do not know the standards. If you do, you will realize that you are probably in the middle of a ferocious storm. However, if you saw a pressure of 1020 mb (102.0 kPa), you would know that you were probably having clear skies.

Layers in the Atmosphere

As we said earlier, each layer of the atmosphere has distinct characteristics. There are 5 main layers within the atmosphere, which we will discuss in turn. They are the troposphere, the stratosphere, the mesosphere, the thermosphere, and the exosphere.



Troposphere

The troposphere is the lowest layer of the atmosphere. This is the layer where most weather takes place. Most thunderstorms don't go much above the top of the troposphere (about 10 km) . In this layer, pressure and density rapidly decrease with height, and temperature generally decreases with height at a constant rate. The change of temperature with height is known as the lapse rate. The standard lapse rate for the troposphere is a decrease of about 6.5 degrees Celsius (C) per kilometer (km) (or about 12 degrees F).

Near the surface, the lapse rate changes dramatically from hour to hour on clear days and nights. Sometimes the temperature does not decrease with height, but increases. Such a situation is known as a temperature inversion. Persistent temperature inversion conditions, which represent a stable layer, can lead to air pollution episodes as we will discuss in Session 6's Focus on Air Quality.

The other main characteristic of the troposphere is that it is well-mixed. The name troposphere is derived from the Greek tropein, which means to turn or change. Air molecules can travel to the top of the troposphere (about 10 km up) and back down again in a just a few days. This mixing encourages changing weather.

The troposphere is bounded above by the tropopause, a boundary marked as the point where the temperature stops decreasing with height and becomes constant with height. Any layer where temperature is constant with height is called isothermal. The tropopause has an average height of about 10 km (it is higher in equatorial regions and lower in polar regions). This height corresponds to about 7 miles, or at approximately the 200 mb (20.0 kPa) pressure level. Above the troposphere is the stratosphere.

Stratosphere

The stratosphere is the layer above the troposphere, characterized primarily as a stable, stratified layer (hence, stratosphere) with a large temperature inversion throughout (see chart above). The main impact the stratosphere has on weather is that its stable air prevents large storms from extending much beyond the tropopause.

The other main impact important to life deals with ozone. Ozone is the triatomic form of oxygen that absorbs ultraviolet(UV) light and prevents it from reaching the earth's surface at dangerous levels. The stratosphere contains the ozone layer which has been such a hot topic as of late. The maximum concentrations of ozone are at about 25 km (15 miles) above the surface, or near the middle of the stratosphere. The interaction between UV light, ozone, and the atmosphere at that level releases heat, warming the atmosphere and helping to create the temperature inversion in this layer.

The stratosphere is bounded above by the stratopause, where the atmosphere again becomes isothermal. The average height of the stratopause is about 50 km, or 31 miles. This is about the 1 mb (0.1 kPa) pressure level. The layer above the stratosphere is the mesosphere.

Mesosphere

The mesosphere is the middle layer in the atmosphere (hence, mesosphere). There are two key points about the mesosphere. First, temperature in the mesosphere decreases with height. At the top of the mesosphere, air temperature reaches its coldest value, around -90 degrees Celsius (or -130 degrees Fahrenheit). The second point is that the air is extremely thin at this level. Over 99.9 percent of the atmosphere's mass lies below the mesosphere. However, the proportion of nitrogen and oxygen at these levels is about the same as at sea level.

The mesosphere is bounded above by the mesopause. The average height of the mesopause is about 85 km (53 miles), where the atmosphere again becomes isothermal. This is around the 0.005 mb (0.0005 kPa) pressure level. Above the mesosphere is the thermosphere.

Thermosphere

The thermosphere is a warm layer above the mesosphere. In this layer, there is a significant temperature inversion. The few molecules that are present in the thermosphere receive extraordinary amounts of energy from the sun, causing the layer to warm. Though the measured temperature is very hot, if you exposed your skin to the thermosphere, the perceived temperature would be very cold. Because there are so few molecules present, there would not be enough molecules bombarding your body to transfer heat to your skin. Temperature is a measurement of the mean kinetic energy, or average speed of motion, of a molecule. So, although there are only a few molecules, each has a huge amount of kinetic energy.

Above the thermosphere is the exosphere. Unlike the layers discussed previously, there is no well-defined boundary between the thermosphere and the exosphere (i.e., there is no boundary layer called the thermopause).

Exosphere

The exosphere is the region where molecules from the atmosphere can overcome the pull of gravity and escape into outer space. The atmosphere slowly diffuses into the void of space. The exosphere usually begins about 500 km up (notice, this is well off the chart above), but there is no definable boundary to mark as the end of the thermosphere and the beginning of the exosphere. Even at heights of 800 km, the atmosphere is still measurable. However, molecule concentrations are very small and considered negligible.

Chapter 3

Measures of the Atmosphere

This session will describe the few measurable parameters of the atmosphere such as temperature, humidity, and pressure and the many methods that are used to measure them. In addition, we will discuss how those few measures are used to determine other parameters like the lifting condensation level (LCL) and potential temperature, which are used in evaluating atmospheric conditions and predicting the dynamics that drive the transport and dispersion of pollutants.

Part 1: Why We Measure the Atmosphere
Part 2: What We Measure (A snapshot of the Atmosphere)

- Temperature
- Moisture
- Pressure
- Wind

Part 3: Values We Infer

- Equation of State
- Virtual Temperature (Tv)
- Potential Temperature ( )
- Lifting Condensation Level (LCL)
- Equivalent Potential Temperature (e)
- Reduction to Sea Level (Hypsometric Equation)

Part 1:
Why We Measure the Atmosphere

Why do we measure the atmosphere? The obvious answer is to predict the weather. We regularly record variables such as temperature and pressure because we want to and need to know what the weather will be. We must know the current state or condition of the atmosphere as well as changes in atmospheric conditions over a period of time in order to forecast weather events. We need to know of major weather systems that are elsewhere and heading our way. We need to know when a severe storm may develop near us or "pop up out of nowhere." We need to know the strength and duration of various weather systems so that we may adequately prepare in advance. If a drought is forecasted, we may need to plan ahead for crops, livestock, and city water supplies. Evacuations may need to be executed in preparation for severe storms that could produce damaging winds or flooding. Our livelihood often depends on accurate weather forecasts.

A less obvious answer to the question above is to forecast air quality. Actually, forecasting air quality is still only part of the answer. You see, we cannot control the weather. There are ways that we can modify some aspects of weather events. We can increase and decrease to some degree the amount of rain a storm may produce, as well as the size of hailstones produced. But these modifications are usually not cost effective and they require adding chemicals that may later be detrimental. In contrast, we can often significantly control the quality of the air by limiting what we emit into the atmosphere and when we emit it. When pollutants are released into the atmosphere, as we discussed in Session 1, they become part of the composition of the atmosphere. When the atmosphere moves, they move. When the wind blows, they are a part of the wind. Thus, wind patterns can tell us where pollutants will travel.

Atmospheric stability, determined from the measurements we take, can also help forecast pollutant concentrations at given locations. A stable atmosphere will trap pollutants, often at the surface, while an unstable atmosphere will enable them to mix and dilute with cleaner air. Storms can cleanse the atmosphere of many pollutants. Interestingly, all precipitation begins as an aerosol. But wait, in

discussion of aerosols in Session 1 we said that liquid and frozen water were not considered aerosols. Water and ice are not aerosols. Rain, snow, sleet, and hail all form when water vapor either condenses or freezes onto a solid particle, often a pollutant. Those particles, whether pollutants or not, are called cloud condensation nuclei or CCN. When precipitation falls to earth, those particles are removed from the air. Wind speeds can influence pollutant deposition as well. Heavy particles will fall out of the air if winds are light while stronger winds will keep them aloft and in motion.

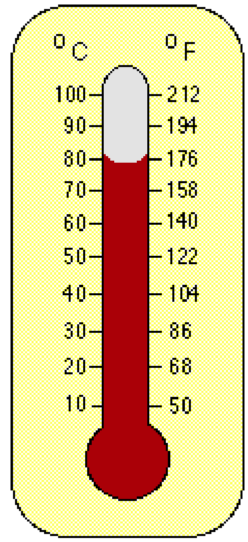
From this short discussion we begin to see why it is necessary to record regular atmospheric measurements. The more we understand about the atmosphere, the better we are able to predict not only the weather, but predict and control air quality as well. In this session, we will briefly discuss the varying properties of the atmosphere we can measure and introduce some of the more discreet parameters we can infer from them. We will also discuss where and how in the atmosphere the data is obtained. As in the case of many of the sciences, the study of meteorology requires learning many small pieces of information in order to build up to a complex process. It is often difficult to convey the significance of each piece until the process itself is better understood. Some of the material that follows is analogous to looking at individual pieces of a puzzle. You may not be able to grasp the importance of each piece until you see the whole puzzle completed. But keep in mind, each piece is important.

Part 2:
What We Measure
(A Snapshot of the Atmosphere)

You may be surprised to find out there only a few parameters that are regularly measured in the atmosphere. The most common include temperature, humidity, pressure, and the speed and direction of the wind. These few pieces of information taken at any point in the atmosphere basically provide us with a snapshot of the atmosphere at that point. From these snapshots taken simultaneously at predetermined locations in the country and at various heights in the atmosphere, we can infer nearly every other parameter of the atmosphere we need in order to model the motion of the atmosphere. So what's the catch? The catch is integrating all of the many factors that cause these snapshots to change over time, such as the earth-atmospheric heat budgets, discussed in Session 2, and the relationships between the measured parameters themselves. Modeling atmospheric motion is essential for both weather forecasting and predicting the transport and dispersion of pollutants. In the remainder of Part 2 and Part 3, we will look at those parameters we can measure and a few of the important relationships between them that we need to understand in order describe how the motion in the atmosphere is initiated.

Temperature

When we hear the word "temperature" we normally think of hot and cold. But what is hot and cold? When we measure the temperature of something we are actually measuring the average kinetic energy of molecules. Therefore, when we measure the air temperature, we are measuring the average kinetic energy of the molecules that make up the composition of the air. When air molecules bombard a thermometer, the kinetic energy of the air molecules is transferred to the liquid in the thermometer. This transfer of energy causes the liquid to heat up and expand and the thermometer to "rise." In the upper layer of the atmosphere, the temperature is very hot, at 115 km it can be as hot as 65 C, but you would freeze because there are so few molecules that high up to bombard your body.



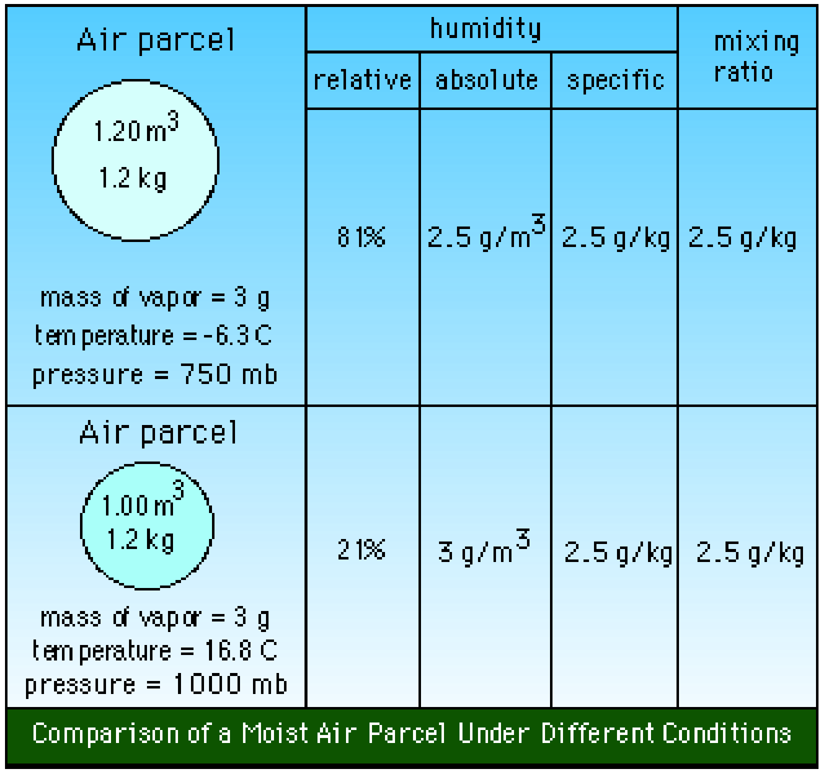
In meteorology we usually record temperatures in degrees centigrade or Celsius (C), but for computations we convert degrees Celsius to degrees Kelvin (K). Water freezes at 0 C and boils at 100

C. The centigrade temperature scale was, in fact, derived from these properties of water to create a scale with convenient reference points. The Kelvin scale is referred to as the absolute temperature scale. Zero degrees Kelvin (0 K) is absolute zero. If a substance has a temperature of 0 K then this means that all molecular motion has ceased. The molecules have no motion, thus no kinetic energy, and absolutely no temperature. Zero degrees Kelvin converts to centigrade as -273 C. In fact, this easy conversion is another reason we use the Celsius and the Kelvin scales. All you have to do to convert units of Celsius to Kelvin is add 273 degress to the centigrade temperature.

As in the case of the thermometer, increased molecular motion of the air molecules results in the expansion of the air. In other words, as the speed of the molecules increases, the molecules spread further away from each other making the air less dense. As mentioned earlier in this course, the "normal" density of air is 1.2 kg/m3. Less dense or lighter materials "float" on denser materials like a cork floats on the water. This is the reason warm air rises. Convection in the atmosphere is caused by heating of the air nearer the heat source, usually the ground, pavement, or a body of water. The heat is transferred to the air, causing it to expand and rise. Convection is one way in which pollutants are transported away from the surface of the earth and diluted with cleaner air.

Moisture

The term most frequently used to express the amount of moisture in the air is relative humidity (RH). The relative humidity is the ratio of the actual amount of water vapor in a sample of air compared to the total amount



of water vapor the same sample can hold before condensation begins (i.e., it becomes saturated with water vapor) at a given temperature and pressure. As the term suggests, RH only gives us a relative sense of the amount of moisture in the air, not the actual amount. On the east coast, for example, the temperature may be 95 degrees Fahrenheit (35 degrees Celsius) with a relative humidity of 90 percent, while in the western desert, the temperature may be similar with a relative humidity of only 20-30 percent. Clearly, there is more water vapor in the air on the east coast, but in either case, the relative humidity does not tell us how much is in the air. Absolute humidity and specific humidity are measures of the actual amount of water vapor in the air. The absolute humidity is the mass of the water vapor per unit volume of air. Does this sound familiar? The absolute humidity is actually the density of the water vapor in the air, while the specific humidity is a ratio of the mass of the water vapor in a sample of air to the total mass of the sample. Similar to the specific humidity is the mixing ratio (w). The mixing ratio is the mass of the water vapor in a sample of air divided by the mass of the dry portion of the air (rather than the total mass of the sample as in the case of specific humidity). The mixing ratio is often recorded in grams of water vapor per kilogram of dry air (g/kg). The mixing ratio of saturated air is called the saturation mixing ratio (ws).

$$\text{Relative Humidity} = \frac{\text{mass of water vapor}}{\text{mass of water vapor at saturation}} \times 100\%$$

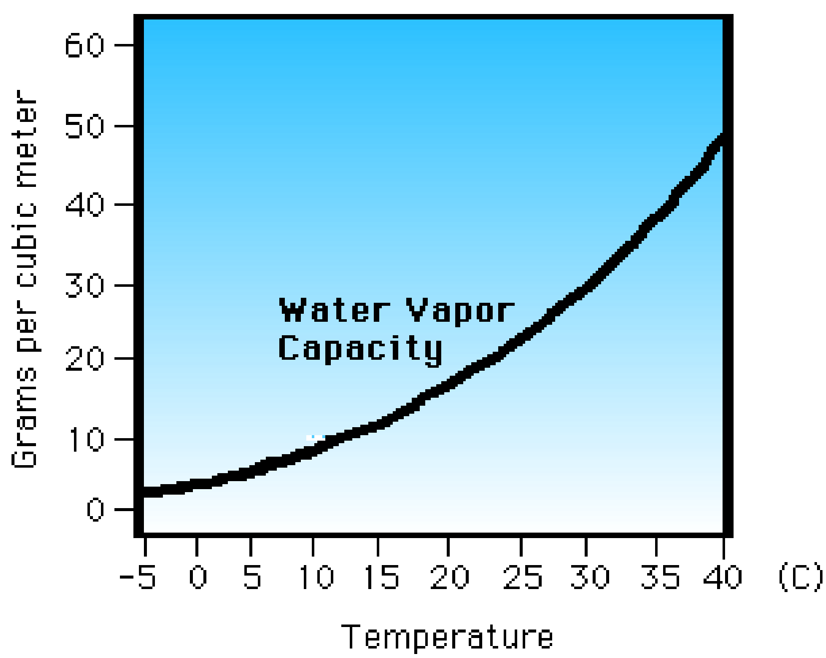
$$\text{Absolute Humidity} = \frac{\text{mass of water vapor}}{\text{volume of air}}$$

$$\text{Specific Humidity} = \frac{\text{mass of water vapor}}{\text{total mass of sample}}$$

$$\text{Mixing Ratio} = \frac{\text{mass of water vapor}}{\text{mass of dry air}}$$

An interesting thing about humidity is the manner in which it is measured. It would be difficult and time consuming, if not impossible, for meteorologists to measure out a volume of air, separate the water vapor from the dry air and weigh them separately. Fortunately, there are quicker and easier methods. The dew point temperature (Td) is a direct measurement that is taken to determine the amount of moisture in the air. The dew point temperature is the temperature to which a sample of air must be cooled for condensation to occur or to make the sample saturated. This measurement is easy to make. You simply take a sample of air and cool it until you first begin to see condensation. Though the dew point temperature does not tell us directly how moist the air is, as we shall see a little later, we can use the dew point temperature to find out how much water vapor is in the air.

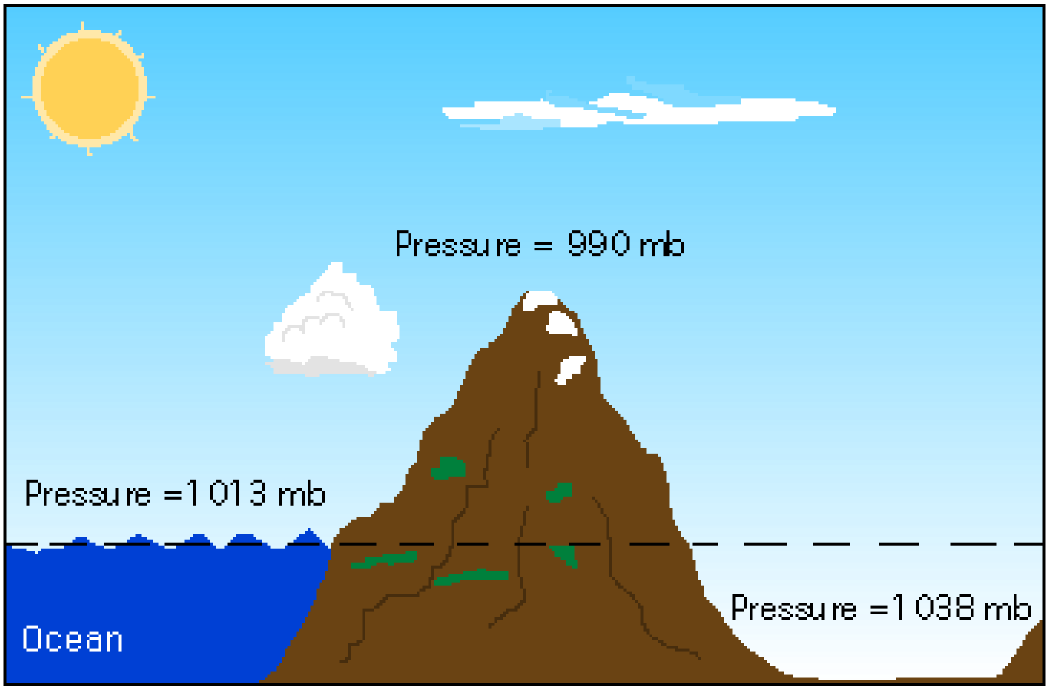
Another measurement of temperature that is related to humidity is the wet-bulb temperature (Tw). Instead of measuring the amount of water vapor already present in the air, the wet-bulb measures the amount of water vapor that is not present in the air. More specifically, it measures the amount of energy necessary to evaporate water into the air. The wet-bulb accomplishes this by measuring the temperature drop over a surface of water as a result of evaporating water into the air. A wet wick is placed over the end of a thermometer, and the thermometer is whirled around in the air. The dryer the air, the more evaporation that will take place. Evaporation requires energy. When a water molecule evaporates and breaks the molecular bond that holds it to the water surface, the body of water loses that energy along with the molecule. This results in the cooling of the body of water. The wet-bulb thermometer measures the drop in the temperature. The wet-bulb depression is the amount the temperature drops on the wet bulb thermometer as a result of evaporation. In saturated air, the air temperature, dew point temperature, and wet-bulb temperature are equal. In unsaturated air, dew point is always the lowest and the air temperature is the highest. The wet-bulb temperature always occurs between the other two.

Much of what has been discussed regarding moisture and humidity suggests that the amount of moisture the air can hold is average speed of the molecules in warmer air is greater than in cooler air.	

variable. Generally, the warmer the air is, the greater the potential the air has for holding water. The reason for this is that the Warmer air is less dense and the molecules are spread out. When air cools, the average movement of all the molecules is slowed, including the water vapor molecules. Water vapor molecules have a strong molecular attraction to each other. Therefore, as the temperature cools and molecular motion slows, the natural molecular attraction between the water vapor molecules becomes greater than the kinetic energy they possess, causing the water vapor to condense. Condensation is obviously significant in that it eventually produces precipitation, but that is not all. When the formation of a droplet begins, the nucleus of each droplet is a solid particle. In many cases that solid particle is a smoke particle or some other pollutant, which may eventually fall out of the cloud in rain droplets or some other precipitation. Condensation also releases latent heat. This heats the surrounding air, causing it to be less stable. If the air is rising, the release of latent heat may prolong its ascent because the additional heat energy keeps it warm and rising, extending the process of convection.

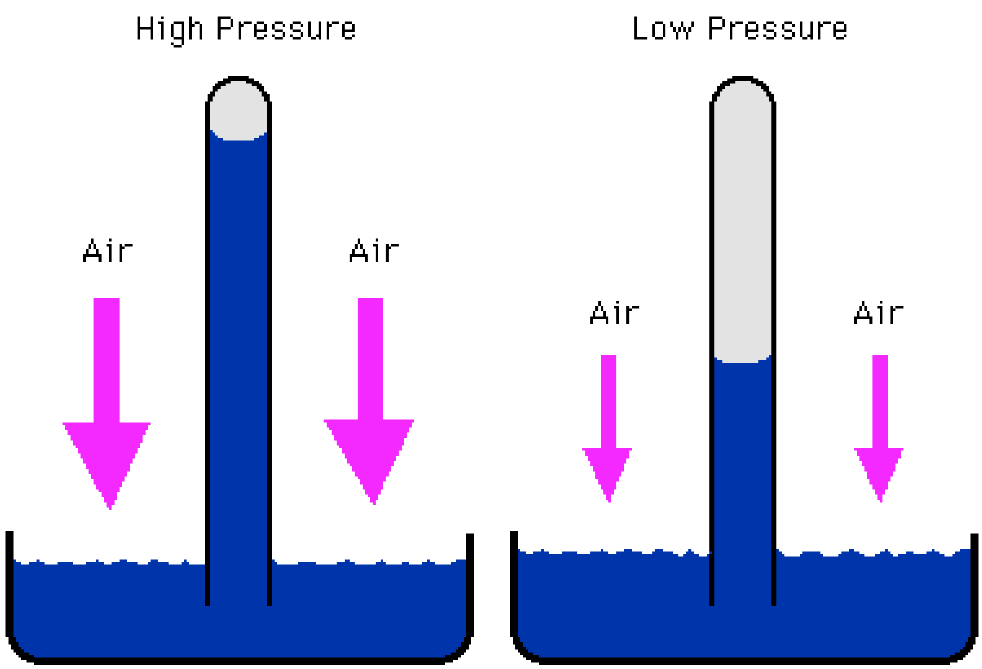
Pressure

The concept of atmospheric pressure is often a hard one to grasp. It is difficult to think of the atmosphere as having mass because we generally do not notice its weight pressing on our body. As discussed in Session 1, pressure is not only due to the mass of the molecules, but also the motion of the molecules bombarding our body. The atmosphere actually exerts an average force of 14.7 pounds per square inch of area at sea level. Pressure (P) is defined as this force (F) per unit of area (A), i.e, P=F/A. In units that meteorologists are more accustomed to and for computational purposes, the average pressure at sea level is 1013.25 millibars (mb) or 101.325 kiloPascals (kPa) (also equivalent to 29.92 in/Hg and 760 mm/Hg where Hg is the symbol for mercury). Sea level is used as a reference point for height. Areas with elevations below sea level can expect higher average pressures because the air is more dense, while higher elevations can expect lower average pressures. Pressure is proportional to density ( ). Since more molecules are exerting forces in dense air, the pressure is greater. The pressure profile of the atmosphere from Session 1 illustrates the decrease in pressure with height.



Pressure at the surface is measured with a barometer. While there are many types of barometers, the most commonly used barometers for meteorological purposes are the mercury barometers. A properly calibrated mercury barometer is extremely accurate. These barometers consist of a glass tube filled with liquid mercury and closed on one end. The tube stands on end with the closed end up and the open end

submerged in a reservoir of mercury that is exposed to the air. As the air pressure rises, it pushes on the liquid in the reservoir. The level of the mercury rises in the glass tube to compensate for the additional pressure exerted on the exposed reservoir. The average air pressure at sea level as recorded on a mercury barometer is 29.92 inches of mercury (in. Hg) That is, the column of mercury is 29.92 inches high at standard sea level pressure. Though we speak in terms of the average or standard pressure at sea level, the pressure at any place on the surface of the earth is variable and continuously changes with time due to the heating and cooling of the earth. Varying pressures result in a pressure gradient force. This force – the wind - is only one of several that produce motion in the atmosphere, transporting pollutants.

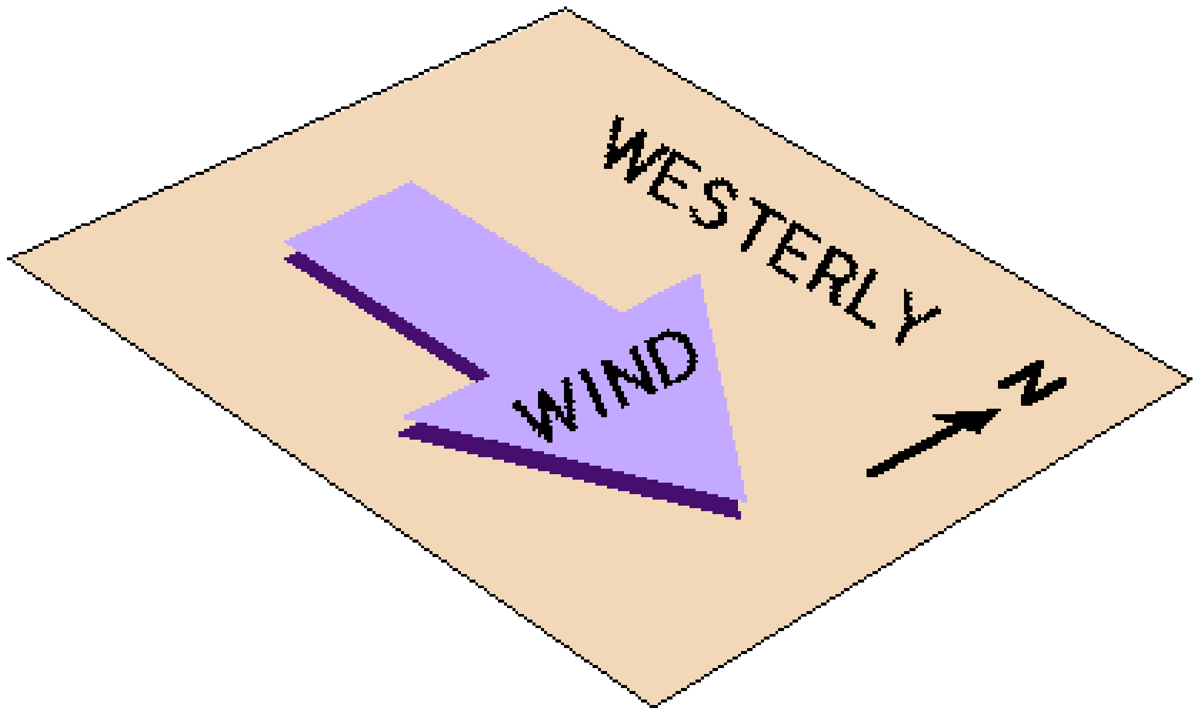


Wind

The heating of the earth, as we have seen in Session 2, is uneven. As parts of the earth are heating, causing the air near the surface to warm, expand, and rise, other areas are cooling. The region vacated by warm rising air becomes an empty pocket in the atmosphere. This empty space is an area of low pressure or low density because there are few molecules left to occupy the space. Air, like all other fluids,

naturally moves from areas of high pressure to low pressure, thus the air near the empty pocket rushes in to fill the void. This movement of air from high pressure to low pressure is what we call the wind.

Two important measurements of the wind are the direction and speed of the wind. Wind speed is generally recorded in knots (kn) or nautical miles per hour where one knot is equal to 1.15 mi/hr. For making computations, however, knots must be converted to meters per second (m/s). One knot is equal to 0.51 m/s. Wind directions are not given in reference to the direction in which they are blowing, but rather the direction from which they are blowing. A westerly wind blows from west to east. A northerly wind blows from north to south. The simplest way to measure the wind direction and speed at the earth's surface is with wind cups and vanes. The vane gives the direction while the cup catches the wind and rotates giving an indication of speed. Wind speeds and directions at the surface and aloft help us to predict where and how fast weather systems will move and pollutants will be transported. We can also infer from the winds whether pollutants will be mixed or deposited. Light winds cannot keep heavy particles aloft while even the slightest breeze may keep lighter particles suspended indefinitely.



In this brief overview of temperature, moisture, pressure, and wind, we have seen how each of these parameters play an important role in predicting weather phenomenon as well as the transport, dispersion, and deposition of pollutants. We have also seen how these properties of the atmosphere can influence each other. For example, heating at the surface causes convection or vertical motion by altering the temperature and pressure of the air at the surface. This rising air produces horizontal motion (wind) as air rushes in to fill the void left by the rising air. As the rising air cools, the condensation of water vapor releases heat which, in turn, increases the instability of the surrounding air, and so on.

While the parameters we have discussed so far can be directly measured, there are many others that we can calculate from them which provide us with even greater detail of the atmosphere. These will be the topic of our next section.

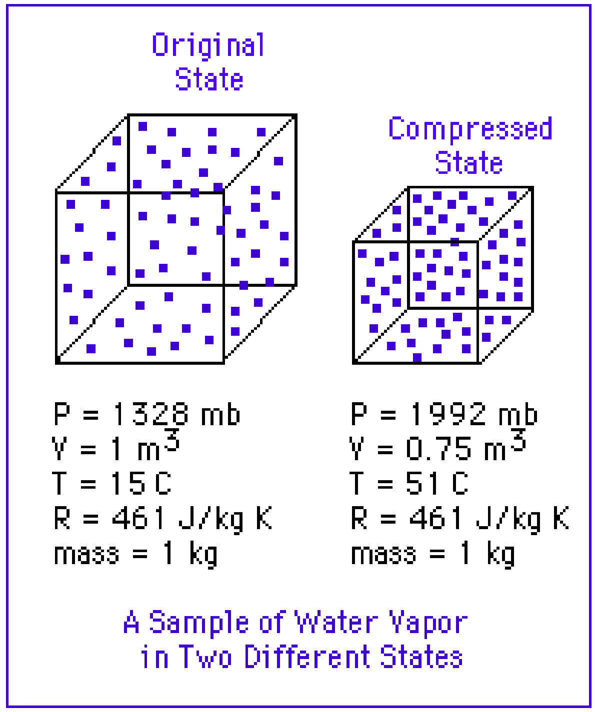
Part 3: Values We Infer

While temperature, humidity, pressure, and wind data tell us much about current atmospheric conditions, in and of themselves, they tell us little about future conditions. We can use them, however, to determine the values of other parameters which are necessary for forecasting and modeling the atmosphere. Computers now calculate these values for us so you will not be expected to make these computations

yourself during this course. However, a look at how a few of these parameters are defined either mathematically or graphically can increase your understanding of the behavioral influences of the atmosphere.

Equation of State

If we want to model the behavior of the atmosphere, which is mostly made up of gases, we must study the natural behavior of gases. All gases are found to approximate the equation of state. The equation of state does just what its name suggests.



A Sample of Water Vapor in Two Different States

It tells us of the state of a particular gas under a broad range of conditions. The equation of state defines the relationship between the temperature (T), density (ρ), and pressure (P) of the gas. These we call the state variables. The equation of state can be written as

$P = \rho R T$

Pressure = Density x Gas Constant x Temperature

where P is measured in pascals (Pa), ρ is in grams per cubic cm.

Virtual Temperature (Tv)

In the last section, we defined the equation of state for a dry atmosphere. In the real atmosphere, the water content is highly variable, from 0-4% by volume. The proportionality constant (R) for dry air is 287 J kg- 1K-1, while R for pure water vapor is 461 J kg-1K-1. But, the atmosphere is rarely ever dry and is never pure water vapor. Thus, the value for R in the real atmosphere is always somewhere between 287 and 461. As the moisture content of a sample of air changes, so does the proportionality constant for that sample. The value for R, though not difficult to calculate, is not a joy to calculate either. For moist air, using the equation of state could become a bit cumbersome since you would need to compute a new R value each time the moisture content changed. Necessity is the mother of invention and for this reason, virtual temperature (Tv) was invented. We have stated many times that less dense or lighter air rises, and we have generally referred to the less dense air as warmer air. To make matters more interesting, we will now inform you that a sample of moist air is less dense than a sample of dry air at the same volume, temperature, and pressure. The reason for this is water vapor has a smaller apparent molecular weight than dry air. In other words, one mole (6.02 x 1023 molecules) of water vapor weighs less than one mole of dry air. The definition of virtual temperature (Tv) is the temperature to which a sample of dry air must be heated in order to have the same density as a sample of moist air at the same pressure. The virtual temperature can be calculated using the equation

$$T_v = (1+0.61w)T$$

Virtual Temperature = (1 + (0.61 x Mixing Ratio)) x Temperature

where w is the mixing ratio in units of kg water vapor/kg dry air and T is in degrees Kelvin. (NOTE: the mixing ratio is usually recorded in g/kg and has to be converted to kg/kg when used in computations.) The mixing ratio can easily be obtained from meteorological charts using only the dew point temperature and pressure. Calculating Tv is much simpler than calculating R. The virtual temperature allows us to use the equation of state for dry air by substituting Tv for T. The equation of state for moist air then becomes

$$P \alpha = R_d T_v$$

Pressure x Specific Volume = Dry Gas Constant x Virtual Temperature

where Rd is the proportionality constant for dry air. The virtual temperature simplifies the complications that the presence of water vapor in the atmosphere creates. It allows us to convert a moist sample of air into a theoretical dry one. Virtual temperature is a term that is used extensively in meteorology. We generally always use Tv instead of T when speaking of a moist atmospheric environment.

Potential Temperature (θ)

Let's suppose you are working outside on a hot summer day. When you go inside to cool off and check out the score of the big game you are missing, a weather alert comes on calling for severe thunderstorms with large damaging hail. How could such large balls of ice fall from up there when it is so hot down here? And if it is cold enough up there to form ice, why can't we just bring some of that cold air down to cool things off a bit?

The equation of state can help us answer these questions. The equation of state helps us figure out the state of the atmosphere under various conditions. For instance, suppose the air at the surface of the earth is dry and in equilibrium, stable, and has a pressure of 1000 mb (100 kPa), a specific volume of 0.83 m3/kg (ρ =1.21 kg/m3), and a temperature of 25 C. If we were to lift a volume of that air (without adding any heat to the air) to a height where the pressure was 700 mb (70 kPa), the volume of the parcel would increase and the temperature would decrease in order for the lifted air to assume a pressure of 700 mb like the surrounding air. To say it another way, the parcel would expand to a pressure of 700 mb. Because the parcel was lifted without adding heat, the change of state of the parcel is called a dry adiabatic change. Adiabatic simply means to alter the state of a gas without adding or withdrawing heat from the gas. The change in the temperature with the change in pressure is called the dry adiabatic lapse rate. Through several computations using the laws of thermodynamics, the specific volume cancels and a new equation is used to describe the relationship between the temperature and pressure of the atmosphere during adiabatic expansion and compression. This equation is referred to as Poisson's equation and is written as

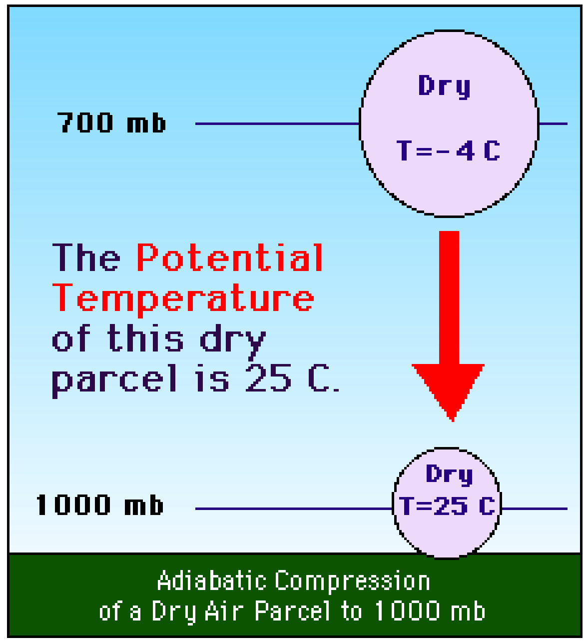
$$\frac{T_0}{T} = \left(\frac{p_0}{p}\right)^{0.286}$$

where T0 is the temperature at the higher pressure in degrees Kelvin (K), T is the temperature at the lower pressure in degrees Kelvin, p0 is the higher pressure, and p is the lower pressure. For the parcel of air described above being lifted to a height where the pressure is 700 mb (70 kPa), the temperature would decrease from 25 C to -4 C. That's cold!

To answer our initial questions about hail on a hot day and cold air aloft, the air above is so much colder than the air at the surface because the temperature is in equilibrium with the density and pressure at the height it is residing. So, hail can form on a very hot day because the temperature aloft can be below freezing when the temperature at the surface is quite scorching. But if we were to try to bring that same cold air to the surface, the temperature would increase in order to compensate for the adiabatic compression.

For convenience of comparison, meteorologists have defined the potential temperature(θ) as the temperature an air parcel would have if it were expanded or compressed adiabatically to a pressure of 1000 mb (100 kPa). Thus, for the air parcel we lifted, because we started at 1000 mb (100 kPa) with a temperature of 25 C, the potential temperature (θ) is 25 C. To calculate the potential temperature, Poisson's equation can be rewritten as

$$\Theta = T\left(\frac{1000}{p}\right)^{0.286}$$

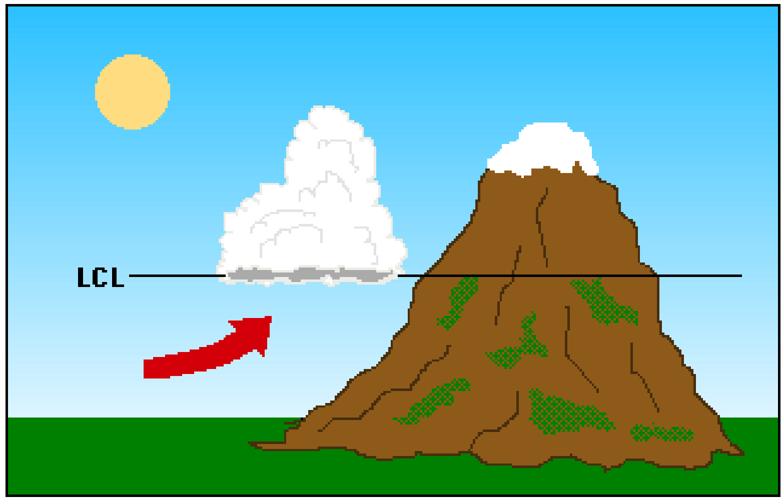


Adiabatic Compression of a Dry Air Parcel to 1000 mb

where T is the initial temperature in degrees K and p is the initial pressure in millibars. In addition to calculating potential temperature, adiabatic charts (thermodynamic diagrams) have been constructed which enable us to easily look up the temperature and pressure of a parcel of air and determine its potential temperature.

Potential temperature is going to be an important term later when we discuss the vertical stability of the atmosphere. Comparing the potential temperature of a parcel to that of its surroundings can tell us weather a region of air is stable or subject to lifting or sinking. We will see that the issue of stability is a major factor when trying to determine the transport and dispersion of air pollutants. Pollutants trapped in a stable layer can sit stagnant in very high concentrations. Once air becomes unstable and buoyant the air may rise and mix and dilute the pollutants.

Lifting Condensation Level (LCL)



Now, consider that same parcel of air we just described above with just one deviation. Instead of dry air, the air now contains moisture.

The amount of moisture in the air is expressed by the dew point temperature of the air. If we begin to lift (which is equivalent to expanding) that moist parcel of air adiabatically, it will cool as it expands. As the parcel expands, the dew point of the parcel also changes in response to the pressure change. When the temperature of the air parcel and the dew point temperature are equal, condensation will occur, and a cloud will form. The dew point temperature represents the temperature at which an air parcel will be saturated with water vapor and condensation will occur. The height at which a parcel is lifted adiabatically and cools to the dew point temperature is known as the **Lifting Condensation Level** or the **LCL**. The LCL can also be obtained very easily from an adiabatic chart using the temperature, dew point temperature, and pressure. In the atmosphere, air is lifted and cools adiabatically when moist air is lifted up over a mountain or when moist air lifts up over dry air because the moist air is less dense, as in the case of an advancing cold front.

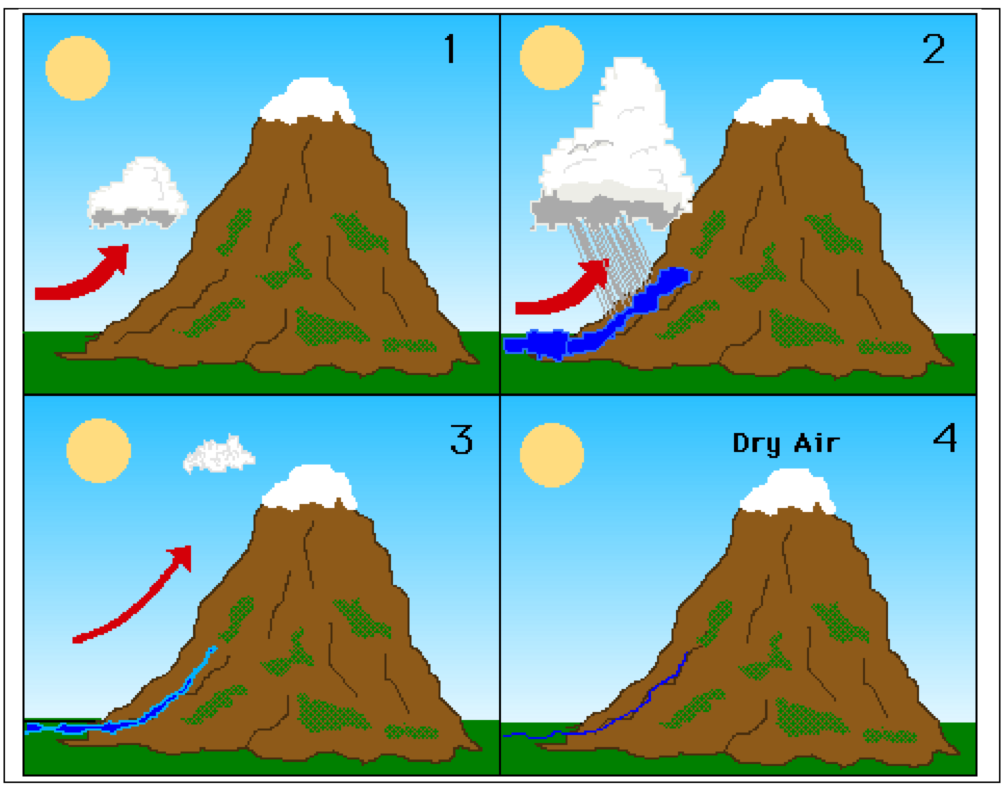
In the summer, the sky is often dominated by puffy cumulus clouds. While cumulus clouds often form as a result of adiabatic lifting, as in the case of air moving up a mountain side, they can also form from the convection of warm air heated at the surface of the earth. The term "lifting condensation level" is used in reference to this type of lifting as well. Once a parcel leaves the heat source, which is the ground, heat is no longer added, and the process is considered adiabatic. So, when you look up in the sky, the base of the lower forming cumulous clouds is considered the LCL.

The LCL will help us to evaluate atmospheric stability in Session 6. In fact, most of these parameters we are discussing are pieces of the process of determining atmospheric stability. In terms of developing storm systems, stability is the issue in which we are most concerned because it helps us predict the intensity of a storm . Atmospheric stability is also important in the study of air quality. The stability of the atmosphere influences the transport and dispersion and, ultimately, the concentrations of pollutants in the atmosphere.

Equivalent Potential Temperature (θe)

Once rising air reaches the LCL or, condensation level, and clouds begin to form, we can no longer refer to the lifting as a dry adiabatic process. When a parcel of air becomes saturated and condensation begins, the process of condensation releases latent heat into the surrounding air. This latent heat further warms the air making the air even more buoyant. We refer to this as a moist adiabatic or saturated adiabatic process. The moist adiabatic expansion increases the instability of the parcel. The moist adiabatic lapse rate is not as steep as the dry adiabatic lapse rate meaning a parcel of air that has risen above the LCL will not cool as rapidly as it did before reaching the LCL. If this process of moist adiabatic expansion continues, all of the water may condense out of the rising parcel and precipitate out, yielding a dry parcel. The potential temperature of that new dry parcel is called the equivalent potential temperature (θe) of the original moist parcel. So, the equivalent potential temperature is the potential temperature of a parcel of air after all of the water vapor has condensed and fallen out of the parcel.

Let's continue with our example of the air lifting up the side of the mountain. If the mountain is high enough, and the parcel or layer of air continues to lift, all of the moisture in the lifting air could condense and fall out of the air in the form of precipitation. Once this occurs, the potential temperature of the now dry air is considered the equivalent potential temperature of the original moist parcel before condensation began. Like the potential temperature and the LCL, the equivalent potential temperature will also be a key player when we discuss vertical stability. The equivalent potential temperature will help us determine the stability of a thick layer of air versus the stability of a single isolated parcel. Once again, this is one piece in the puzzle that will be necessary later when we try to attach all of the pieces together.



Reduction of Pressure to Sea Level

In our discussion on sea level in Session 1, we discussed the importance of the standard pressure at sea level. Because the surface of the earth is not flat and pressure decreases with height, it is hard to compare surface pressures taken at locations of different elevations. In order to standardize surface pressures we **reduce** them to sea level, meaning we calculate what they would be if the elevation at which they were taken were the same as the sea level. We use the term "reduce" whether the elevation of original pressure is obtained at a level above sea level or below.

Consider the surface pressure at the top of a mountain compared with the surface pressure at the base the mountain. We would expect the pressure at the base to be higher. Now suppose a low-pressure system was moving in and the two pressures were equal. It is difficult to evaluate the low-pressure system unless we have a standard point of reference. We can create a reference by reducing both pressures to sea level. Once this has been done, we can compare the reduced pressures and determine the strength of the system. When we watch weather broadcasts on TV, the meteorologist reports the local surface pressure. That pressure is a real surface pressure, but the pressures that appear on the surface maps showing high and low-pressure systems have been reduced so that pressure trends are easily discernible. The hypsometric equation enables us to easily reduce surface pressures. This equation can be written as

$$p = p_o \exp\left(\frac{gZ_g}{Rd\,\overline{Tv}}\right)$$

where p is the pressure reduced to sea level, po is the surface pressure, g is the acceleration due to gravity, Zg is the elevation of the ground, Rd is the gas constant of dry air, and Tv is an estimated average virtual temperature of the theoretical layer of air between the ground and sea level.

By reducing surface pressures to sea level we can evaluate pressure changes over large scales such as across the entire country or even globally. This enables us to recognize pressure trends that will indicate horizontal and vertical motion in the atmosphere. Session 4 will discuss how these pressure changes actually influence motion in the atmosphere.

Chapter 4

Atmospheric Stability

"Up, up, and away." Sustained vertical motion in the atmosphere can result in turbulent mixing and produce violent storms. In this session we will discuss the characteristics of a stable versus an unstable atmosphere and explore several methods that are used to determine atmospheric stability.

Part 1: A Review of Lifting Mechanisms in the Atmosphere

m Orographic Lifting

m Convergence

m Diabatic Heating

m Frontal Systems

Part 2: A Review of Adiabatic Processes

m The Dry Adiabatic Process

m The Moist Adiabatic Process

m The Pseudoadiabatic Process

Part 3: Hydrostatic Equilibrium Revisited

Part 4: Stable, Neutral, and Unstable Atmospheres

m The Stable Atmosphere

m The Neutral Atmosphere

m The Unstable Atmosphere

m Conditional Instability

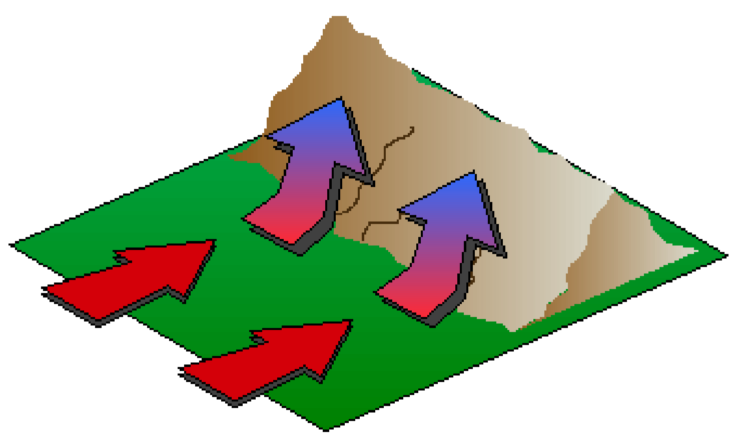
m Convective Instability: Lifting Entire Layers

Part 1:
Lifting Mechanisms in the Atmosphere

Even though the atmosphere at a given time may not be showing a great display of activity, you cannot be too sure that the atmosphere is stable. A bucket of water balanced on top of a slightly opened door (waiting for you to walk through) may appear to be stable. However, with just a small nudge you quickly understand how unstable that bucket really is. You can think of an unstable atmosphere in much the same way as the bucket balanced on the door. The term stable means that the atmosphere is in a state that is resistant to change. An unstable atmosphere may be in balance but not at all resistant to change. A little push in the right direction (in this case, up) may produce all the change necessary to sustain vertical motion, causing turbulent mixing and the production of storms. Here, we will review a few mechanisms which serve as catalysts that can bring about that change.

Orographic Lifting

When air in motion reaches a barrier that it cannot go through or around, it often goes over it. We see this in nature when air lifts over



Orographic Lifting of Air

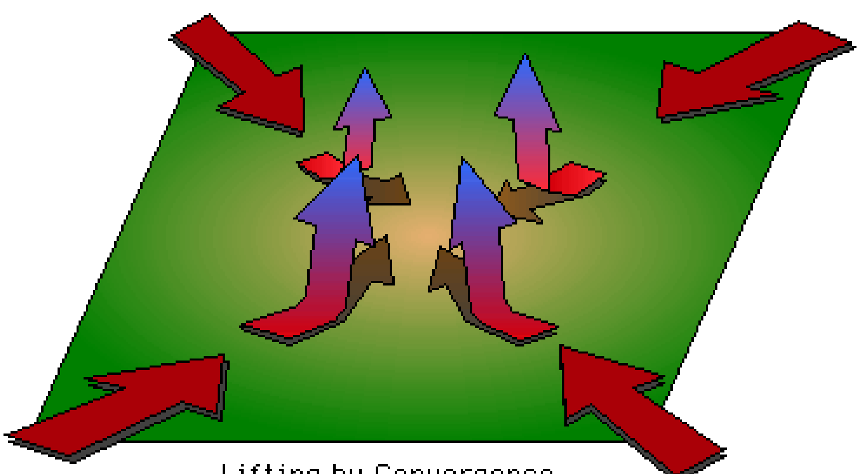
a mountain. This process
of a parcel or layer of air
rising as a result of the

topography is referred to as **orographic uplifting**. If you remember from Session 3, the equation of state and Poisson's equation illustrate the relationship between the temperature and the pressure of an air parcel. As air lifts over a mountain, the pressure and temperature decrease according to the dry adiabatic lapse rate until reaching the lifting condensation level (LCL). Above the LCL the temperature decreases according to the moist adiabatic lapse rate. If the temperature of the rising air decreases faster than the lapse rate of the air around it (the environmental lapse rate), then the parcel will continue to rise only as long as it is forced from below. A parcel of rising air that cools at a slower rate than the environmental lapse rate will continue to rise as a result of its buoyancy. We shall see a little later that this comparison of lapse rates is the process we use in determining the stability of the atmosphere. (Part 4: Stable, Neutral, and Unstable Atmospheres) Click on the image below to view an animation of orographic lifting. Or, if you do not have a frames-compatible browser, you may follow this link to view the animation without frames.

Convergence

Convergence, as we discussed in Session 4, is another mechanism that can force air near the surface to rise. If winds blowing in different directions meet each other, the different moving air masses become an obstacle to one another. The air converges and has no place to go but upwards. At the surface air flows inward to the center of low pressure where it converges and then rises. Convergence also occurs when air

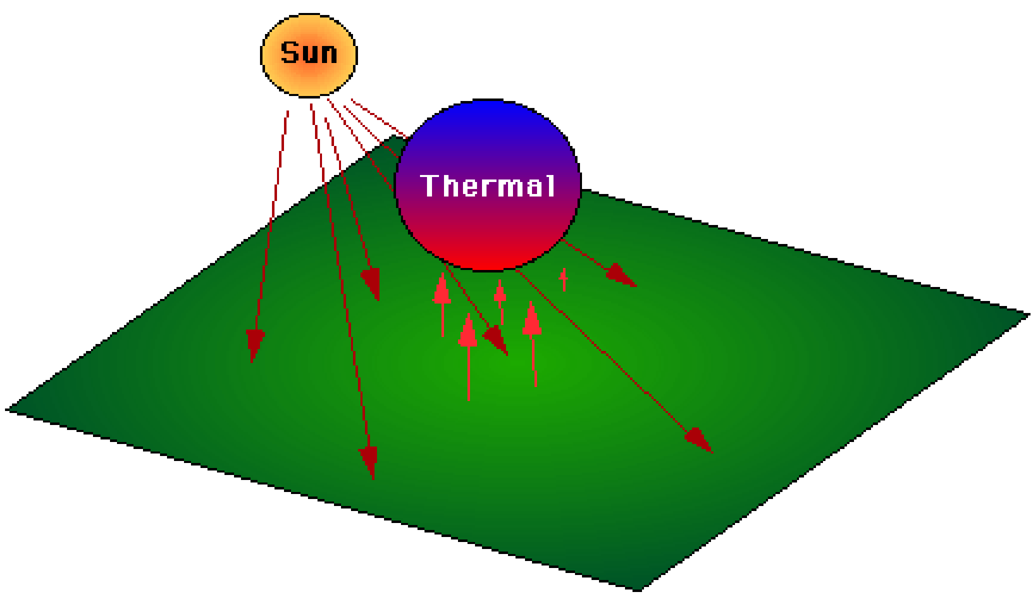
flowing over a smooth surface suddenly hits a rougher surface and slows due to increased friction. The air piles up at the rough surface where the friction is greater, and this causes some of the air to move in a vertical direction.



Lifting by Convergence

Diabatic Heating

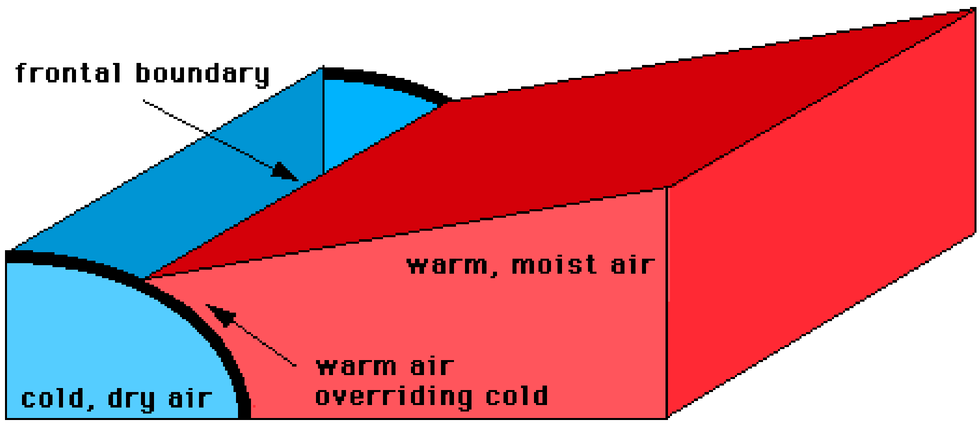
We have mentioned many times how the radiation emitted by the earth heats the air at the surface. When the warm air rises, it may cool adiabatically, meaning without the exchange of heat between the parcel and the surrounding air. The temperature drops in response to the change in pressure. In contrast, we refer to the heating that occurs by the radiation as **diabatic** heating. Diabatic is the opposite of adiabatic. In the case of diabatic heating, the temperature of the air changes, at least in part, in response to the heat that is added by radiation. The air expands as it warms, making it less dense than its environment. Because less dense materials are more buoyant, the warmer air rises producing a thermal.



Lifting by Diabatic Heating

Frontal Systems

The final lifting mechanism which we will discuss is the overriding of air at frontal boundaries. An air mass has certain temperature and moisture characteristics. In fact, these characteristics are the basis of categorizing air masses. For instance, a continental-polar air mass (cP) originates from a continent near a polar region. Continental air tends to be dry while polar air is relatively cold. If a cP air mass meets up with a continental tropical mass (cT), which is dry and warm, the warm tropical air will override or rise above the polar air. Likewise, moist air, because it is less dense, will override dry air. But how is this different from convergence? In the case of convergence, the lifting results from air molecules pushing one another upward, like pushing two small piles of sand together with your hands, forcing a larger pile to form. When two frontal boundaries meet, the lifting that occurs is due to the relative buoyancy of the two air masses. The more buoyant air mass will override the lesser buoyant air mass. The buoyancy is determined by the characteristics of the air masses (i.e., temperature and moisture content).



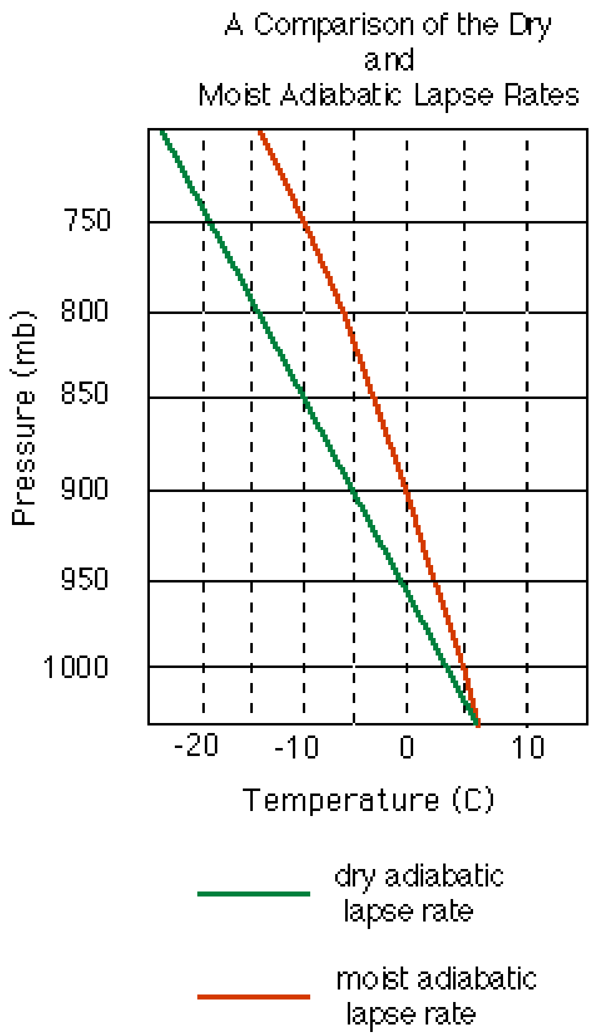
Part 2:
Adiabatic Processes

Knowing the rate at which rising air cools is vital in determining the stability of the atmosphere. We have briefly introduced dry, moist, and saturated adiabatic processes in previous sessions, but because these concepts are so important to the discussion on atmospheric stability, we will take a few minutes to review them, as well as introduce a couple of others that are of equal importance.

The Dry Adiabatic Process

When a parcel of air rises, it expands, and the temperature decreases. Likewise, when air sinks, it compresses, and the temperature increases. This phenomenon occurs without adding or withdrawing energy from the parcel and is illustrated in Session

3 by the equation of state and Poisson's equation. When a parcel of air, either dry or containing water vapor, rises or sinks without the addition or extraction of heat, that process is said to be a **dry adiabatic** process. Even though a parcel of air may contain moisture, if the parcel is rising, then it cools according to the dry adiabatic lapse rate until it reaches the dew point temperature (Td). (Note: The dew point temperature for a rising air parcel is not equal to the dew point temperature of the same parcel at the surface. As the parcel rises, the dew point temperature decreases slightly in response to the decrease in pressure.) We refer to the pressure where the actual temperature equals the dew point temperature as the Lifting Condensation Level (LCL). At the LCL, the cooling process becomes a moist or saturated adiabatic process.



Moist Adiabatic Process

The amount of water vapor a sample of air can hold is dependent on the temperature and pressure of the sample. As a parcel of air cools, its relative humidity increases, provided no moisture is either added or removed from the parcel. In general, as a sample cools, its capacity to hold water decreases. When the air reaches a point of saturation, condensation begins. As the water vapor condenses it goes from a higher energy state to a lower one, and as a result, latent heat is released into the air. In an effort to simplify the process, meteorologists assume that an air parcel is bounded by an imaginary balloon-like skin that contains the parcel and does not allow any mixing with its surrounding environment. Therefore, any latent heat released by condensation is contained in the parcel and is utilized to heat only the sample parcel. So, for the case of a rising, expanding, saturated parcel of air, the only heat added to this theoretically closed system is the heat generated within the parcel itself. The energy of the parcel is said to be conserved because the latent heat is a conversion from existing energy and not energy that was added to the system.

If a rising saturated parcel suddenly changes directions and starts to sink (compression), then evaporation of the condensation products will consume the latent heat to re-form water vapor. Since no heat is exchanged between the parcel and the environment, we still refer to this heating and cooling as an adiabatic process. However, during the processes of condensation and evaporation, the cooling and heating of the saturated parcel varies somewhat from the purely dry adiabatic process we discussed above. A rising saturated parcel cools at a slower rate due to the release of latent heat, and a sinking saturated parcel heats more slowly due to the conversion of heat energy during evaporation. This cooling and heating of a rising and sinking saturated parcel is called a **moist** or **saturated adiabatic** process.

The Pseudo adiabatic Process

We have covered the cases of dry air, unsaturated moist air, and saturated air, so what could possibly be left? Well, we all know from personal experience that if condensation continues long enough we get rain. The process of forming a rain drop is more complex than just simple condensation. But for our purposes here, the point is that all that condensing moisture does not remain in the cloud just to be evaporated at some later time. So, we must address the case where the moisture precipitates out of the cloud.

If you remember our discussion of temperature in Session 3, all substances are composed of molecules in motion which have kinetic energy. As rain, snow, or any other form of precipitation falls out of a cloud, it carries with it that energy it possesses. With this true loss of energy, the process is no longer adiabatic and, therefore, is called **pseudo adiabatic**. Fortunately, the amount of energy lost through precipitation is very small compared to the energy of the air molecules. The **pseudo adiabatic lapse rate** is so close to the moist adiabatic lapse rate that meteorologists tend to ignore this difference and often refer to them synonymously. In fact, on adiabatic charts (or pseudo adiabatic charts), the moist adiabatic lapse rate lines are referred to interchangeably as moist, saturated, or pseudo adiabatic.

Part 3:
Hydrostatic Equilibrium Revisited

In the remainder of this session, we will be evaluating the stability of a layer of air in the atmosphere. First, we will look at the criteria of stability in reference to a single isolated parcel being lifted through a layer of the atmosphere. Later, we will evaluate the atmosphere in reference to an entire layer being lifted. To simplify matters, we will assume that the **environmental** air, or the air surrounding the rising air, is in hydrostatic equilibrium. Let's take a few minutes to briefly review the basics of an atmosphere that is in hydrostatic equilibrium.

$$\mathrm{PGF}_{(\mathrm{vertical})} = \mathrm{Gravity}$$

$$-\frac{1}{\rho}\frac{dp}{dz} = g$$

$$\frac{dp}{dz} = -\rho g$$

You were introduced to the **hydrostatic equation** in Session 4's discussion on Hydrostatic Equilibrium. It expresses the relationship between the pressure gradient force in the vertical, which forces air molecules upward, and gravity, which forces them downward. By this equation, the pressure gradient force and gravity are equal to each other, and there is no motion either upward or downward, that is, if there are no other forces acting on the air molecules. So, why do we need to assume the environment is in hydrostatic equilibrium for our discussions on stability? Well, by assuming a hydrostatic environment, we can greatly simplify the motion in the atmosphere in order to see the basic responses of the parcel air and the environmental air while a parcel or a layer is being lifted. On the scales that we are discussing, the meso- and microscales, the atmosphere probably is not in hydrostatic equilibrium. There very well could be other forces acting on the environmental air surrounding a rising parcel. Friction and surface heating are two factors which could cause vertical motion in the environmental air. When the environment is not in hydrostatic equilibrium, issues can get very complicated very quickly because you must take into account these forces and responses when determining stability. We just want to understand the basics here, so we will save nonhydrostatic complications for another discussion.

Part 4:
Stable, Neutral, and Unstable Atmospheres

The stability of the atmosphere is basically determined by comparing the lapse rate of a parcel of air to the lapse rate of the surrounding air, which we also refer to as the environment. If we know the temperature and dew point of the air parcel before it begins to rise, then we can pretty accurately determine the temperature change as it rises, as discussed in A Review of Adiabatic Processes. Radiosonde and/or remote sensing data provide us with a profile of the environment with which to compare our rising air. But, keep in mind, our profile is merely a snapshot of the atmosphere. The sounding only gives us a view of the atmosphere at a point in time. Motion in the atmosphere makes these comparisons complicated because motion causes changes in the lapse rates we wish to compare. In order to simplify our discussion, we are going to make a few assumptions. These will be helpful to remember as we talk about stability.

We will assume:

1. The atmosphere is in hydrostatic equilibrium.

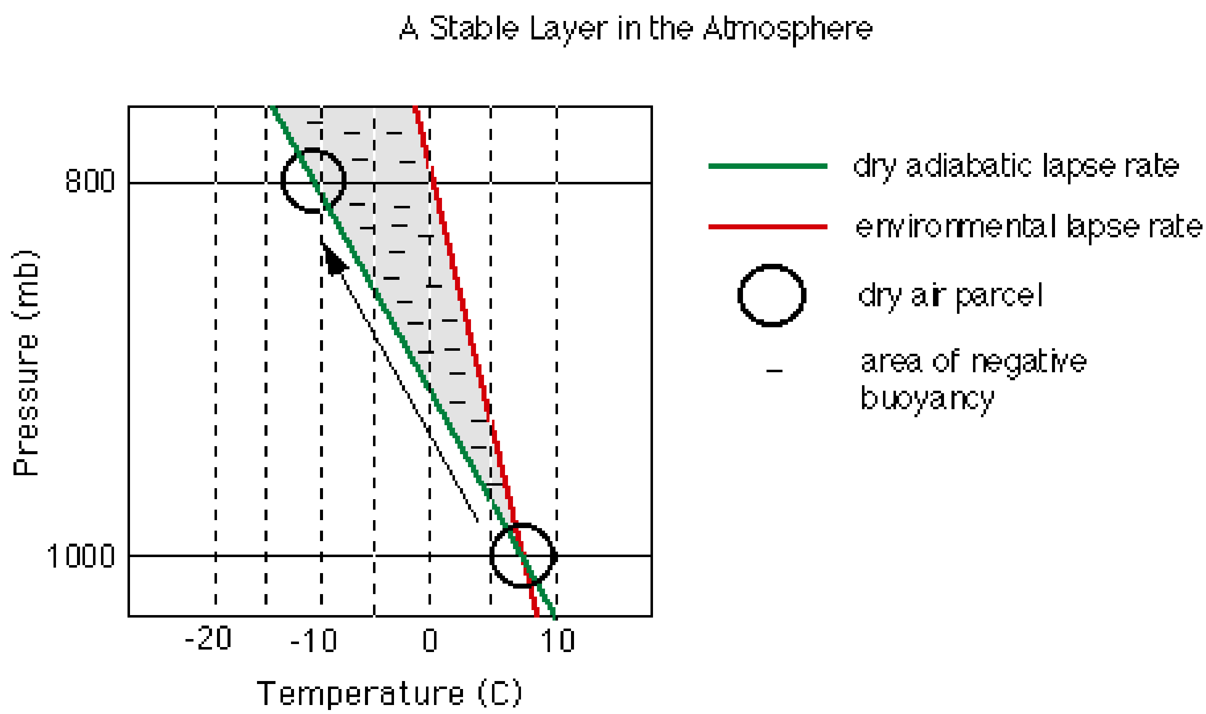
2. As a parcel of air rises or sinks, there is no compensating motion in the displaced environmental air. The environment around the parcel is static.
3. The rising or sinking parcel is isolated from the environment such that the rising and sinking air and the environmental air do not mix.

We have found that there are three basic categories in which the atmosphere or a layer in the atmosphere can be classified in terms of stability. These categories, which we discuss next, are stable, neutral, and unstable.

The Stable Atmosphere

A stable atmosphere is one that is strongly resistant to change. If some external force such as orographic lifting or convergence pushes the air upward, the temperature of the rising air relative to the environment suggests that the air would prefer to go back to its original position. In other words, though a parcel is being forced up, it has negative buoyancy meaning it wants to sink to its original position where it was in equilibrium with the environment. If pushed down, the air has positive buoyancy and wants to rise. Imagine a cork floating on a lake. If you were to push the cork under water then release it, the cork would float back to the surface. Likewise, if you were to lift it out of the water then let it go, it would fall right back to the surface. Pushed under the water, the cork has positive buoyancy due to the difference in the density of the cork and the water. Lifted up out of the water, the cork has a negative buoyancy. But wait, doesn't the cork fall back to the water due to gravity? Gravity does pull the cork back to the surface, but it also has the same pull on the air molecules. The cork is more dense than the air so it falls or sinks to the surface where it is in equilibrium with its environment. This is very much like a stable atmosphere.

In a stable atmosphere, if you lift a parcel of air, the temperature of the rising air will decrease fast enough that its temperature will always be colder than the temperature of the environment. Colder air sinks. If the force pushing the air up suddenly disappeared, the parcel would sink back down to its original position where its temperature and pressure would be in equilibrium with the environment. Another way of stating that the atmosphere or a layer in the atmosphere is stable is to say that the lapse rate of the rising air is greater than the lapse rate of the environment. (Note: A positive lapse rate indicates a decrease in temperature with height.) A layer characterized by a temperature inversion, defined by a negative lapse rate, is considered extremely stable. These inversions near the surface often occur in the early morning hours before sunrise.

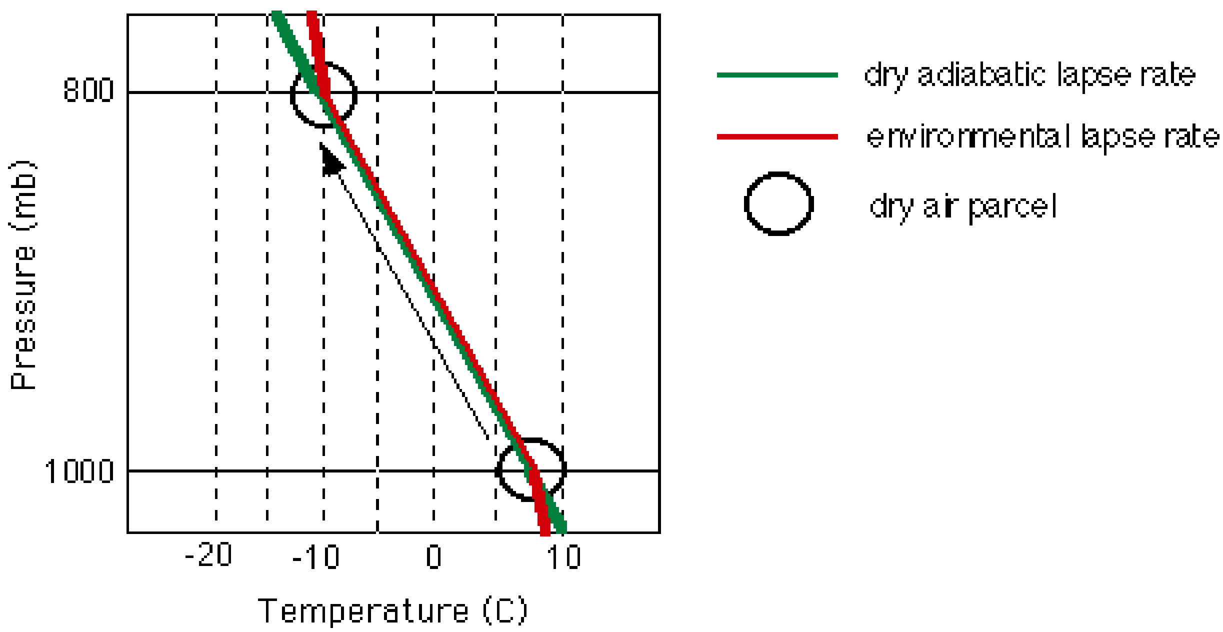


In the diagram above, a parcel of dry air is lifted from a pressure of 1000 mb (100.0 kPa) at 7 degrees Celsius to a pressure of 800 mb (80.0 kPa). For simplicity we used a dry air parcel so the entire expansion is a dry adiabatic process. A comparison of the dry adiabatic lapse rate to the environmental lapse rate reveals that at every pressure level from 1000 mb (100.0 kPa) to 800 mb (80.0 kPa) the rising parcel is colder than the environment. If at any time during the expansion the lifting force disappeared, the parcel would sink back to the 1000 mb (100.0 kPa) pressure level. For example, at 800 mb (80.0 kPa) the temperature of the parcel is -10 degrees Celsius while the temperature of the surrounding air is about 1 degree Celsius. In the diagram, at every point above 1000 mb (100.0 kPa) the parcel has negative buoyancy because it is colder than the surrounding air. This is an example of a stable atmosphere. If the parcel in the example above had been a moist parcel, we would compare the environmental lapse rate to the dry adiabatic lapse rate below the LCL and to the moist adiabatic lapse rate above the LCL. The next classification of stability we will discuss is the neutral atmosphere.

The Neutral Atmosphere

While in a stable atmosphere or layer, the lapse rate of a rising air parcel is greater than the lapse rate of the environment. In a neutral atmosphere, the two lapse rates are equal. If a parcel of air is lifted through a neutral layer, the temperature and pressure of the parcel will be identical to the temperature and pressure of the surrounding air at every height and is always in equilibrium with the environment. Thus, the parcel is not buoyant. If the force producing the motion ceases, the parcel will neither continue to rise nor begin to sink, rather, the motion of a parcel will also cease.

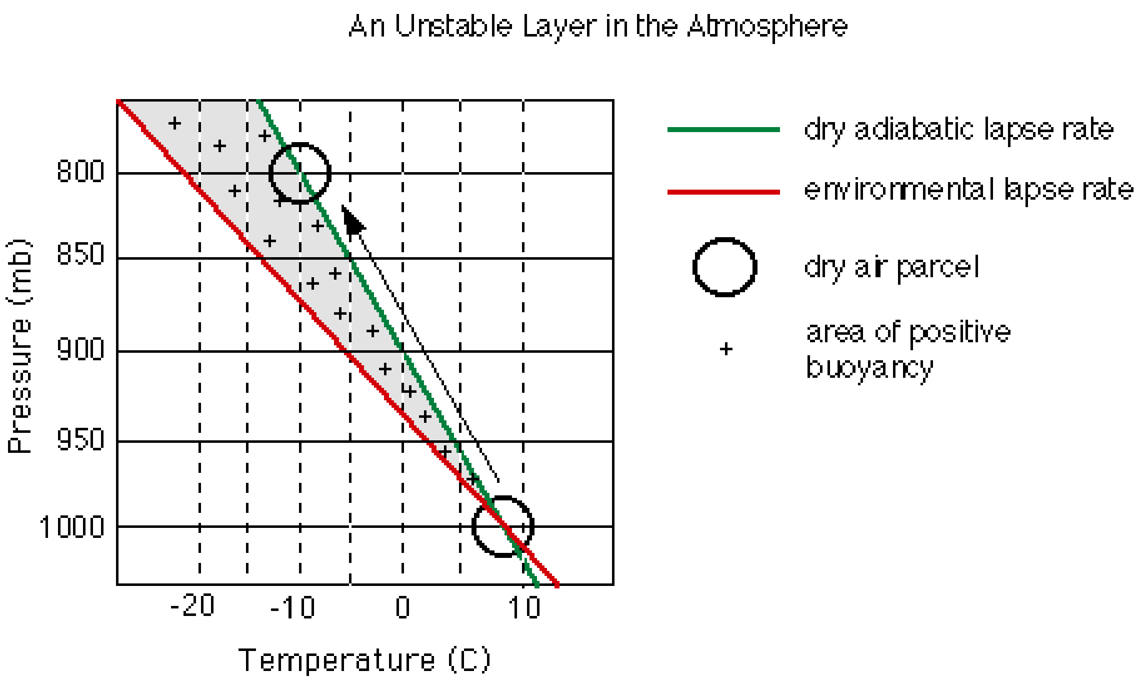
A Neutral Layer in the Atmosphere



In the diagram above, our dry parcel is lifted, once again, from 1000 mb (100.0 kPa) to 800 mb (80.0 kPa) but in different atmospheric conditions. As the parcel rises and cools according to the dry adiabatic lapse rate, it remains at the same temperature and pressure as the surrounding air. For instance, at 1000 mb (100.0 kPa) the parcel and the environment have a temperature of about 7 degrees Celsius. After the parcel is lifted to 800 mb (80.0 kPa), the temperature of the parcel and the environment is -10 degrees Celsius. At every height in this neutral layer, the parcel is in equilibrium with the environment.

The Unstable Atmosphere

At times, in the atmosphere, a little push goes a long way. If a parcel of air is lifted and continues to rise after the lifting force disappears, the atmosphere is unstable. In an unstable layer, the lapse rate of a rising parcel is less than the lapse rate of the environment. Though the parcel cools as it rises, its temperature remains warmer than the surrounding air during its ascent through an unstable layer. Because the parcel is warmer than the environment, the parcel has positive buoyancy and continues to rise on its own.



In the diagram above, we have altered our atmospheric conditions slightly, once again, so that we may illustrate our dry air parcel rising through an unstable layer. In contrast with our previous examples, rather than lifting the parcel, we only need to give it a little push to get it going. Once the parcel gets just above the 1000 mb (100.0 kPa) pressure level, we find that the temperature of the parcel is actually warmer than the environment. Just as before, at 1000 mb (100.0 kPa) our parcel and the surrounding air are about 7 degrees Celsius. At 950 mb (95.0 kPa) the parcel has cooled to about 5 degrees Celsius but is 2 degrees warmer than the environment, and thus, has positive buoyancy. The parcel will rise on its own as long as it remains warmer than the surrounding air.

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General questions

- Define Parallels, Prime meridian, Great circle, Small circle, Date line, Latitude, and Longitude.
- Define Observer's Horizon, Celestial Projection, Celestial Equator, Celestial axis and poles, Elevated Pole, Vertical Circles, Celestial Meridians, Hour Angles.
- Define the azimuth and the Altitude of a celestial body
- Define the declination of the celestial body
- What do we mean by the Greenwich Hour Angle of the body

The Greenwich Hour Angle of the body is the arc on the celestial equator between the Greenwich Meridian and the celestial meridian of the body. It is called the hour angle because due to the earth's rotation, 15 degrees of longitude corresponds to 1 hour.

- What do we mean by Spherical triangle

When drawn on the same sphere the vertical circle of the body and the celestial meridian of the body form a spherical triangle. Its three corners are the zenith (Z), the elevated pole, and the location of the body on the celestial sphere.

- Define the altitude angle

It is the angle between the horizontal plane at the measurement location and the line connecting the viewer and the sun's disk.

- Define azimuth angle

Azimuth varies from 0° to 360° starting from north to east then to south, west, and north.

- Draw the spherical triangle
- Write down the first law of the spherical triangle
- Complete the following sentences:
- the declination angle changes during the year, reaching its maximum on............., and its value is 23.5 degrees, when the sun is perpendicular to the Tropic of Cancer.
- The lowest possible declination angle is .... degrees on December 21 when the sun is perpendicular to the Tropic of Capricorn.
- Declination angle is equal to zero on both March 21 and September 21 when the sun is perpendicular to ....
- Write down the solar declination formula:

$$\boxed{\delta_s = 23.45 \sin\left[360 \frac{n-82}{365}\right]}$$

- Define the local hour angle

it is the angle between the meridian of the observer and the meridian of the geographical position of the celestial body (GP).

Chose the correct answer:

□ Due to the Earth's rotation, the Sun moves through (15^o , 20^o, or 30^o) of longitude in 1 hour.

Write the equation of hour angle calculation

$$\boxed{h_s = 15(T - 12)}$$

What do we mean by the equation of time

The equation of time describes the discrepancy between two kinds of Solar time; the apparent solar time and the mean solar time. The apparent solar time directly tracks the diurnal motion of the Sun, and the mean solar time tracks a theoretical mean Sun with Noons 24 hours apart.

Apparent solar time can be obtained by measurement of the current position (hour angle) of the Sun, as indicated (with limited accuracy) by a sundial.

Mean solar time, for the same place, would be the time indicated by a steady clock set so that its differences from apparent solar time would resolve to zero over the year.

Write down the azimuth angle equation

$$\boxed{\sin \alpha_s = \frac{\cos(\delta_s)\sin(h_s)}{\cos(\alpha)}}$$