Basic Meteorology:

A Short Course

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2013

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Preface

As a sailor (well, motorboater) I have long had a practical interest in weather. Many (too many) are the times when I have been in heavy seas wondering how this came upon me. The worst to date was in the Caribbean. Following an absolutely glorious ride from St. Martin to Antigua (flat seas, warm sky, the green flash) strong Christmas Winds and high seas kept us in Antigua for a week. Finally we had to return to St. Martin in what I judge to be 15-foot breaking seas on our starboard quarter. The only good outcomes were that we made it and that I developed a confidence in the roll angle of our boat,

Retirement gave me the time to learn more about the weather so I bought a DVD on Meteorology from the Great Courses Institute. It was outstanding, as was the previous course I had bought (*Quantum Mechanics*, taught by Kenyon's Benjamin Schumacher). *Meteorology* is taught by Robert G. Fovell, a professor of Atmospheric Science at UCLA. Aimed at the undergraduate, it is an excellent balance of theory and exposition. I highly recommend this course and the Great Courses Program!

So I took careful notes and wrote my prose version of Fovell's video course so that I could share it with others interested in weather. And here it is! I hope that you find it useful. If so, all credit goes to Fovell, whose amanuensis I became.

Peter Fortune
Naples, Florida
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1. Earth's Atmosphere

In this section we look at the basic structure of Earth's atmosphere, beginning with the Ideal Gas Law summarizin the relationship between three major weather-creating actors: air pressure, air density, and air temperature. We then look at the chemical composition of the theoretical Standard Atmosphere, and at the influence of *greenhouse gases*.

Temperature, Pressure and Density

The Ideal Gas Law describes the relationship between gas temperature (T), density (ρ) and the gas constant (R) for the specific characteristics of the gas. T is the kinetic energy of the gas created by vibrations and collisions of its atoms; P is the force per unit area—at sea level the atmospheric pressure is about 15 pounds per square inch (psi), 30 inches of mercury or 1000 millibars (mb); ρ is the density of the air, mass per unit of volume (for example, kilograms per cubic meter).

The Law says that air pressure is proportional to both density and temperature, the proportionality constant (R) varying with the specific gas.

Ideal Gas Law

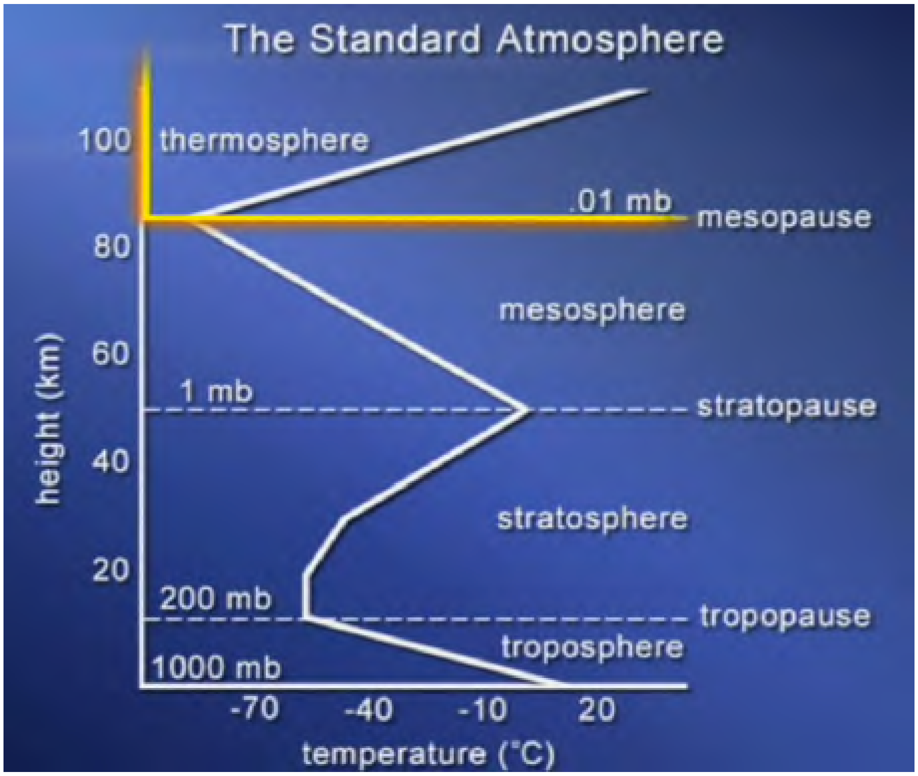
$$P = R(\rho T)$$

P = pressure, ρ = density, T = temperature, R is gas-dependent constant

Thus, given density and R, pressure increases with temperature; given temperature, pressure increases with density; and given pressure, temperature and density are inversely proportional—increased temperature means proportionally lower density.

The Standard Atmosphere

The *Standard Atmosphere* is derived from global averages of temperature, density, chemical composition of air, and so on, ignoring the regional and local variations that are the central source of weather. A brief picture of the Standard Atmosphere is shown below.



Altitude, Temperature, and Pressure

The lowest portion of the atmosphere is the *troposphere*, the "turning layer" at the Earth's surface, extending from the surface to about 12km altitude. The troposphere has air pressure of 1000mb at the surface and 200mb at its top (the *tropopause*),with a temperature range of 20°C at the surface to -60°C. As altitude increases, temperature falls, the air pressure declines and air density diminishes.

Next is the *stratosphere*, an area between 12km and 80km altitude, with a pressure range of 100mb to 1mb and a temperature range of -60°C to 0°C. The stratosphere has the unusual property of *temperature inversion*: temperature increases with altitude because, although air density and pressure decline, the sun's radiation is absorbed in the stratosphere's *ozone layer*. The rising temperature in the stratosphere acts to increase air density and pressure.

Our daily experiences revolve around the troposphere and the stratosphere, where most of our weather-creating actions take place. Above the stratosphere is the *mesosphere*, extending from 50km to 85km. In this "middle layer" air temperature declines from 0°C to -80°C, pressure falls from 1mb to 0.1mb, and air density declines.

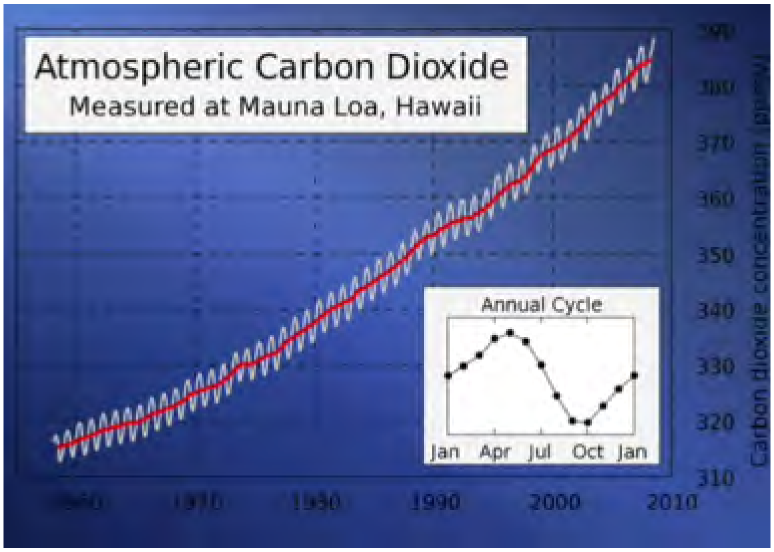
Finally we get to the *thermosphere*, from 85km up. In this "heated layer" pressure falls from 0.01mb to 0mb and temperature rises from -80°C to extremely high levels as the sun bakes the atmosphere.

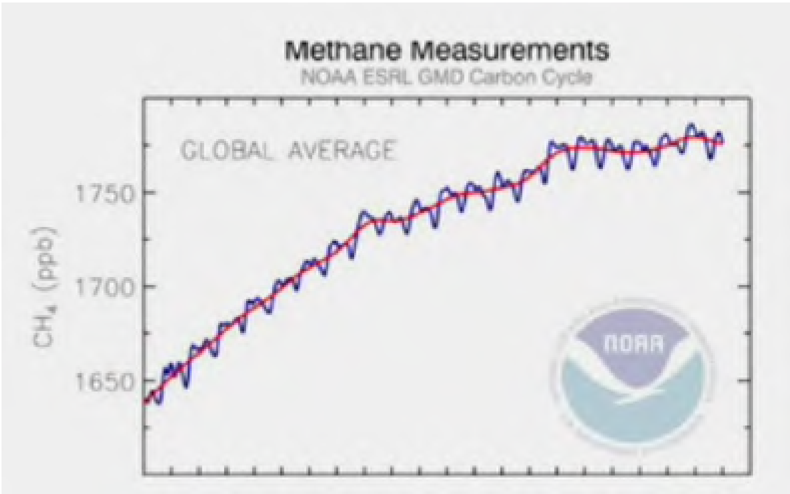
<u>Chemical Composition</u>

The atmosphere's chemical composition depends in part on the amount of its water vapor, which is concentrated in the troposphere. The standard atmosphere's chemical mix is "Dry Air," defined as containing moisture but at subsaturation level. Rounding to nearest percentages, the major constituents of dry air are: 78% Nitrogen (N_2), 21% Oxygen (O_2), and 1% Argon (AR). There are other minor, but very important, constituents called *greenhouse gases*. Chief among these are Carbon Dioxide (CO_2), inversely correlated with plant growth (hence highest in in the northern hemisphere's winter and lowest in its summer); Methane (CH_4), less abundant but more potent, is directly associated with plant decay and ruminants (cattle, sheep); Nitrogen Oxide (NO_2), Ozone (O_3), and water vapor (H_2O) make up a very tiny proportion of the chemical mix but round out the actors in the greenhouse effect. Ozone, created by lightning, is concentrated in the stratosphere.

A purely manmade greenhouse gas, Chlorofluorocarbon (CFC), is used as a propellant in aerosols and a refrigerant in air conditioners. CFCs attack the radiation-absorbing ozone layer by interacting with solar radiation to make Chlorine, a gas that destroys ozone.

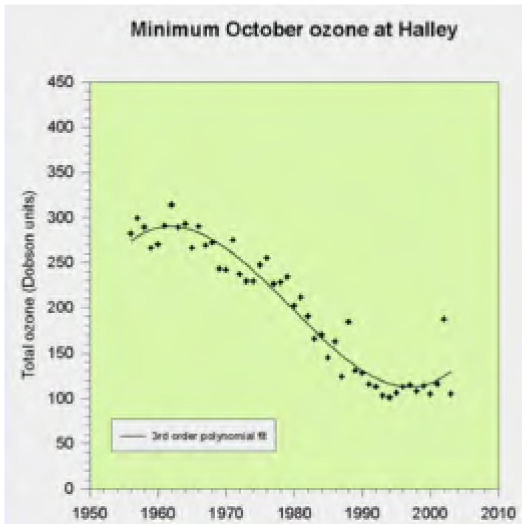
The Standard Atmosphere's chemical composition of air has changed considerably in the past 50 years. The chart below shows a steady carbon dioxide increase from 310 parts per million (ppm) in 1960 to 390 ppm in 2010. The annual cycle follows the cycle of plant growth. The same trend is shown in methane content.





<u>Greenhouse Gases and the Ozone Hole</u>

Ozone absorbs the ultraviolet radiation that creates damage from sunburn to cancers. The ozone hole, discovered in 1979, is over Antarctica even though much of its source is in the northern hemisphere. Greenhouse gases drift in atmospheric winds from the north to the south. As noted above, the hole arises from the interaction between solar radiation (photons) and greenhouse gases. This interaction creates chlorine, which bonds with ozone and removes it from the atmosphere.



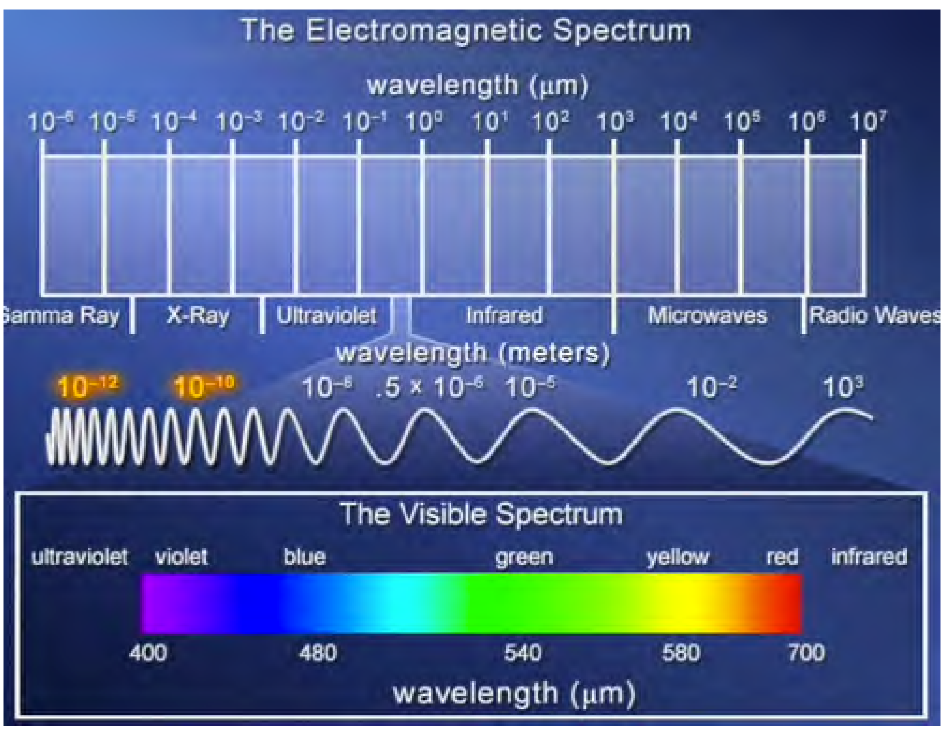
The ozone hole at the Halley station in Antarctica has clearly declined since 1950, though it seems to have stabilized in 2000-2004. Whether this is temporary is not known.

2. Radiation and Absorption

Energy Radiation and Temperature

All "things" both radiate and absorb energy in *all* wavelengths. However the most prominent wavelength ranges emitted by an object are different from those absorbed by the same object. For example, we will see that the Sun's average energy radiation is at short wavelengths of ultraviolet to visible light but Earth's atmosphere absorbs mostly UV and infrared (IR) light, giving passage to visible light. The Earth also re-radiates energy at longer wavelengths than the energy it absorbs—mostly in the IR range.

The electromagnetic spectrum of energy is shown below. The visible range is at wavelengths of 400 – 700 μm (microns, or millionths of a meter). Shorter wavelengths are in the invisible UV and gamma ray range, while longer wavelengths are in the invisible near-IR and IR ranges.



The *intensity* of radiation energy is the energy emitted per unit volume of the emitted light, measured, for example, in joules per cubic meter). Thus, intensity can be increased by raising the energy content (joules) or by reducing the area into which a unit of energy is emitted. Energy is related to temperature through the *Stefan-Boltzman Law*: the intensity of radiation is proportional to the fourth power of temperature. The Sun and Earth have average temperatures of 5800°K and 260°K, so the Sun's average radiation is 247,650 times that of Earth's.

Radiation Intensity and Temperature

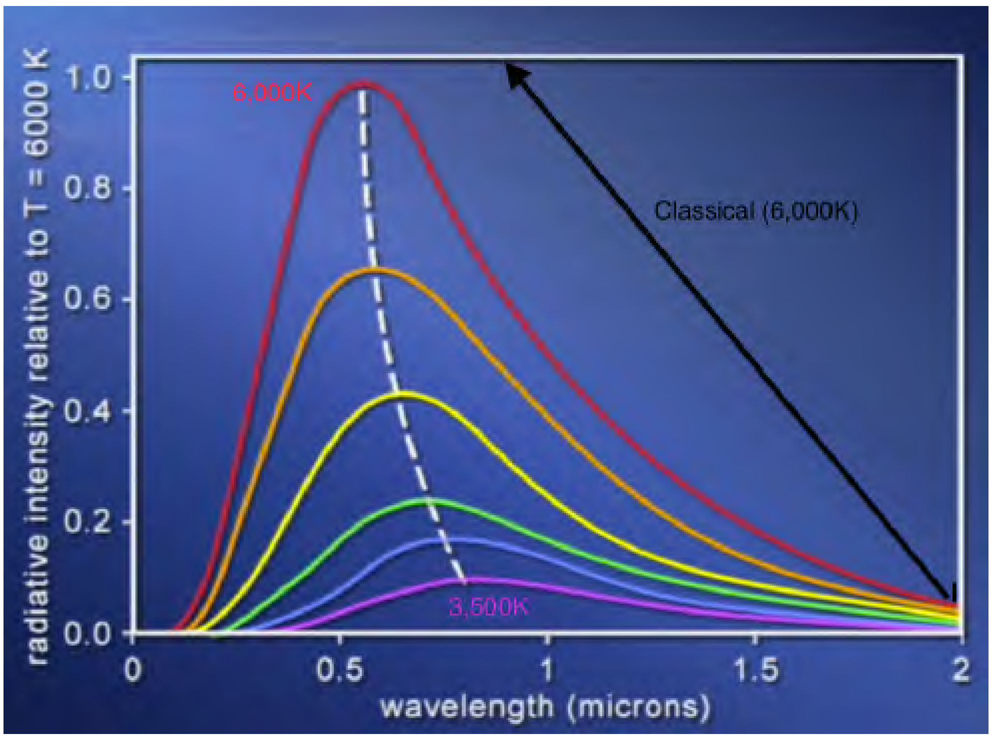
Stefan Boltzman Law: $I = \sigma T^4$

I = Intensity, or Energy per cubic meter; T = Temperature;
σ = the Stefan-Boltzman Constant

Planck Curves and the Ultraviolet Catastrophe

At the turn of the 20th century a controversy developed over the relationship between radiation intensity and the temperature of an ideal concept called a *blackbody*, an object that absorbs all radiation it receives and radiates all the energy it absorbs. Classical physics predicted that the intensity of radiation by a blackbody at a given temperature increase indefinitely as the radiations' wavelength shortens, and that as the wavelength enters the ultraviolet range at about 400 microns (μm) the ultraviolet energy intensity becomes infinite. This is shown as the "Classical Theory" line in the figure below (it assumes a 5000K temperature). This was called the *ultraviolet catastrophe* because it implied that a blackbody emitting extremely short wavelength radiation would fry the Earth.

Clearly something was wrong. The experimental evidence contradicted the ultraviolet catastrophe, as shown in the Planck Curves below: the Planck Curve for 6,000°K is quite different from the classical curve for the same temperature: as wavelength shortens a blackbody at a given temperature emits more intense radiation but only up to a point, after which intensity actually declines. The figure below shows several Planck Curves, each drawn for a specific temperature from 6,000°K (red) down to 3,500°K (violet). At higher blackbody temperatures the intensity of radiated energy at every wavelength is higher (a la Stefan-Boltzman). And at each blackbody temperature the energy intensity reaches a peak, then falls toward zero at extremely short wavelengths. We are rescued from the ultraviolet catastrophe.



The area under a Planck Curve measures 100% of the radiation emitted by a black body at that temperature, so the area under the curve for any wavelength range is the proportion of radiation emitted in that range. For example, the area under the red Planck Curve between 0.5 and 1.0 microns is the proportion of radiation emitted by a 6,000°K black body that is in that wavelength range.

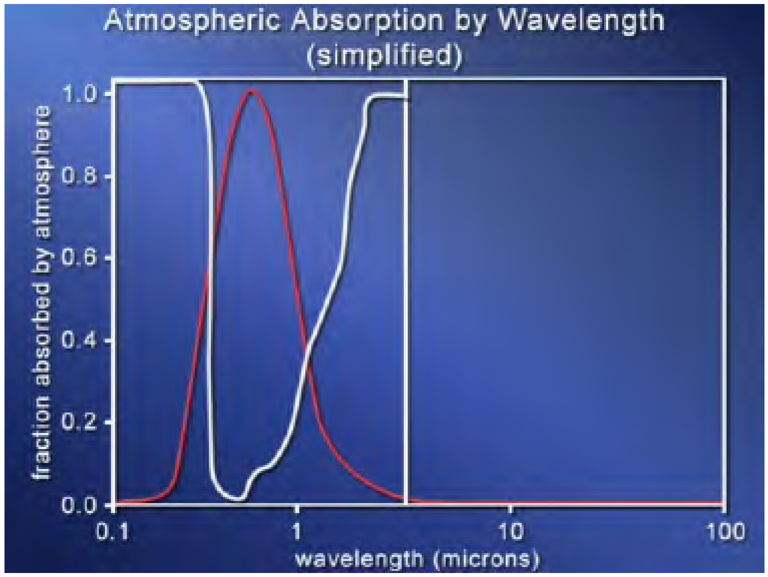
The Planck Curves show several important features of the relationship between energy intensity and its wavelength: (1) as temperature increases, the peak intensity is at lower wavelengths. Thus, the hot Sun (about 5,800°K) emits its peak intensity in the UV to visible light range, but the Earth (260°K) emits its maximum intensity in the Infrared range. This has been institutionalized as *Mean's Law*—as temperature rises the peak intensity occurs at shorter wavelengths—and is shown by the dashed white curve; (2) for any chosen wavelength range, the total intensity emitted is greater at higher temperatures—the sun emits more intense radiation (more energy per cubic meter) at *all* wavelengths than does the Earth; and (3) a cool body like Earth will radiate in the longer near-IR and IR ranges.

As an aside, Planck explained the failure of classical theory by postulating the notion that energy is emitted in discrete amounts, called *quanta*. The quantum (minimum energy packet) is higher at short (more energetic) wavelengths, so it becomes increasingly difficult to input enough extra energy to raise the intensity level farther as wavelength gets shorter. It is as if higher and higher speed bumps are encountered as energy intensity increases. This hypothesis explained the peak, then decline, of intensity as wavelength shortens, and it was the beginning of *quantum physics*.

Energy Emission and Atmospheric Absorption

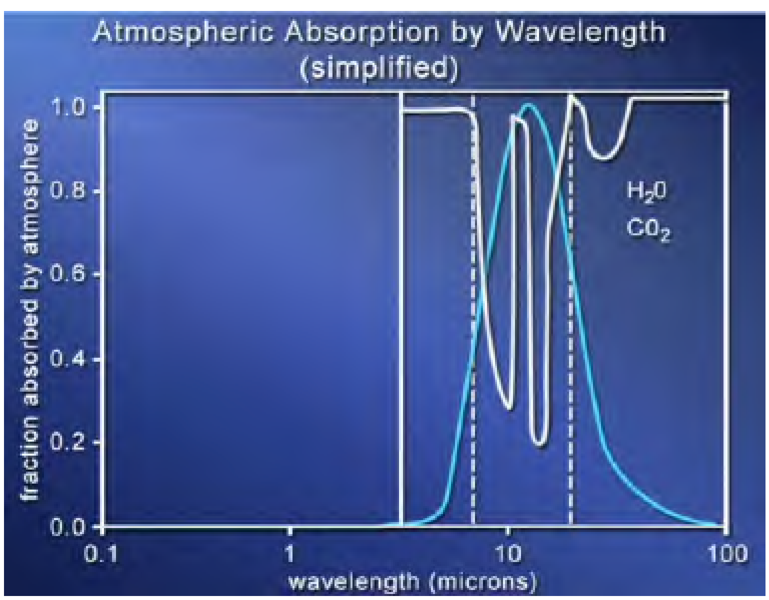
The figure below shows the proportion of light emitted by the Sun (the red line) at each wavelength and the proportion of that radiation absorbed by Earth's atmosphere (the white line). The Sun emits about half its energy in the UV and visible light range, and about a quarter of its radiation in the UV and IR ranges. Earth, on the other hand, absorbs most of the sun's UV radiation and IR radiation, leaving the visible light to pass through to the surface. Thus, our atmosphere gives us what we need (visible light) and stops much of the bad stuff (UV light).

The second figure below shows the wavelengths of the Earth's radiation from its internal energy and re-radiation of the Sun's energy (the blue line),and the wavelengths absorbed by our atmosphere (the white line). Earth's atmosphere absorbs most of the far infrared radiation, and most of the low visible range (blue to violet) but it lets a large share of the mid-to-upper visible light pass through.



Wavelengths Emitted by Sun and Absorbed by Earth

The figure below shows the wavelengths of the Earth's radiation from its internal energy and re-radiation of the Sun's energy, along with atmospheric absorption. Earth's atmosphere absorbs most of the far infrared radiation, and most of the low visible range (blue to violet) but it lets a large share of the mid-to-upper visible light pass through.



Earth's Radiation Emissions and Absorption

3. Atmospheric Heat Transfer

Earth's atmosphere heats from the ground up as the Sun's radiation warms the Earth's surface and the energy is re-radiated upward. In the process, different parts of the Earth develop different temperatures—there are often strong global, regional and local variations in temperature. The most obvious example of a global difference is the steady 100°F difference between the equator and the poles.

<u>Global Temperature Differences</u>

Regional and local temperature differences arise for a variety of reasons. The primary cause of global variation is the *tilt of the Earth's axis* relative to the Sun: the Sun's radiation travels through more atmosphere at the poles than at the equator, allowing greater atmospheric absorption—and less warming—at the poles. Axis tilt also creates annual seasons are also due to the: as the Earth rotates around the Sun, the North pole tilts toward the sun in the northern summer and away from the sun in the northern winter; the opposite is true of the South pole. Winter in Chicago is summer in Sydney!

A second—but minor—factor in global temperature variation is the eccentricity of the Earth's orbit: the Earth is close to the sun at some points and farther away at other times. But this effect is dominated by other factors and each hemisphere remains coolest in winter and warmest in summer.

An important source of regional and local temperature differences is *the material that absorbs the Sun's radiation*. Each material differs in its *heat conductance* and *thermal inertia*. Heat conductance is the direct transfer of heat from high temperature and rapidly-vibrating atoms toward low temperature and slowly-vibrating atoms, as when air molecules are in contact. Materials like air, with its low density, have very low heat conductance, while water has high density and high heat conductance, so water transfers heat by this method better than air. Metals have extremely high density and therefore are very efficient heat conductors.

Thermal inertia, once called *heat capacity*, is the resistance of a material to temperature change. Some materials, like sand on a beach, have low thermal inertia—it heats up quickly and cools quickly. Water has higher thermal inertia, taking longer to effect a given temperature change.

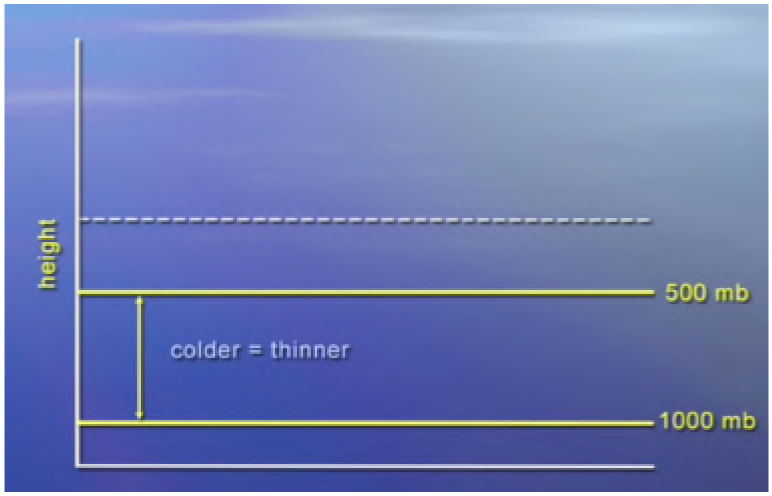
To add to the regional and local variations we can look to topological differences in the surface of the Earth. Valleys tend to get less radiation and to be cooler. Hills get more radiation and are warmer.

In general, conductance is a weak source of temperature differences. *Heat convection*—transfer by wind arising from pressure differences—is the primary source of temperature equalization. Cold air (at lower pressure and density) flows toward warm air. This *wind shear* can be vertical, as when winds are faster (or lower) at higher altiudes. Or it can be horizontal, as when wind speeds differ at different locations: vertical shear appears horizontal (right-left) in an atmospheric cross section, while horizontal shear appears as vertical (up-down) in the cross-section. This is an important part of nature's way of moderating extremes in temperatures at different altitudes or locations

<u>Wind Shear</u>

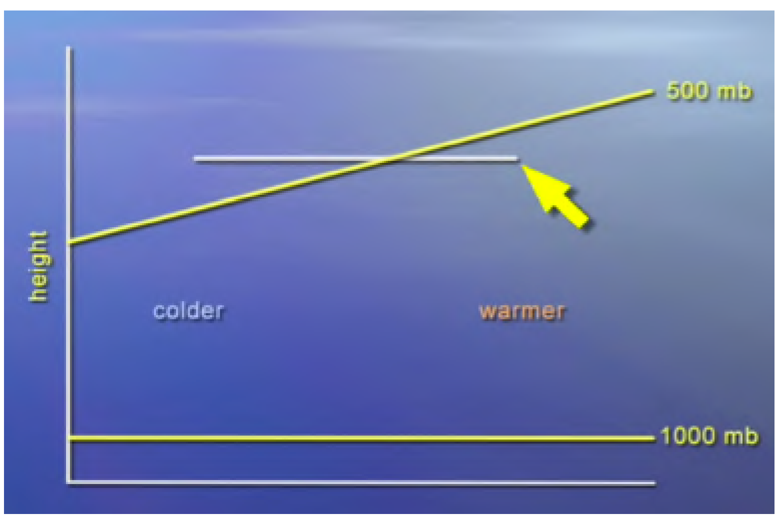
Wind shear is created when air moves by *convection* from high pressure/density areas to low pressure/density areas, or from low temperature to high temperature areas by *advection*. The figure below shows air between altitudes with 1000mb and 500mb pressure. The dashed white line shows the 500mb altitude in an initial situation with constant air pressure at each altitude; that line is an *isopressure line*. Now suppose that the air cools; this reduces the altitude at which a 500mb prevails. The altitude difference between 500mb and 1000mb isopressure line has narrowed

The second figure below shows the effect of a horizontal difference in temperature—a local or regional variation. The cooler air at the left becomes compressed so that the 500mb altitude drops, while the warmer air on the right expands and the 500mb pressure altitude rises..



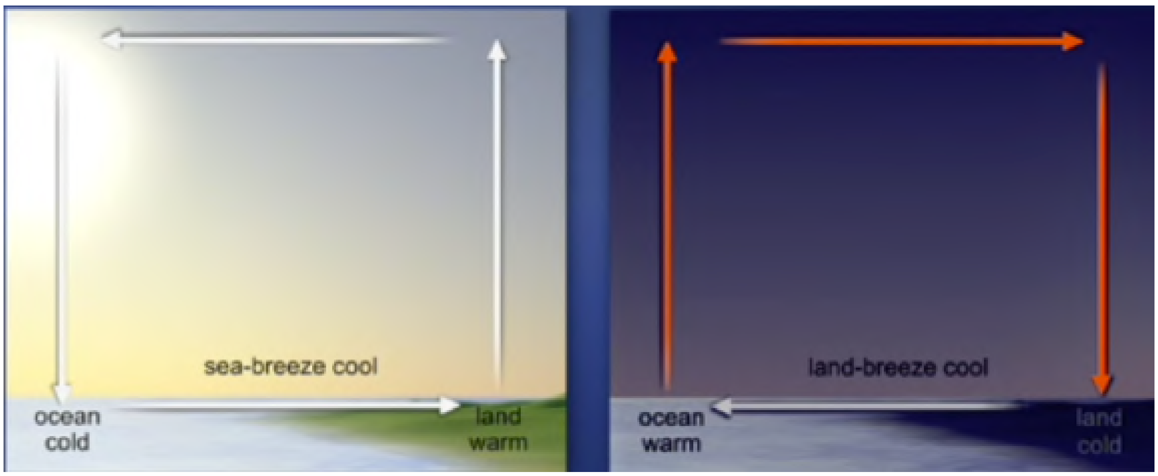
Temperature and Pressure

The isopressure lines now tilt to the "northeast." As a result, at every altitude (say, along the solid white horizontal line) there is a pressure/density difference: more dense on the left, less dense on the right. The denser cooler air on the left now flows toward the right: convection has created a pressure-driven wind from left to right.



Wind Created by Convection

This is the principle behind land and sea breezes, shown below. When the land is warmer than the water (as in the left chart) the surface air is onshore (from sea to land—a sea breeze) and the higher air moves in the opposite direction; a counterclockwise vertical circulation develops. When the land is cooler than the sea, the circulation reverses as land cools more than water, so at the surface there is a land breeze. Because the Earth's land has less thermal inertia than water, the land heats and cools more, and more quickly, so sea breezes tend to occur in the day and land breezes at night.



Sea Breeze Circulation **Land Breeze Circulation**

<u>Katabatic Winds: California's Santa Ana</u>

Katabatic winds are winds that rise over mountaintops and plunge downward on the other side. When air from a dry region is pushed against a mountain it rises and cools. As it falls over the backside, the air expands air pressure and density fall,and the air acquires latent heat. This creates high winds and air temperature increases. This is called the *dry adiabatic process of temperature change*—a change in temperature due to compression rather than energy content.

Katabatic winds have many different names (in the Mediterranean area they are *mistrals*, in southern California they are S*anta Ana* winds). The Santa Ana Winds are a perfect example. Cool air from the Mojave Desert moves westward, is channeled through canyons where it picks up speed, then moves up the eastern face of the mountains. When it falls down the western face it compresses quickly and heats up by as much as 60°F before blasting and baking the Los Angeles region.

4. Water's Phase Transitions

All objects have three temperature-related phases: gas, liquid, and solid. For water, these phases are water vapor, liquid, and ice. The shift between these phases is largely a function of air temperature and pressure.

<u>Humidity</u>

The ability of air to hold water vapor is called its *vapor capacity* (vc); it is measured in units of weight, e.g. as grams of water per kilogram of air. The air's actual water vapor, also measured in g/kg, is its *vapor supply* (vs), called *absolute humidity*,. Relative humidity (RH) is the ratio of vapor supply to vapor capacity (vs/vc).

Celsius	Fahrenheit	Vapor Capacity
–20° C	–4° F	0.75 g/kg
–10° C	14° F	2 g/kg
0° C	32° F	4 g/kg
10° C	50° F	8 g/kg
20° C	68° F	15 g/kg
30° C	86° F	28 g/kg
40° C	104° F	50 g/kg

Temperature and Vapor Capacity

Vapor capacity is largely a direct function of temperature and (much less so) air pressure. As its temperature increases, the ability of air to hold water vapor increases exponentially (roughly doubling for every 10°C increase). This is shown above. But vapor supply can be treated as simply the amount of water vapor available. Clouds exist where relative humidity is 100% (*saturation*), rain occurs when RH is so high that condensation forms (*supersaturation*) , and deserts exist where there are extremely low relative humidity (*extreme subsaturation*).

Paradoxically, deserts typically have a vapor supply similar to cooler tropical areas, but the vapor capacity is so high that relative humidity is quite low—the water is there, but you don't feel it. The low thermal inertia of desert sand creates extremes in temperature and, therefore, extremes in vapor capacity: the cool air explains the prominence of dew and fog at either dawn or dusk, when vapor capacity is low and relative humidity increases.

Condensation of water vapor—the transition from vapor to liquid—requires a *condensation nucleus*, something to which the water molecule can attach. The condensation nucleus can be a particulate in the air (grit), so rain drops are not "pure." An excellent condensation nucleus is salt, explaining the predominance of haze and fog along seacoasts,

The transition between the vapor and liquid phases of water depends on the balance between condensation and evaporation. Consider the air over a lake. It is constantly taking water molecules from the lake (evaporation) and giving water to the lake (condensation). *Saturation* exists when this exchange is balanced; subsaturation is when evaporation exceeds condensation; and supersaturation exists if condensation exceeds evaporation. Again, this balance depends on the air's vapor capacity, hence on air temperature.

As water transits from one phase to another it releases or absorbs *latent heat*: as it moves from solid (ice) to vapor (gas) through melting and heating, latent heat is acquired; as it moves from vapor to solid by condensation and freezing, it absorbs latent heat. The reason that fog and clouds are warmer than surrounding air is that they contain the latent heat created drawn from the now-cooler surrounding air when water moves from its liquid to vapor form. The increased temperature is one reason that clouds billow upward, an observation that will be discussed when we introduce air buoyancy. The latent heat—and the moisture it represents—also explain the color of clouds; the wavelength of sunlight is affected by the moisture. This is compounded by the particulates which form the nuclei of water droplets.

By the same token, as water vapor condenses it releases latent heat and the air around it becomes warmer, and as water evaporates to become vapor it absorbs latent heat and the air around an area of evaporation gets cooler. As we will see, storms are stimulated by, among other things, the extreme water evaporation and the incorporation of latent heat in the vapor.

Condensation, Evaporation and Latent Heat

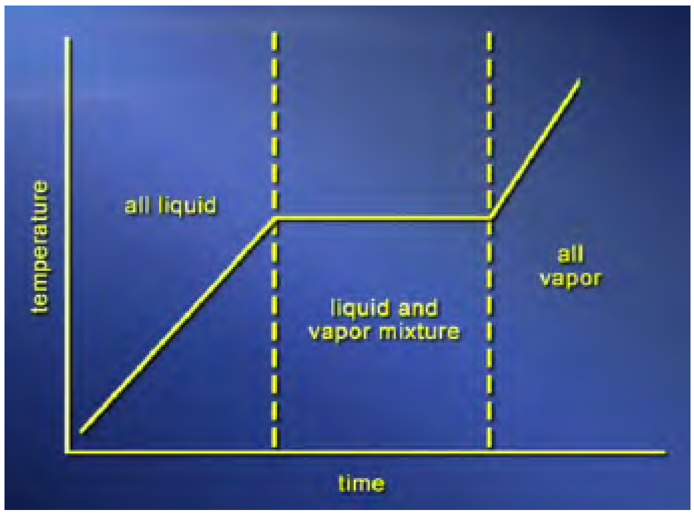
"Temperature" is the result of the motion of atmospheric atoms: when they are energetic and frequently collide, temperature is higher; when they are vibrating less and with fewer collisions , temperature is lower. Thus, the atoms of ice have little motion and ice has an extremely low temperature, while combustible materials can have very high temperatures.

There are three particularly important ways of measuring temperature:

- *Dry Bulb Temperature* (or "regular" temperature) is the temperature that exists when air is subsaturated. One can think of it as dry air (RH < 100%) around the mercury bulb of a common thermometer.
- *Wet Bulb Temperature* is the temperature measured for saturated air under conditions of evaporation. A crude way of measuring wet bulb temperature is to wrap a thermometer's bulb in a moist cloth and blow air on it. Latent heat is absorbed by the evaporating air and drawn away from the thermometer. Hence Wet Bulb Temperature is lower than Dry Bulb Temperature.
- *Dew Point Temperature* is the temperature at which vapor begins to condense, creating fog. Further cooling becomes increasingly difficult because it contracts vapor capacity, creating *supersaturation* as vapor capacity declines. Supersaturation leads to condensation ("dew") and to the release of latent heat that warms the area.

At saturation, all three temperature definitions coincide, but as supersaturation develops due to cooling air, the wet bulb temperature and dew point temperature both decline relative to regular (dry bulb) temperature, the wet bulb temperature declining most. Thus, the lowest "temperature" is the wet bulb temp, the next highest is the dew point temp, and the highest is the regular temp.

Boiling water is at the opposite extreme. The water begins as a liquid and as energy input heats it up its phase shifts into a water/water vapor mix, then finally to a pure vapor form.



Phases of Water Vapor

The shift from pure liquid to a vapor/liquid mix initially creates small bubbles suspended in the liquid; this is due to the opposing forces of atmospheric and water pressure pushing down, and water vapor pressure pushing up. As the boiling point is reached the small bubbles rise to the surface; they break the surface as the vapor pressure from below overcomes the atmospheric/water pressure from above.

The boiling point is the temperature at which the bubbles become long-lived and larger, rising to the surface. Here there is a phase shift as the all-liquid phase becomes a liquid-vapor mixture. Further increases in energy input do not increase the temperature because they are devoted to breaking apart the water molecules in the liquid, increasing the vapor/liquid ratio. Once the water has been entirely converted to vapor, temperature will begin to rise and the vapor will become less dense, expanding as it absorbs latent heat until it overflows or, if capped, explodes like Old Faithful.

The boiling point of water decreases with elevation because the downward atmospheric pressure is lower, as is (by a miniscule amount) the downward force of gravity that reduces water pressure. As a result the bubbling point and boiling point are reached at lower temperatures. Food cooks more slowly at elevation, and coffee is cooler. Both reasons not to climb mountains!

5. Cloud Formation: Air Parcels and Buoyancy

Clouds are *moist air parcels* with internal temperatures greater than the ambient air, allowing them to retain moisture due to their higher vapor capacity. A relatively dry cloud reflects sunlight readily, appearing white. As moisture content increases the cloud absorbs more light and becomes darker.

Cloud Types

Clouds are classified by several characteristics, particularly altitude, structure and subtype. The table below gives some of the cloud types in the taxonomy.

Some Cloud Classifications

Altitude	**Structure**	**Subtype**
Strato- (“Low”)	Stratus (“Flat”)	Lenticaris (“Lense-Like”)
Alto- (“Middle”)	Cumulus (“Heaped”)	Humilis (“Humble”)
Cirro- (“High”)	Nimbus (“Rain”)	Undulatis (“Wavy”)
		Fractus (“Broken”)

Cumulus clouds are vertical formations, often with sizable billows. At the lower layer, a cumulus is *stratocumulus*, looking pillow-like. At the middle level it is *altocumulus* and appears billowy yet broken up, like a “mackerel sky.” The appearance is due to the colder temperature and higher density of the air parcel, tending to smaller and more distinct air parcels. At the highest level a *cirrocumulus* cloud appears thin and wispy, with tendrils coming off of the main body. The cirrocumulus is broken into many small air parcels that look like a sheet of haze with wispy edges

Stratus clouds are spread horizontally rather than vertically. The stratus cloud is a *stratostratus* at the lower level, *altostratus* at the middle level, and *cirrostratus* at the high level.

Stratus and Cumulus clouds tend to be benign, unlike the Nimbus clouds associated with storms. The *Nimbostratus* is a flat dark cloud containing a significant amount of light-absorbing moisture. The *Cumulonimbus* is an anvil-shaped dark cloud piling high into the atmosphere, the "storm cloud."

Within each group there are a variety of subgroups. A *lenticaris* subgroup is a flat oval cloud shaped like a lens; if cumulus it is called *cumulus lenticaris*. The *undulatis* subgroup is "wavy," the *fractus* subgroup is "broken up."

A stratus cloud formation is flat—spread out horizontally over a narrow altitude range. The *humilis* subgroup is a "humble" cloud sharing the common structure of its type: *altocumulus humilis* is a "standard" altocumulus cloud—lumpy and middle-layer. The *undulatis* form is a "rolly" cloud with a wave-like structure; altocumulus undulates will appear as separate clouds like long bedrolls.

Several cloud types are shown below. The stratocumulus nimbus, or "rotor" cloud, is associated with katabatic winds rushing down mountain slopes and pushing back the clouds at the edge of a front.

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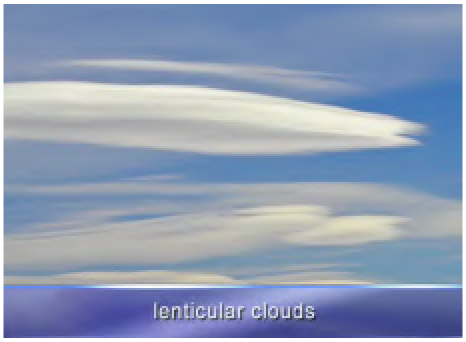
Cumulus ("Pillow")



Cumulus Lenticaris ("Lens")

Cumulonimbus (“Anvil”)

Altocumulus Lenticaris (“Rotor”)

Altocumulus Undulatis (“Band”)

Stratocumulus

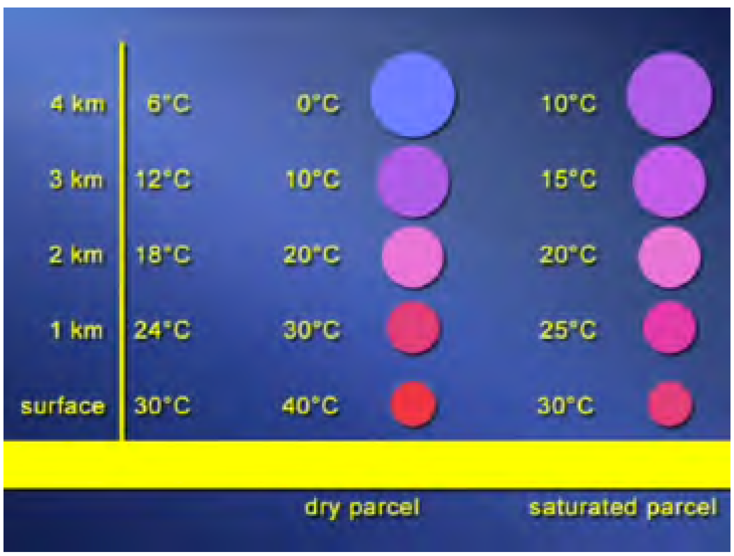
Altocumulus

Altocumulus (“Mackerel Sky”)

Air Parcels and Air Buoyancy

As noted above, clouds are air parcels with different internal characteristics than the ambient environment. To understand an air parcel, we begin with a *dry air* *parcel*: it is assumed to be subsaturated (RH < 100%), to have the same internal pressure as the surrounding air, to be closed (allowing no heat exchange with its environment), and to have a different temperature than surrounding air. An example is a release of hot air from an industrial process. A *moist air parcel* differs only in that its vapor content is at saturation, i.e. RH = 100%.

The dry air parcel has a higher internal temperature than the surrounding air and, therefore, tends to expand and become less dense. This causes it to rise, and as it rises it cools off but (as shown below) at a slower rate than the surrounding air because compression increases RH and releases latent heat (the dry adiabatic process).



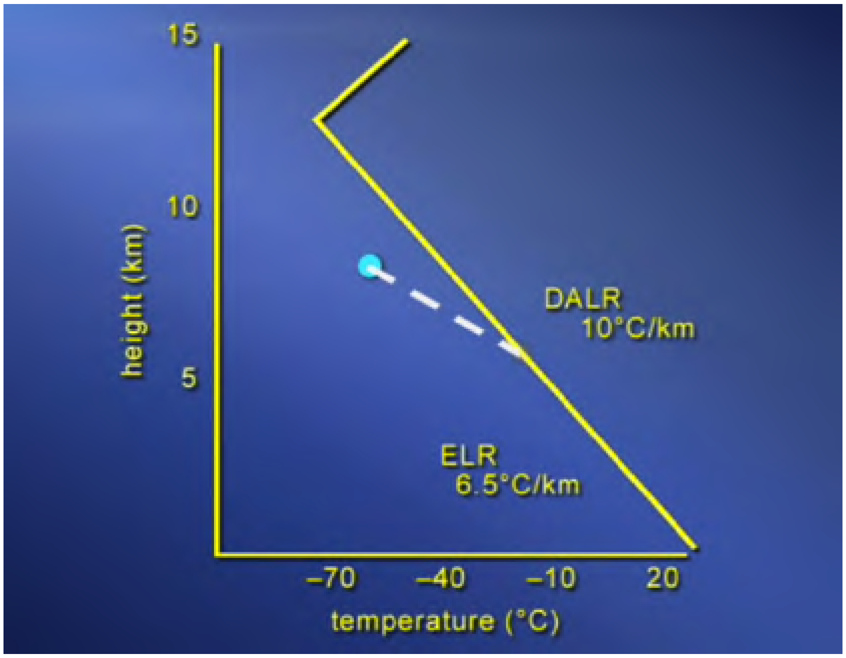
Internal Temperature of Dry and Moist Air Parcels

A moist air parcel with the same internal temperature has, by definition, a higher initial relative humidity. As it rises and compresses it becomes supersaturated, undergoing a phase transition from vapor to liquid. This releases more latent heat than the dry parcel. As a result, the moist parcel maintains a higher temperature as it rises. This gives it greater buoyancy than a dry parcel so it loses heat more slowly and rises farther.

The loss of temperature as altitude increases is called the *Lapse Rate*. The *Dry Adiabatic Lapse Rate* (DALR) is the change in the dry parcel's internal temperature per kilometer of altitude due to air compression or expansion—the

DALR of a dry air parcel in the troposphere is about 10°C/km and, as shown below, the dry parcel's internal temperature changes more than the ambient temperature as it rises. The Lapse Rate of a moist air parcel, the Moist Adiabatic *Lapse Rate* (MALR) is about 5°C/km, half of the DALR. This is because the moist parcel generates more internal heat so as it rises its temperature falls less than the dry parcel.

In short, a dry air parcel has less buoyancy and reaches equilibrium (equalized internal and external temperatures) at a lower altitude than a moist air parcel.



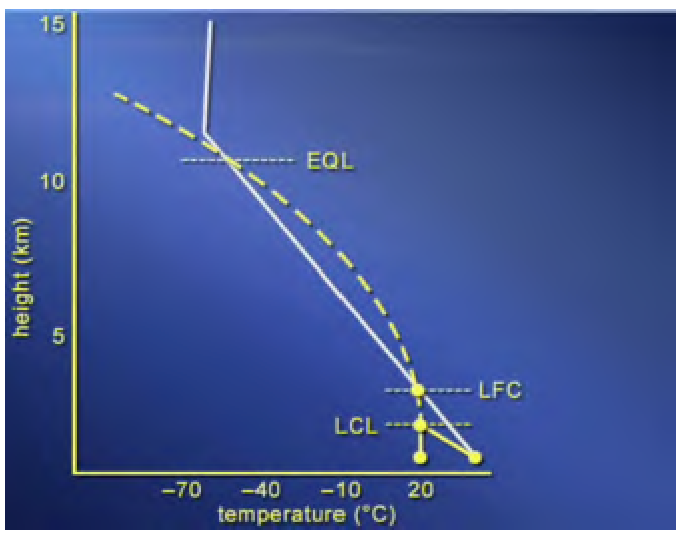
Dry Adiabatic and Environmental Lapse Rates

Yet another lapse rate definition is the *Environmental Lapse Rate* (ELR). This is the actual temperature-altitude relationship for the ambient air. Because the ambient air consists of both moist and dry air parcels, the ELR is about 6.5°C/km, between the DALR and the MALR.

Cloud Formation

Cloud formation has several sources: first, saturation of the air as altitude increases and vapor capacity declines (the adiabatic effect of compression at colder temperatures); second, a dew point lapse rate of 2°C that is far below the DALR, creating supersaturation at a low altitude; and third, the influence of wind on evaporation and cloud lift..

The figure below shows these sources. The straight white line shows the adiabatic lapse rate of a dty air parcel (DALR); its intersection with the temperature axis at about 25°C is the surface temperature. The solid near-vertical line at the lower right is the dew point lapse rate (DPLR); its intersection with the temperature axis is the dew point, so the surface dew point is 20°C. The curved dashed yellow line is the moist adiabatic lapse rate (MALR).

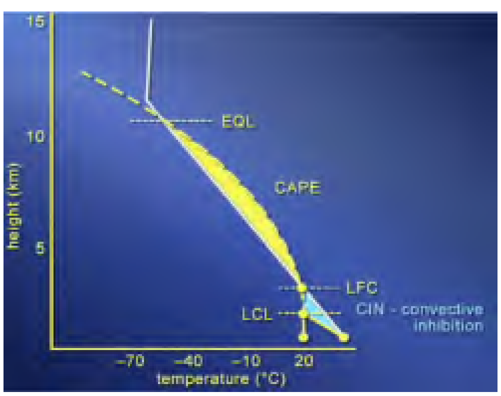


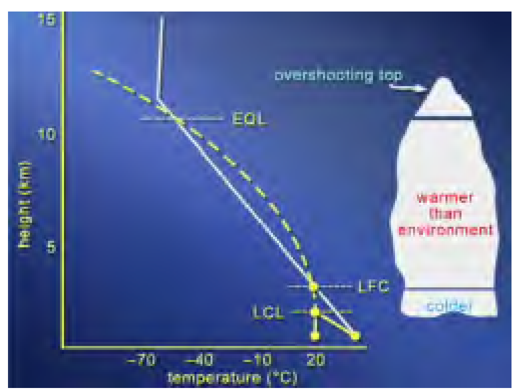
Let's start at the surface. The dew point is below the actual temperature so the air parcel is subsaturated. As altitude increases the moist air parcel's temperature drops faster than the dew point; at an altitude labeled LCL (for *Lifting Condensation Level*) the dew point and actual temperatures are equal and the parcel becomes saturated. The LCL is also called the Cloud Base because it is the altitude at which enough moisture is in the parcel to begin cloud development.

At the LCL the air parcel has negative buoyancy—it is cool and will fall if left alone. But if an external force (wind) pushes the air parcel up enough to reach the LFC (*Level of Free Convection*) the air parcel develops positive buoyancy and begins to rise on its own. This occurs until the EQL (Equilibrium) level is reached, when negative buoyance reemerges. EQL is the top of the cloud.

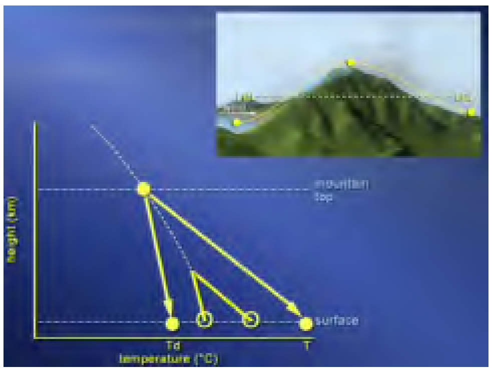
Note the area between the solid line and the curved dashed yellow line between LFC and EQL. That is the total area for which the parcel's temperature exceeds ambient air temperature. That area represents the accumulation of *potential energy* in the form of latent heat embedded in water vapor. If a cold front comes along and condensation occurs, the vapor can turn to liquid (rain), releasing the latent heat. This is what drives storms.

This potential energy is shown more clearly in the left figure below. *Convective Available Potential Energy* (CAPE) is the accumulated latent heat in the altitude range of positive air buoyancy. The area shaded blue below the LFC is the area of negative buoyancy and loss of potential energy. The right graph shows the vertical cloud layer associated with the left chart.

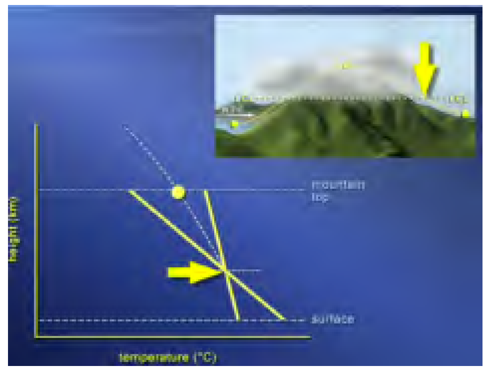




The figures below show development of a cloud formation around a mountaintop. The left figure is the cloud developing on the windward side of the mountain, as wind pushes moist air up to the LFC forming a cloud base near the mountaintop. Note the large and small triangles; these represent two different pairs of air temperature and dew point. The large triangle shows a temperature and dew point range in effect for the air conditions; the smaller triangle shows the temperatures for a situation that would produce clouds below the mountaintop. This represents the importance of dew point-ambient temperature differential to the height of the cloud base.



Windward Side



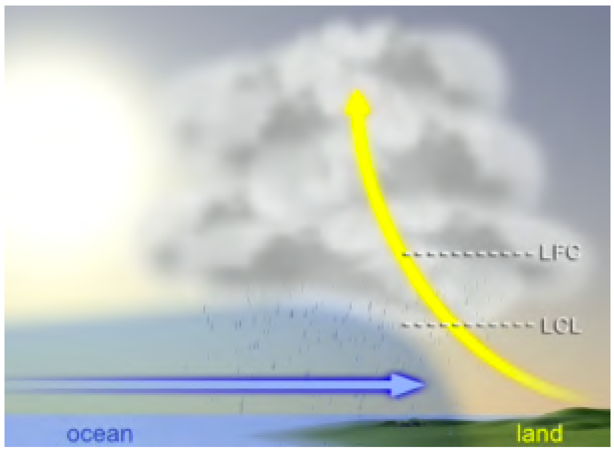
Leeward Side

The right figure shows the action on the leeward side. The air passes over the top of the mountain and falls down the lee side. As it falls it cools, its temperature and water capacity decline, and it becomes subsaturated as it drops below the LCL. This forms the cloud base on the lee side.[1] Thus, all of the moisture (and cloud) builds at the top of the mountain.

[1] Note that this not a source of katabatic wind—not all wind passing over a mountain top becomes katabatic—katabaticity depends on specific conditions creating high turbulence.

Sea and Land Breezes

The figure below shows a Sea Breeze Front. A Cold front arriving from offshore encounters warm air over land. The warm air is less dense and rises over the incoming cold air. The onshore breeze (land to sea) develops as the sea air is pulled in to replace the rising warm air.



Sea Breeze Circulation

The rising warm air creates a cloud base at the LCL and the warm front's wind pushes the moist air up to the LFC, where it becomes buoyant and a vertical cloud develops. The air at the higher elevations condenses, becoming supersaturated and creating rain. As part of the process, a sea breeze circulation builds in which air near the surface flows toward land and air at high altitudes flows toward the sea. The circulation continues until the cold front passes through.

Thus, cloud development and cloud height depend on the relative dew point and air temperature at the surface, which set the LCL or cloud base,, and on the wind that pushes saturated air to the LFC at which it becomes saturated and buoyant, rising to higher altitudes where it becomes condenses into rain.

6. Regional Wind Patterns

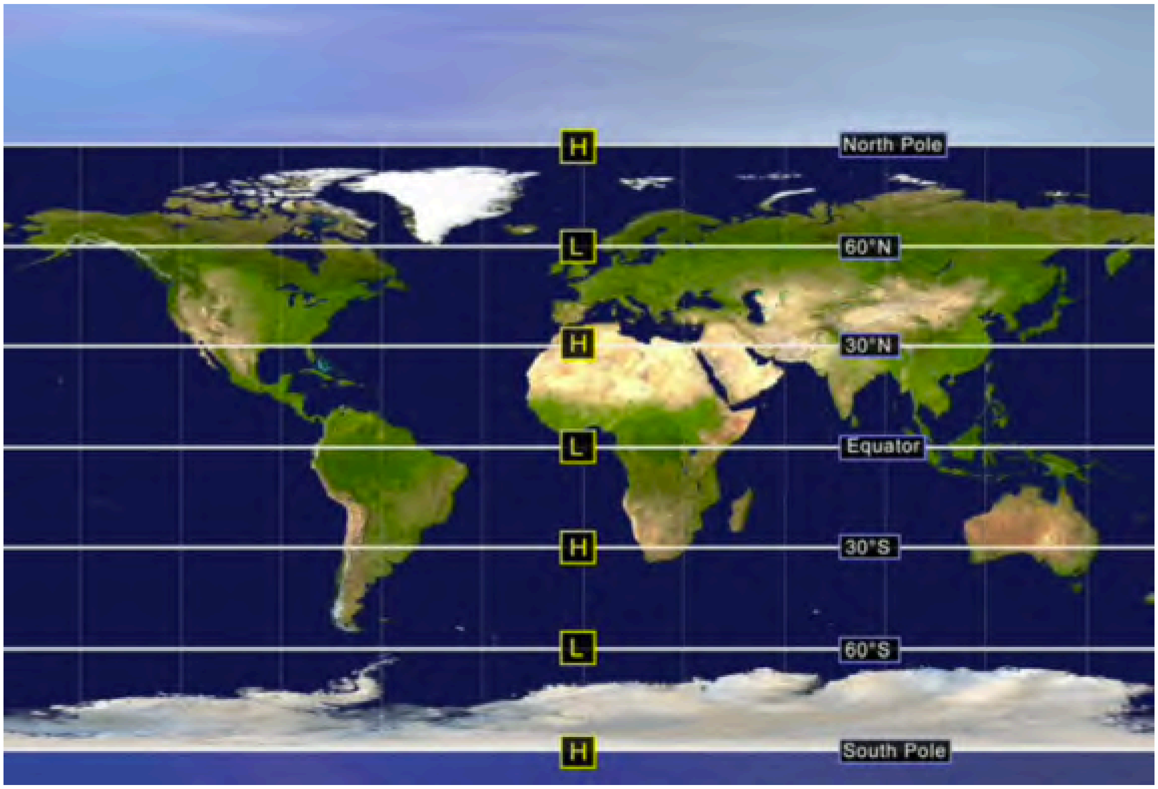
As we have seen, convection (winis created by pressure differentials) causes air to flow from areas of high pressure to areas of lower pressure. But this can't be the whole story of wind direction; if it were there would be a steady wind from the cold, high pressure over the Poles toward the Equator. Instead we see a normal southeast trade wind over the eastern U. S. and a northwest trade wind over Canada,

There are four forces driving the regional wind patterns: the Pressure Gradient Force (PGF) due to regional differences in air pressure; the Coriolis Force due to the Earth's eastward rotation; Frictional Forces at the surface where wind and land meet; and the Centripetal Force associated with rotation of air masses.

<u>The Pressure Gradient Force</u>

The *Pressure Gradient Force* is the air pressure difference *per kilometer* of distance between the points of measurement. Consider a pressure difference of 500 millibars. If this occurs between points 1000 miles apart, the PGF will be only one-one thousandth of the PGF between points one mile apart. Because PGF drives *wind velocity*, the first case would be a barely noticed breeze but the second case could be a hurricane-force wind.

The Earth is composed of several bands of high and low pressure, seen below. A low-pressure band at the equator rises to high pressure at about 30° latitude; this is followed by a decrease in pressure to another low pressure band at about 60° latitude. Finally, pressure rises to high at the poles. The same bands are found in the southern hemisphere.



Global Pressure Bands

Thus, the PGF alone would create northerly winds at low latitudes, southerly winds at mid latitudes, and southerly winds again at high latitudes; in the southern hemisphere the wind directions would be reversed. The PGF explains why warm fronts tend to arrive at the mid-latitude U. S. from the south while cold fronts tend to arrive from the north.

But the normal wind direction in the U. S. is not south to north, it is southwest to northeast. To explain this we need to introduce the *Coriolis Effect*.

<u>The Coriolis "Force"</u>

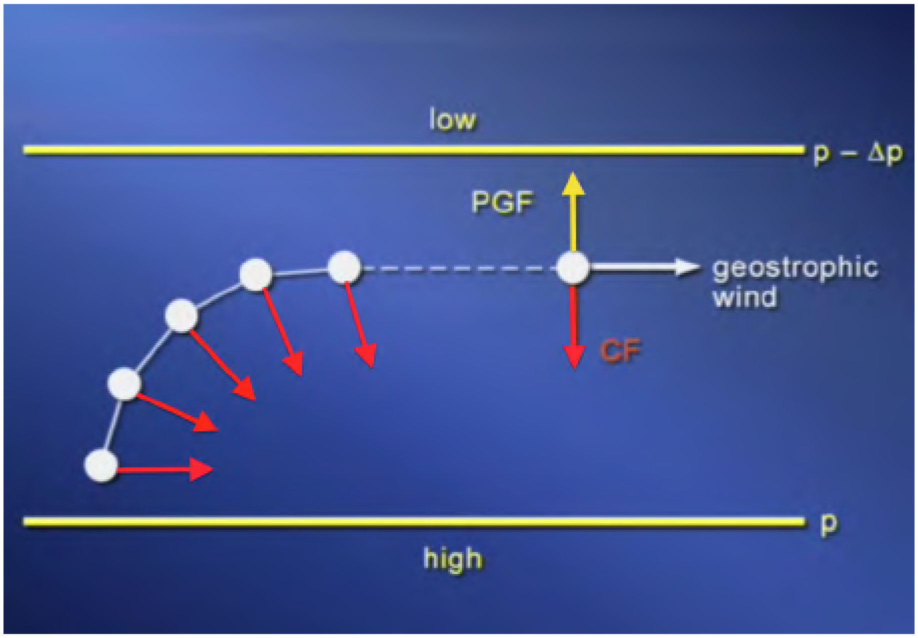
The *Coriolis Force* is often called the Coriolis Effect because it is not really a force like gravity, electricity, or magnetism. Rather, it is a consequence of the Earth's rotation. In the northern hemisphere Earth's rotation is counterclockwise as seen from above the North Pole; in the southern hemisphere the rotation is clockwise as seen from above the South Pole.

Let's focus on the U. S., where the Earth is rotating toward the east. The velocity of rotation is about 1,000 miles per hour near the equator, but it is extremely slow one mile away from the north Pole. The reason is, of course, that a point on the Equator (say, Nairobi) moves 1,000 miles in space in each 24-hour period, while an igloo a mile away from the North Pole moves only about 6.2 miles in the same period.

The effect of this differential angular velocity is that an aircraft departing on a fixed path due north from Nairobi to Chicago. The plane is moving eastward at 1,000mph and northward at (say) 500mph when it leaves Nairobi. Because Chicago is at a latitude with lower eastward rotational velocity, the plane will be drifting to the east of Chicago as it heads due north. That is, it will be moving to the northeast relative to Earth's surface even though its compass reading remains 000°.

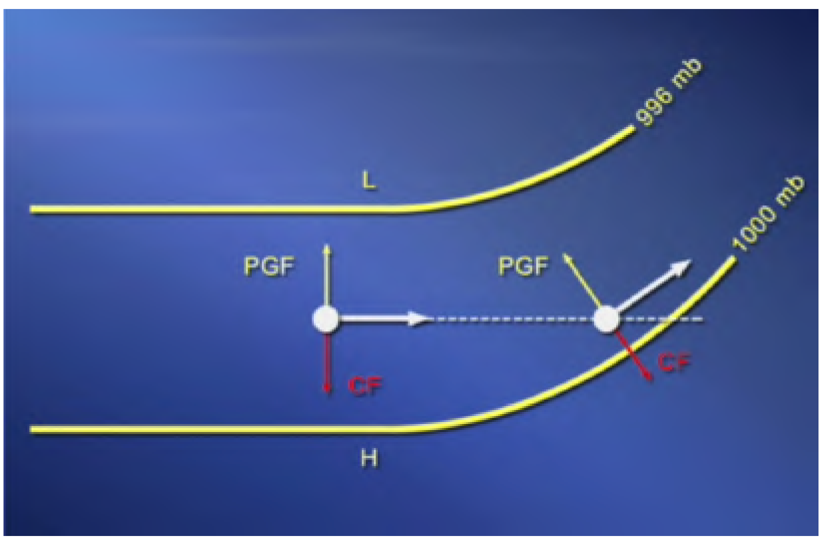
That easterly push is the Coriolis Effect, pushing objects traveling northward toward the east and objects traveling southward toward the west. Over the U. S. air would travel northward (be a southerly flow) if the PGF were the whole story, but the Coriolis Effect means that air traveling north will be a southwesterly (travel toward the northeast), and air traveling southward from upper Canada will be a northwesterly.

This is shown below. Two pressure levels are shown: p and a lower pressure p - Δp. An air parcel begins at the lower left, where it is moved north by the pressure gradient but simultaneously moved east by the Coriolis effect (red arrow), causing it to drift toward the north-northeast. As it moves the PGF remains due north, but the Coriolis force weakens because the air is moving toward a region with lower rotational velocity. This continues with the air parcel arcing to the east and the Coriolis Force weakening, until the north-pushing PGF is exactly offset by a south-pushing Coriolis Force. At that point the air is in *geostrophic balance*, heading due east along a constant isobar (line of equal pressure).



Pressure Gradient and Coriolis Forces, Northern Hemisphere

But, in fact, isobars are not parallel as in the example just show. They are typically curved into a circular form by Highs and Lows. This modifies the analysis a bit. Because isobars are circular rather than straight lines, the wind moving along isobars has a circular flow, as shown below where there is a counterclockwise flow around a low pressure trough.



Wind Direction and Curved Isobars

If the wind continued straight it would move from lower to higher isobars, something wind does not want to do. But as it approached the higher isobar the direction of the PGF shifts toward the northwest because the PGF is always perpendicular to the isobars, and wind flows along isobars. The Coriolis force shifts toward the southeast, and the net effect is to tilt the wind direction toward the northeast, keeping it along the same isobar.

But there is more to the story. In fact, the wind does cross isobars: in the northern hemisphere it is attracted toward low pressure, not equal pressure as the PGF and Coriolis Forces predict. This means that a southwesterly wind around a low pressure area (a Low) is directed more toward the Low's center. Around a High it is directed away from the High's center. To explain this we need two additional forces: the *Frictional Force* and the *Centripetal Force*.

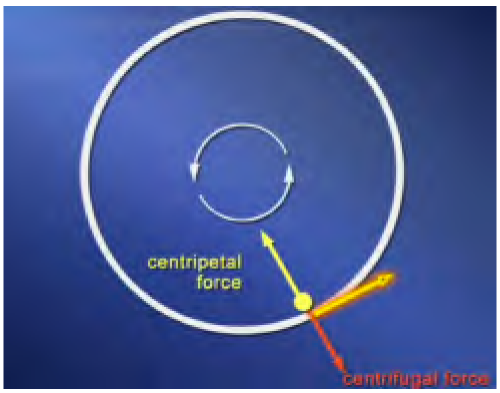
The Frictional Force

The frictional force causes the wind along the Earth's surface to slow down. This tilts the wind toward the center of the Low; it no longer is directed along isobars, it tilts toward the northwest because that is the low pressure trough. This combination of PGF, Coriolis Force, and friction tends to cause a Low Pressure area to spin counterclockwise in the northern hemisphere (clockwise in the southern hemisphere) as the wind veers toward the Low's center. This is enhanced by a fourth force—the centripetal force—that enhances the tilt toward low pressure.

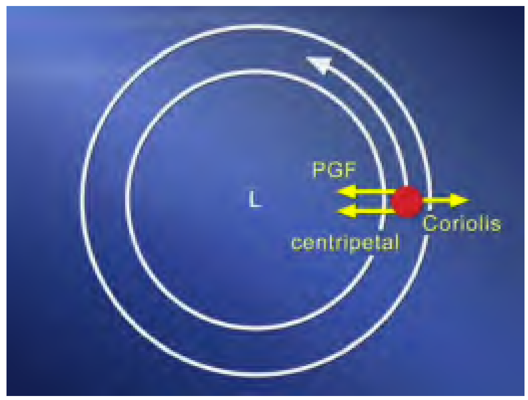
The Centripetal Force

The Centripetal Force is the inward force created by a spinning object, like air around a Low or a High. The centripetal force is a real force that directs spinning objects inward; it is often confused with the fictitious centrifugal force that is said to push spinning objects outward. In fact, the centripetal force and the "centrifugal force" always exactly offsetting, so any problem in physics can be addressed using either force. This is shown in the left chart below. The actual direction of travel is tangent to the radius of the circle; that direction can be deduced as the force arising from the balance of the centripetal force and the “centrifugal force.” However, the “real” force is centripetal—the centrifugal force is the centripetal force in disguise.

We have seen that the first three forces cause the air circulation to follow isobars but with a tilt toward the Low’s center that causes a counterclockwise spin. (around a High the circulation is clockwise with a friction creating a tilt toward the center and a clockwise spin). The centripetal force adds a second inward-directed force to the Low Pressure Area, as seen in the right figure below.

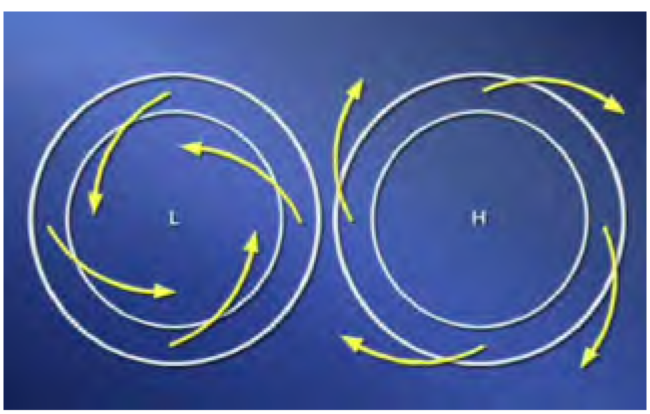


Centripetal and Centrifugal Forces



The Forces against Geostrophic Wind

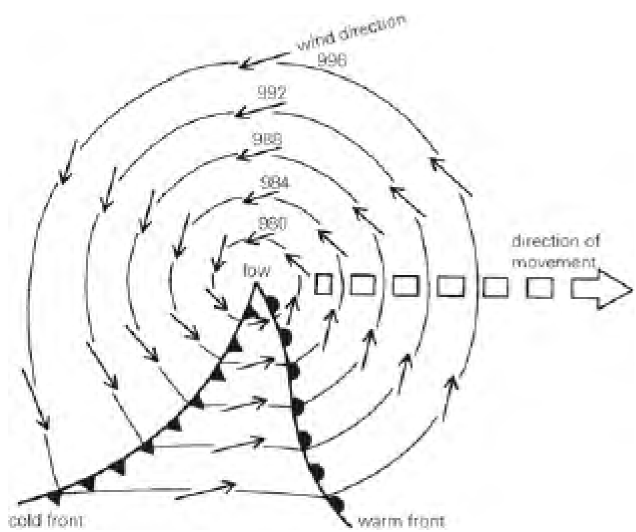
The PGF and Coriolis forces are offsetting when the wind is geostrophic, that is, following an isobar, so the addition of the centripetal effect tilts the wind even more toward the center of a Low. In fact, the wind is tilted about 30° toward the center (30° away from center for a High).



Cyclonic Winds **Anticyclonic Winds**

The counterclockwise rotation of a Low is called a *Cyclonic Wind*, while the clockwise rotation of a high is an *Anticyclonic* Wind. This has no relationship with the term "cyclone" for hurricane-force winds except that hurricanes do rotate counterclockwise. So when you are told that a cyclone is coming, you needn't panic until you know its wind speeds.

A typical wind-pressure chart for a Low is shown below. The Low is formed by the collision of a cold front from the west with a warm front from the east The chart is analogous with topographical terrain maps: isobars on weather maps replace the altitude levels of terrain maps. The height of an isobar is the constant air pressure it represents (the chart shows isobars ranging from 996mb down to below 980mb). The distance between isobars (analogous to the slope of terrain) shows the PGF: the closer are two isobars the greater is the PGF and the greater the wind velocity. Wind direction in the counterclockwise circulation is along the isobars with a roughly 30 orientation toward the Low's center.



Wind-Pressure Chart

Of course, the opposite exists in a northern hemisphere High: the circulation is clockwise, the wind velocity will be higher than in a Low because the wind speeds up to move away from the High's center, and the wind direction is tilted toward outer (lower pressure) isobars.

We have discovered the basis of *Buys Ballot's Law* (pronounced *Bwah Ballo*). In 1857 the Dutch meteorologist claimed (almost correctly) that if you stand with your back to the wind and your arms outstretched to right and left, your left arm will point to the Low and your right arm to the High. If the low is to your west, it is coming toward you and you are on its "dangerous side." Your best bet is to move toward the High.

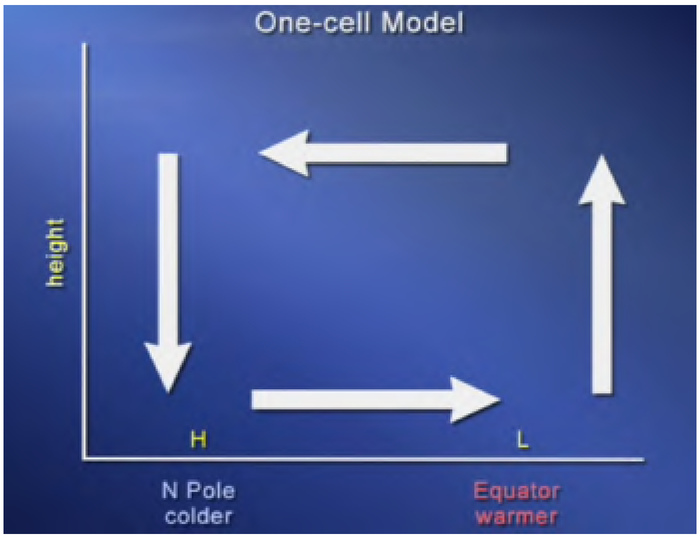
This is almost correct because it assumes that isobars are concentric circles and that wind moves tangent to isobars. The first is an approximation, and the second is true at higher altitudes.. But at the surface, friction turns the wind about 30° toward the Low. Thus, your modified Buys Ballot Law is to stand with your back to the wind, turn about 30° to your right. You will then be facing tangent to an isobar. Now extend your right and left arms outward. As before, the Low is on your lefts and the High is on your right.

7. Global Wind Circulation

The previous section addressed regional and local wind circulation based on four forces: the Pressure Gradient Force, the Coriolis Force, Friction, and the Centripetal Force. The PGF and Coriolis Force determine the geostrophic winds—winds on a large scale that follow along isobars. Friction and the Centripetal Force affect the local and regional wind patterns.

Global Geostrophic Winds

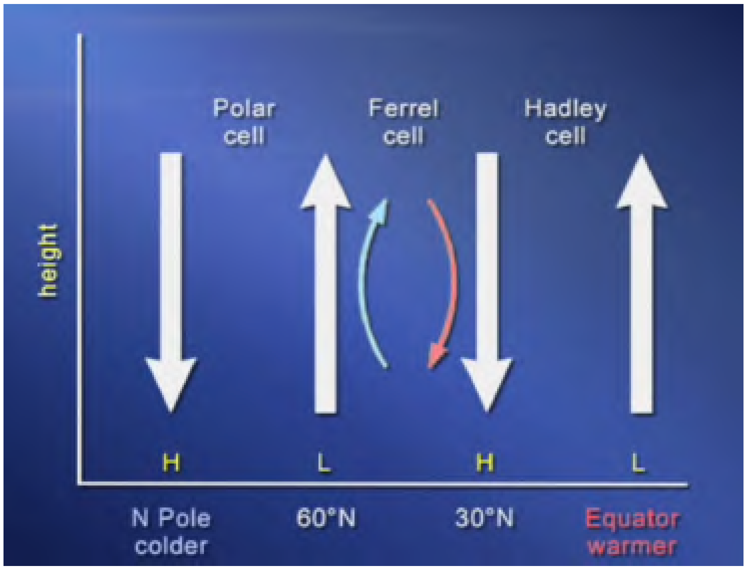
Global winds are primarily geostrophic. The analysis of them can begin with the One Cell Model, in which each hemisphere has a single air circulation pattern. This model, shown below, is a replica of the Land Breeze/Sea Breeze distinction made in earlier sections.



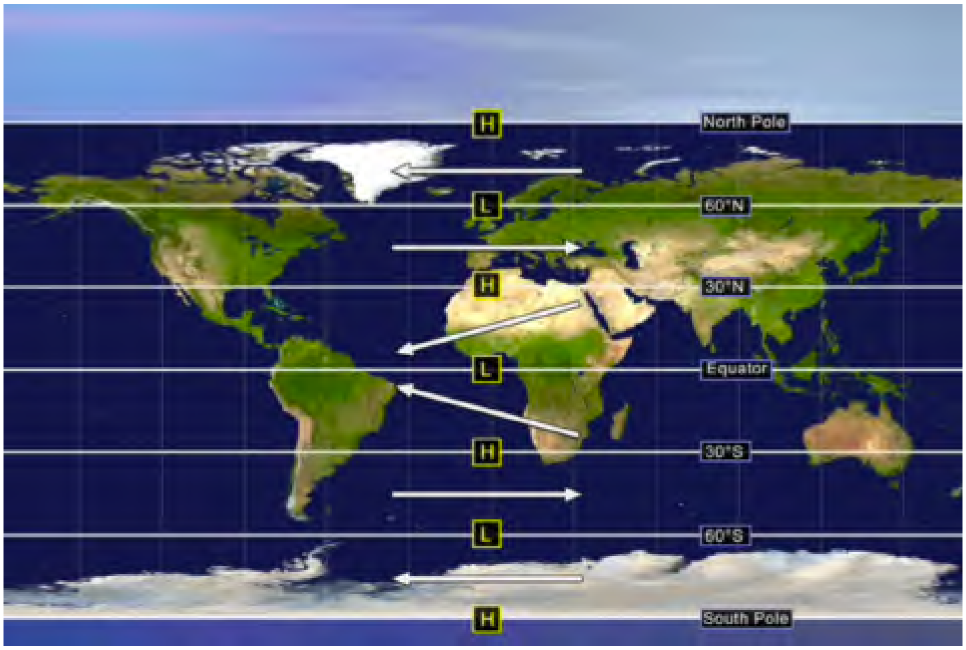
The One Cell Model

The One Cell Model has a high-pressure over the North Pole and low pressure over the Equator. At the surface the air follows the pressure gradient from high to low, where its temperature increases and it becomes buoyant and rises. At some altitude it compresses, maintains altitude, and flows back toward the North Pole. This model predicts that winds travel from north to south at the surface, and from south to north at altitude.

A more accurate model is the Three Cell Model. This recognizes that there are several pressure zones in each hemisphere. In the northern hemisphere air pressure is low at the Equator and rises to become high at about 30° latitude. Then it falls until 60° latitude, finally rising again in the north latitudes. This creates three pressure cells: the *Polar cell* in the northern 30° of latitude, the *Ferrell cell* in the middle latitudes, and the *Hadley cell* from 30° latitude to the Equator.



The Three Cell Model, Northern Hemisphere



Global Air Circulation

As shown in the chart just above, in the middle latitudes the geostrophic wind is westerly (from west to east), while at the lower northern latitudes it is easterly but with a significant tilt to the south.[2] Of course, the opposite flows exist in the southern hemisphere.

Note the wind flows in the lower latitudes of each hemisphere. The pronounced Equatorial Low over Brazil attracts the winds away from their geostrophic path. This low is created by two abutting lows, and gives rise to fierce storms are noted on the northeast coast of Africa.

Note also the alternation of green vegetation and arid desert, particularly extending northeast from the Sahara to the Gobi deserts. This is a band of dense cool air that is dry due to its loss of water vapor. It begins over the Gobi and is driven southwest over the Arabian and Saharan deserts. It is ironic that these deserts are areas of cool air but are extremely hot at the surface. The heat is direct heat from solar radiation.

The winds shown above are geostrophic winds due to the Pressure Gradient and Coriolis forces. In addition there is considerable regional variation from surface friction and the centripetal force associated with Earth's rotation. This is shown below in a summertime chart of flows over the U. S.

The purple areas over the Pacific and Atlantic are regional high pressure areas: the Pacific High and the Bermuda High. The clockwise rotation of these highs alters the wind patterns significantly. Over the Gulf of Mexico the prevailing winds are easterlies, as predicted by the Three Cell Model. But over Texas and northern Florida the Bermuda High pivots those easterlies to become southwesterlies.

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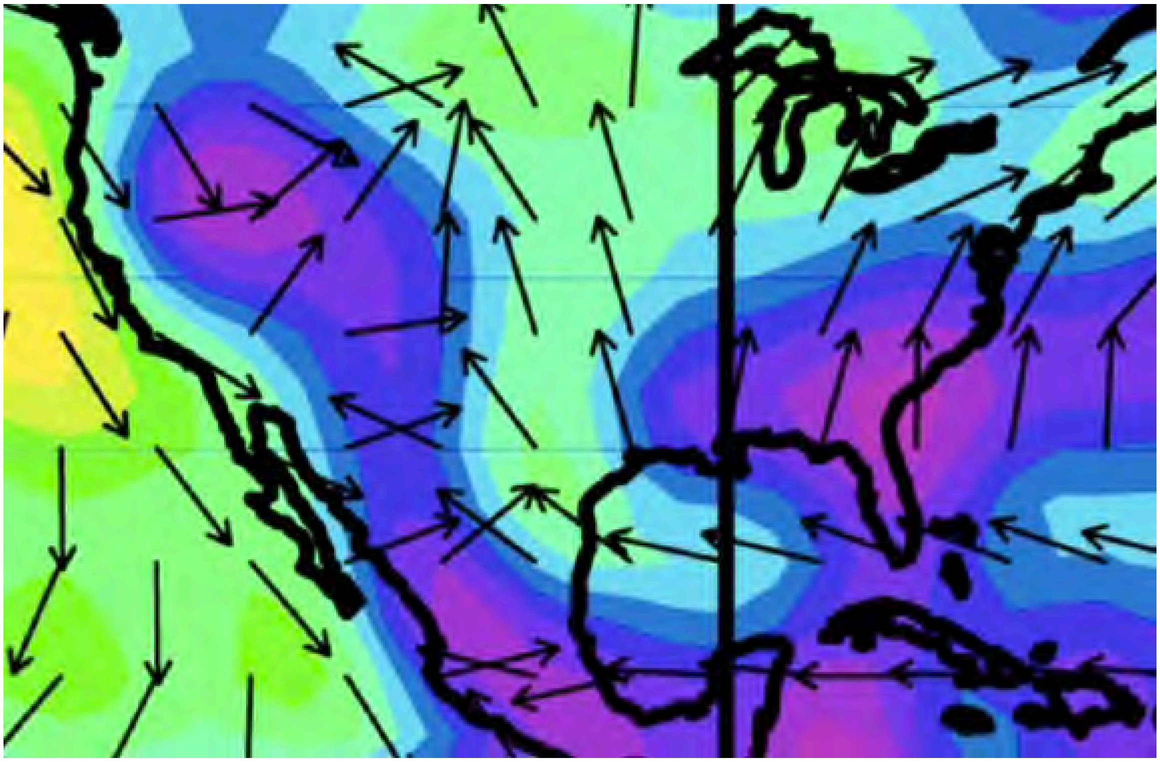
[2] The pressure gradients that determine wind seed are North-South, but the isobars that determine wind direction are perpendicular to the pressure gradients, East-West.



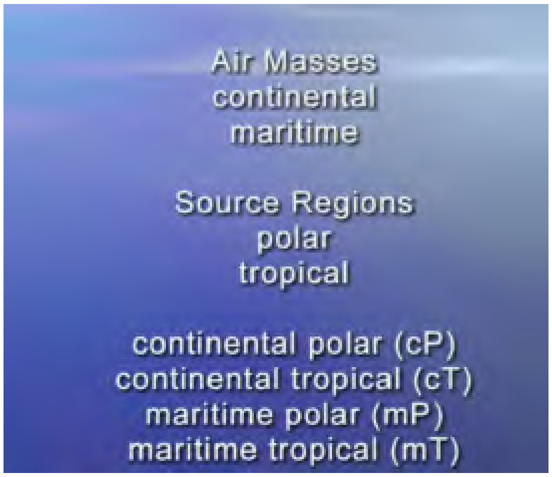
Prevailing Winds—United States, Summer

The prevailing winds on the West coast are more complicated. The arrive from the northwest (a combination of the northerly from Canada via the PGF and the westerly from the Coriolis force. Then they become caught in the Pacific High over California and rotate to become southwesterly, then turn to southerlies as they collide with the Rocky Mountains.

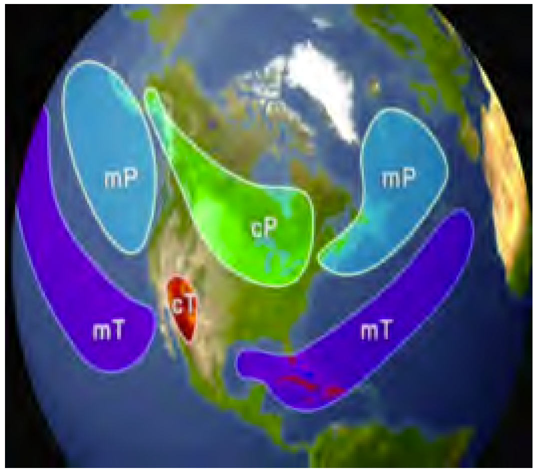
Air Masses

An *air mass* is a large air parcel carrying a common density and temperature. Air masses are the source of *fronts* that drive the harsher weather patterns. As noted before, air masses of different densities do not easily mix—air, like people, seeks its own kind.

The taxonomy of air masses is shown below. They are classified by whether they are over land or water (continental and maritime), and whether they are cold and dense (polar) or warm and light (tropical).



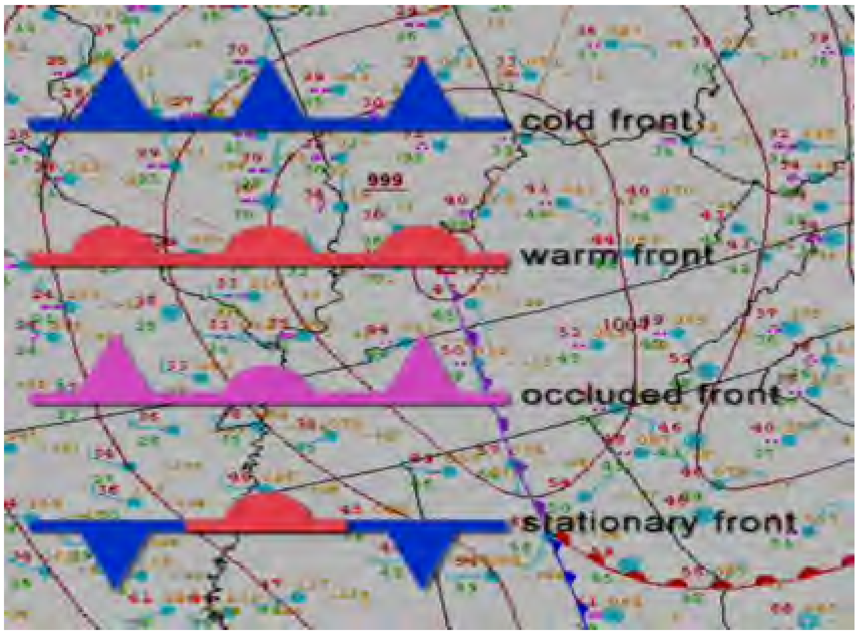
Taxonomy of Air Masses



U. S. Air Masses

8. Fronts and Low Pressure Air Masses

A front is an air mass with a specific air density. Fronts come in four basic forms, shown below



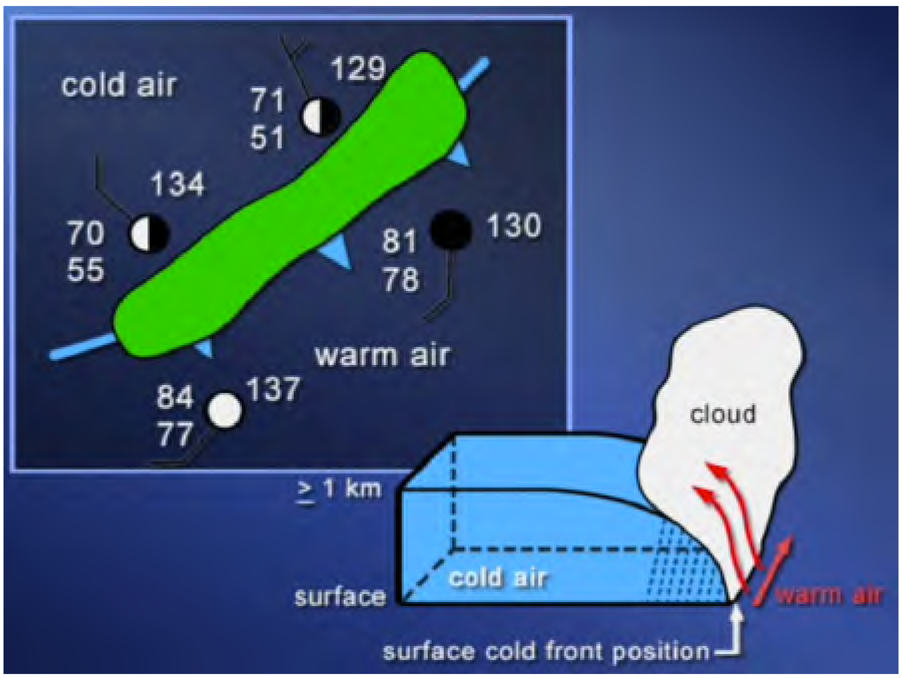
Types of Fronts

A Cold Front is a mass of dense cold air represented by a blue line with triangles in the direction of motion. Warm Fronts are warm masses shown by red lines with semicircles giving the direction. A Stationary Front is a front with warm and cold air masses joined but moving in opposite directions, thus each blocking the other's motion. An Occluded Front has both Warm and Cold Fronts joining as fa aster-moving Cold Front overtakes a Warm Front, both moving in the same direction once joined. Occluded fronts occur near the centers of Lows and usually occur near the breakup of storms.

The collision of warm and cold fronts is a reciprocal event, so when is a front called warm or cold? The answer is that the most energetic and fast-moving air mass gets the nod. Because cold air moves more rapidly, fronts in collision are typically called cold fronts.

Cold and Warm Fronts

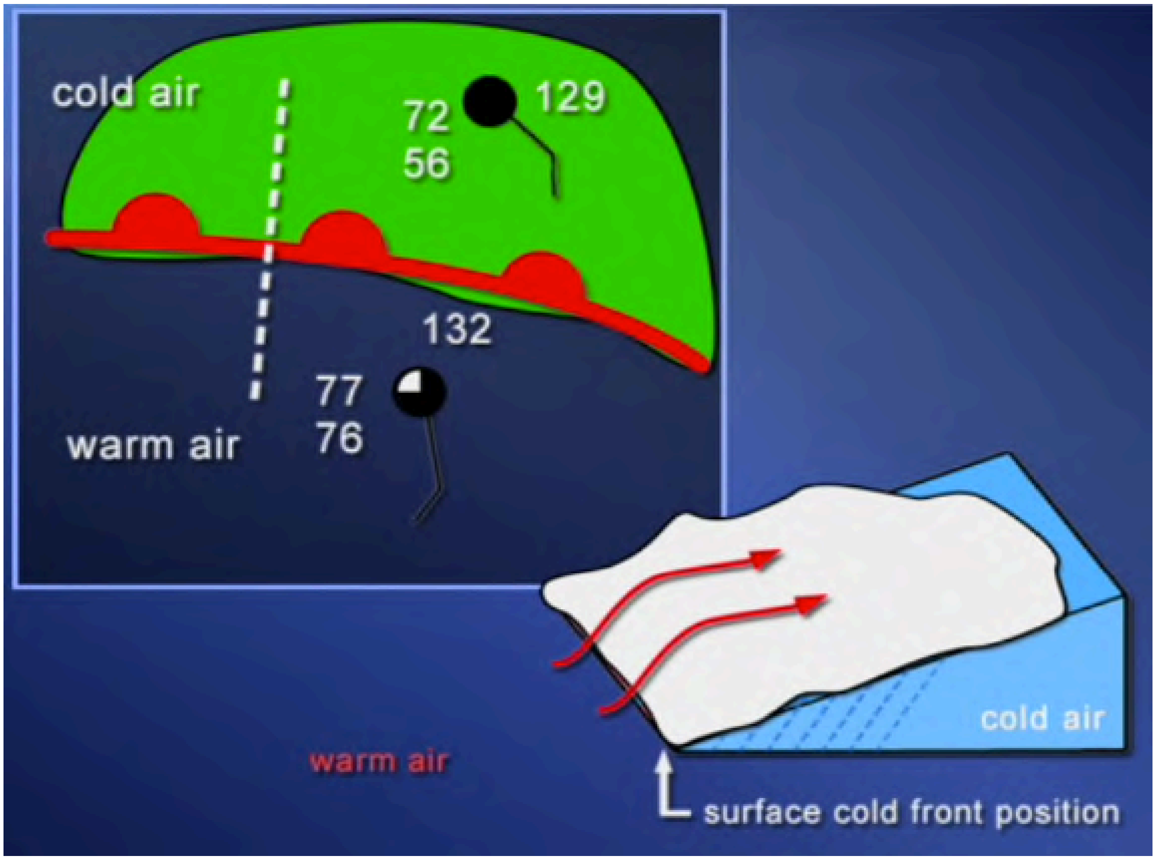
A Cold Front is shown below. Dense cool air is on the march toward warm air. The symbols shown on the chart represent the meteorological conditions. We will go through these later, but this chart shows that at the right side of the cold front (facing in its direction of motion) the air temperature is 70° with dew point 55°, the sky is half obscured by clouds, and the air pressure is 1013.4mb; the facing warm air has temperature 84°, dew point 77°, clear skies, and pressure of 1013.7mb.



A Cold Front Colliding with a Warm Air

A rain band has developed along the junction of the cold and warm air. The warm air, being less dense, rises over the cold front forming a cloud as the warm air's temperature drops to the dew point so condensation of water vapor creates the rain.

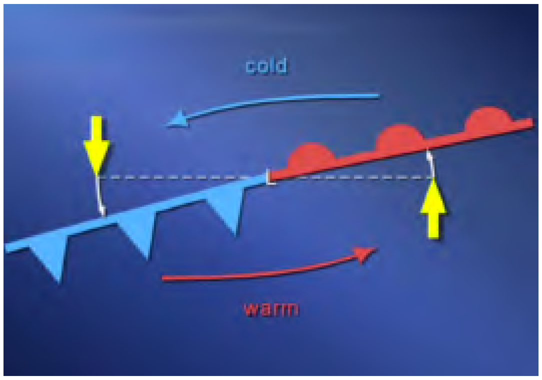
Another form of cold and warm front collision is shown below. In this case the warm front is on the move and the cold front has less momentum and is not as well formed. The warm air slides over the cold air creating a low cloud stratum rather than a vertically formed rain cloud. Rain develops, but it comes from a thinner layer of air and is less heavy than the first situation.

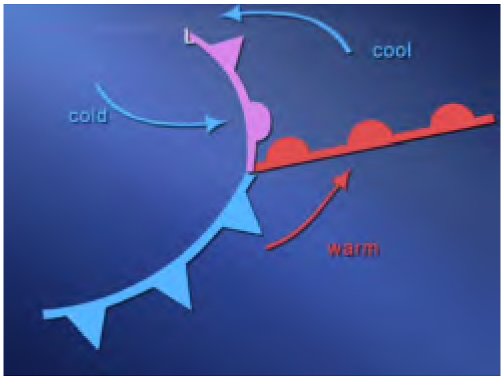


Warm Air Front Encountering a Cold Front

Occluded Fronts

An occluded front occurs when a cold air front encounters a warm front, as in the left figure below. The junction is a low pressure area. The colder air behind the cold front and ahead of the warm front flows toward the cold front, while the warm air on the other side flows in the opposite direction. The result is that the faster moving cold front begins to pivot around the low, as in the right figure.



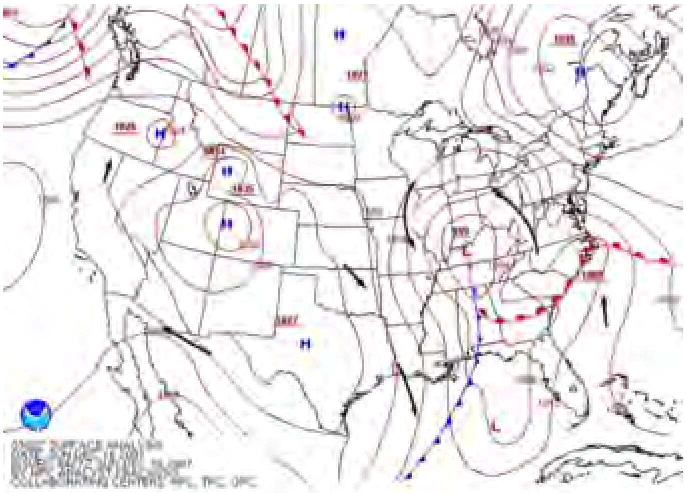


Occluded Front, Phase 1 **Occluded Front, Phase 2**

The occluded front, marked pink with triangles and semicircles, is a mixture of the two fronts. Because cold dense air and warm light air don't mix, the occlusion will not last long. But it exacerbates the counterclockwise flow around the Low.

Weather Charts

Perhaps the most common weather chart is the *surface chart*, shown below.

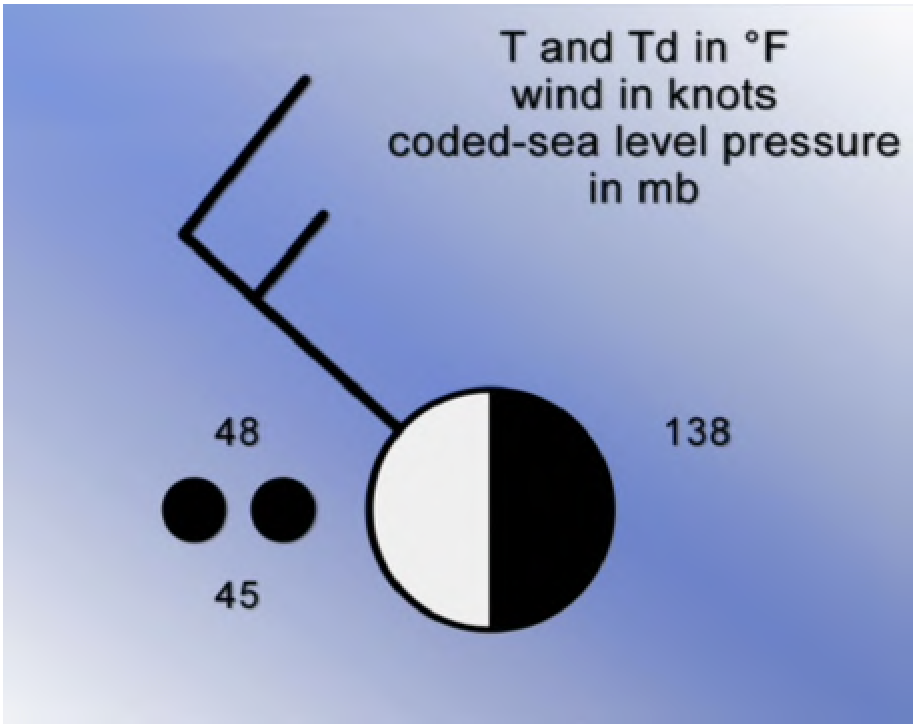


Surface Chart, U. S., September 7, 2007

The surface chart shows the isobars and the associated air pressures in millibars, calibrated to sea level to eliminate the effects of terrain.[3] Over the Western U.S. a large High has developed with its clockwise (anticyclonic) airflow. The fairly wide distances between isobars indicate light winds.

Over the Midwest a Low has developed as cold and warm air fronts collide and create an occluded front beginning at their junction and heading to the Low's center. The circulation is counterclockwise (cyclonic) and the winds are strong as shown by the narrow gaps between isobars. Another occluded front is developing in the Pacific in the northwest, and a warm front is moving eastward in western Canada and the northern U. S.

As seen above, surface charts can also be supplemented with symbols representing temperature, precipitation, cloud cover, pressure, and wind speed. We encountered some of those symbols in the discussion of warm and cold fronts.

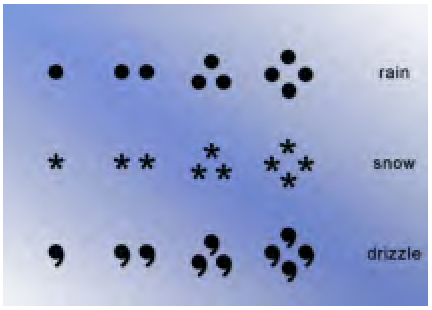


[3] Air pressure is lower at higher altitudes, and pressure sensing stations are at various altitudes above sea level. A surface chart showing measured "station pressure" would conflate the effect of altitude with the air pressure driving winds, so a standard altitude is used: all recorded pressures are transformed to their sea level equivalence.

An example is the figure above. The half-dark circle shows cloud cover: roughly half the sky is covered with clouds; the two numbers at the left are the air temperature (48°F) and the dew point (45°F); the two dots indicate moderate rain, and the "138" is a code for air pressure: if the first number is 0, 1, or 2, put "10" before the number; otherwise do not change the number. So "138" becomes 1013.8 millibars. If the number were 9963 it would be 996.3 millibars.

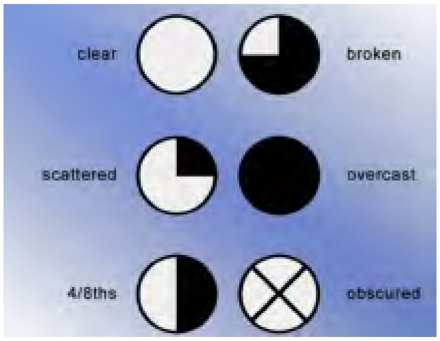
The line coming into the circle shows wind direction . Its source direction is that from which the line comes as it reaches the circle. The direction of that line tells us that there is a northwest wind; if the line were coming in on the right it would be an east wind. The symbols coming off of the line represent wind speed: one long line is 10 knots, a half-line is 5 knots, and no line is less than 3 knots. These symbols are used up to fifty knots (five long lines); wind speed over fifty knots is represented by a pennant and a series of lines, so 55 knots would be one pennant plus one 5-knot line.

The figures below show other symbols used in weather reporting.



Precipitation Symbols

Wind Speed Symbols



Cloud Cover

Thunderstorm

9. Troughs, Ridges, and Vorticity

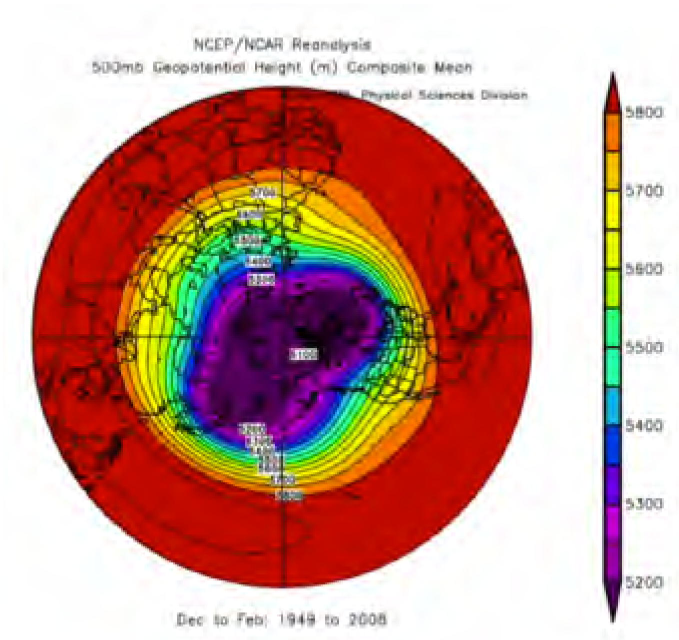
The pressure gradient within a low pressure area does not uniformly decline as the Low's center is approached. The phenomenon of *vorticity* creates troughs of low pressure and ridges and ridges of high pressure, just as a ski slope's average gradient is down but there are intermediate dips and rises in terrain.

In this section we focus on the relationship between pressure and altitude in the cyclonic wind of a Low.

Troughs and Ridges

Isopressure charts are contour charts showing the altitudes associated with a common pressure, unlike isobar charts which show different pressures. An isopressure chart allows the reader to focus on the altitudes associated with the same pressure. Isopressure charts are drawn for one of several standard pressures: 850mb, 700mb, etc. We will use 500mb as the common pressure

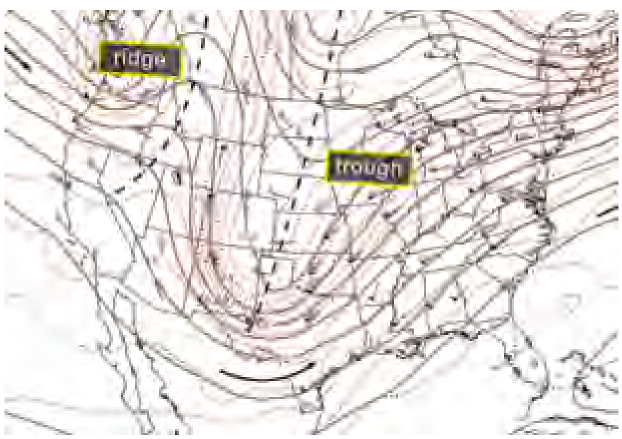
A global view of isopressures is shown below. This chart's color scale shows the altitudes associated with 500mb pressure in the northern hemisphere, as viewed from above the North Pole.



An Isopressure Chart

The scale runs from the blue-violet color at around 5,250 meters altitude (centered around the North Pole) up to dark red at about 5,800 meters (centered around the Equator.[4] Thus, high pressure areas like the Pole are attached to low-altitude isopressure lines; low pressure areas (the Equator) are attached to high-altitude isopressure lines. This makes sense: if a 500mb pressure exists at a 10 mile altitude while the same pressure exists at a 1 mile altitude elsewhere, the first is a higher-pressure area.

Isopressure charts show ridges and troughs. The weather chart below shows 500mb isopressure contours for the U. S. at a particular time. The solid lines indicate 500mb contours at higher altitudes, the dotted lines are isopressure contours at lower altitudes. The first are ridges, the second are troughs. Thus, in the Midwest there is a trough with a predominance of solid contours, while in the west there is a ridge associated with dotted contours.



500mb IsoPressure Chart, U. S.

What creates these troughs and ridges, and what is their effect on the weather? The primary factor is *advection*, to which we now turn.

[4] The global chart shows altitude in geopotential meters (gpms)—meters adjusted for the gravity effects of altitude. For the troposphere and stratosphere—with which we are concerned—these are not much different from meters.

Warm Advection and Vortical Advection

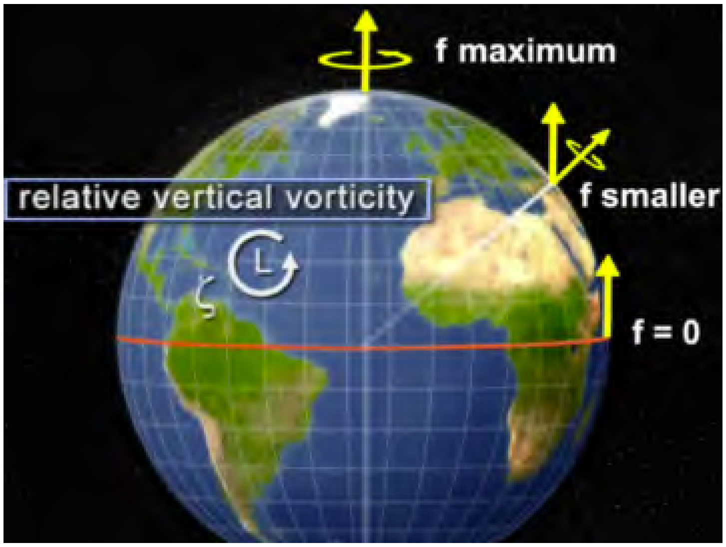
Advection is the flow of air (wind) due to temperature differentials, not pressure differentials (*convection*). Air moves between cool and warm areas even if it as at constant pressure. This motion is primarily vertical because warm air is more buoyant than cold air. Rising warm air is called *warm advection*.

Another source of advection is *verticality* or *vortical advection*. As its name suggests, it is associated with spinning air masses, such as those around Highs and (especially) Lows. The effect of spin is to throw the air to higher altitudes, much as stirring your coffee causes it to make a trough in the center and liquid pile up along the edge of the cup.

Vortical advection is the main source of ascending warm air in storm systems revolving counterclockwise around a Low. The process of *vorticity* is called *vertical vorticity* when the spin is around a vertical (up-down) axis, as when you stir coffee or spin a top; it is *horizontal vorticity* when the spin is around the horizontal axis, as when you spin a yoyo. *Vertical vortical advection* (say that ten times!) is the key to storm systems that rotate on an axis going through Earth's center.

Vertical vortical advection can be positive or negative. This is determined by the right hand rule: take your right hand with thumb up and curl your fingers; the result is a counter clockwise rotation (as seen from your thumb), just like the air flow around a Low. Your up-pointing thumb says that the vorticity is positive. To make your curled fingers form a clockwise flow, as around a High, your thumb must be pointing down—this means negative vertical vorticity.

The spin that creates vorticity comes from two sources. The most important is a local source: *relative vorticity*—the rotation of the air around a pressure area. *Absolute vorticity* due to the Earth's rotation in space affects advection at different latitudes. The figure below shows these two sources. A Low is formed around the Caribbean, an area of low absolute vorticity, creating *relative vorticity*—vorticality relative to the Earth's surface. At the same time, Earth is spinning, creating *absolute vorticality*, that is, vorticality relative to space.



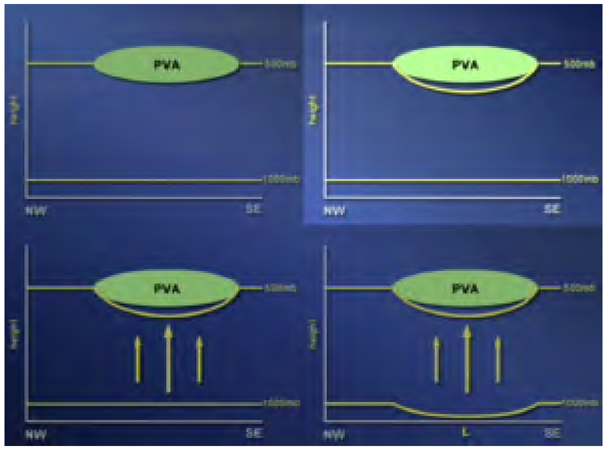
Global and Relative Vorticity

The degree of absolute vorticality is denoted as **f** in the figure. Relative vorticity in the Low is denoted by ζ. In the northern hemisphere there is positive absolute vorticality, highest at the poles, where the spin is more pronounced, and zero at the Equator. Thus, absolute vorticality is relatively minor for our Caribbean Low, but plays a more prominent role in Scandinavian Lows.

Total vorticality is the sum of absolute and relative v0rticality, $\mathbf{f} + \zeta$**.**

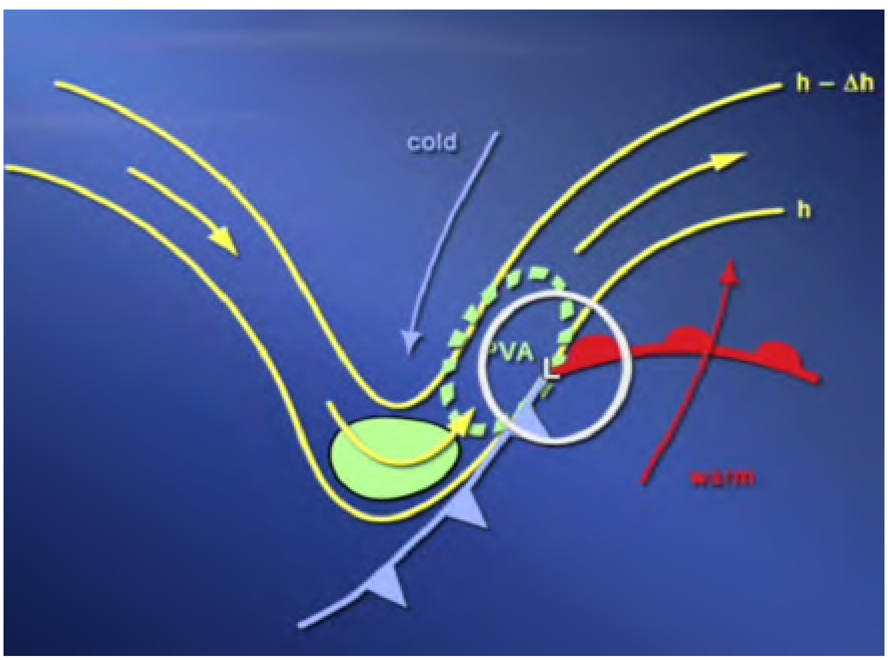
Trough Formation

In air around a Low the air ascent is called total *positive vorticality ascent* (PVA). The process of forming a trough is shown below.



PVA and Trough Formation

In the upper left, an area of PVA forms at, say, the 500mb pressure altitude. This causes air in that area to ascend, reducing local air pressure and reducing the 500mb altitude (right top). But the loss of air pressure above draws air from the lower altitudes (left bottom) and as this continues the 1000mb altitude also decreases right bottom). The sag created in the constant pressure altitudes is the trough.



Air Flow Around a Trough

The air arriving from the left is, therefore, drawn down toward the center of the trough, then up on the other side. Meanwhile, the PVA pocket has moved eastward, and with it the location of the Low's center.

Eventually the cold and warm air fronts that got the process going will merge into an occluded front near the low. The occluded front will break the low up, ending the storm.

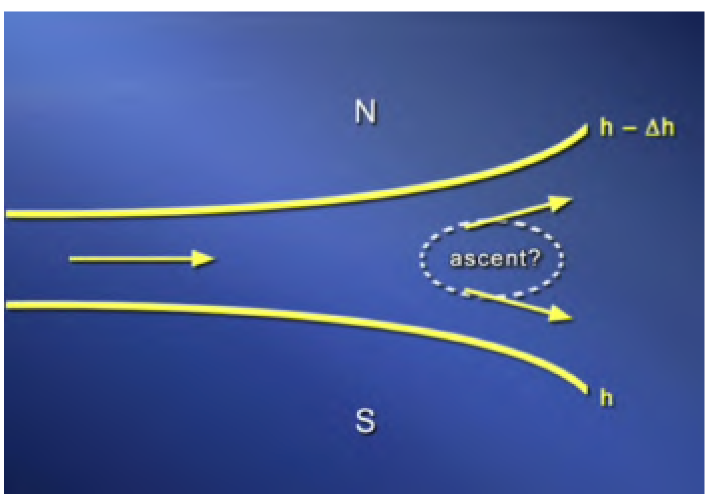
10. Vertical and Horizontal Wind Shear

Wind shear is the difference between wind speeds at different altitudes (*vertical wind shear*) or at different locations with the same altitude (*horizontal wind shear*). The primary question about vertical wind shear—our first topic—is why are horizontal wind speeds normally higher at higher altitudes? For example, the *Jet Stream* rushes along at about 100mph in the high troposphere while surface wind speed is typically much lower.

Vertical Wind Shear

The question raised above is a bit misleading—vertical wind shear is a bit more complicated. In fact, vertical shear does increase with altitude, but only up to the tropopause. Once into the stratosphere, the wind speed declines with altitude.

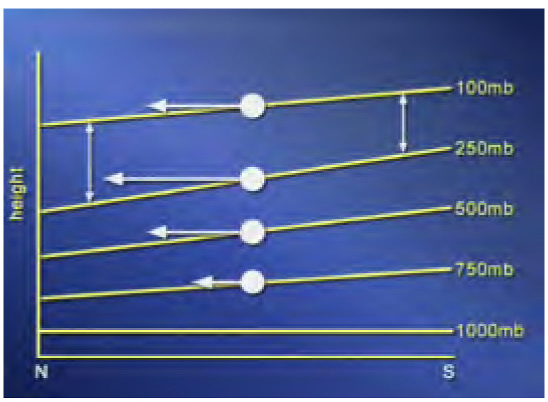
The figure below shows two isopressure lines, **h** and a lower isopressure **h – Δh.** The northern isopressure line is for a lower altitude than the southern because northern air is colder and denser, resulting in a lower pressure at a lower altitude.

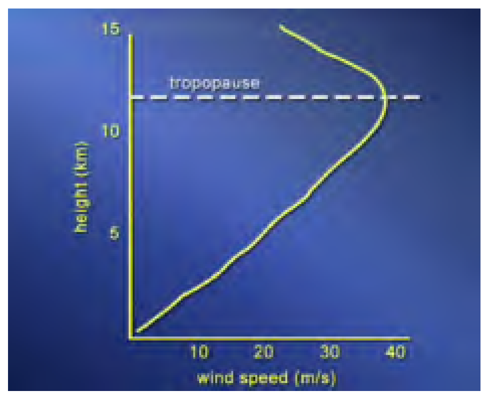


Isopressure Lines and Altitude

The wind direction is westerly (from the west), flowing along an isopressure line. As the isopressure lines diverge toward the east an area of air ascent can be created: the wind divergence is caused by wind slowing as isopressure lines diverge.

Below we see several isopressure lines with pressure declining as altitude increases. At the north (left) the isopressure line is at lower altitude than in the south. This tilt means that at a constant altitude pressure is lower to the north, creating a southerly wind. The length of the wind arrow, showing the relative wind speed, is higher as pressure falls because the air becomes colder with altitude in the north faster than in the south. The result is more tilt of the isopressure line as altitude increases, and a rise in the pressure gradient force with altitude.

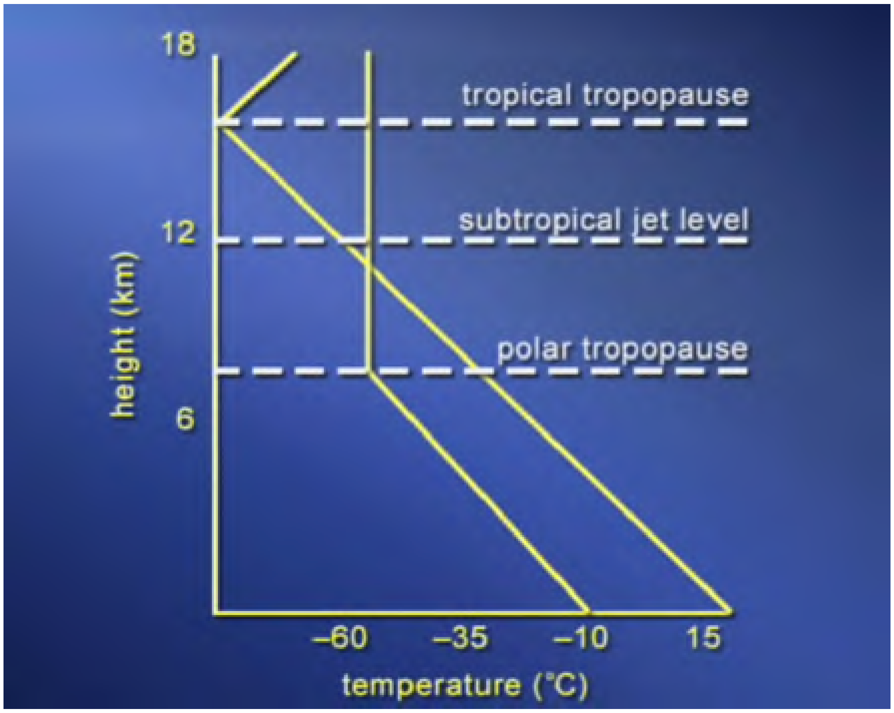




Wind Direction and Speed at Different Latitudes

But this reverses when pressure becomes less than 250mb, where stratospheric altitudes begin. Why do wind speeds increase with altitude in the troposphere but fall with altitude in the stratosphere? Well, the tilting of the isopressure line decreases with altitude in the stratosphere, reducing the PGF, and at some altitude in the stratosphere the tilt of isopressure lines actually reverses, creating a northerly wind. But why is that?

Recall that temperature actually increases with altitude in the stratosphere, as shown in the left side of the chart below. This is because the sun's radiation warms the air more than higher altitude cools it. On the right side we see that the tropopause (the altitude of 200mb pressure) is at a much lower altitude at the pole than at the equator. The two lines rising toward the left show the polar temperature (lower line) and the tropical temperature (right line) as altitude increases: they both decrease at the same rate but the polar temperature begins at a lower level.



Polar Temperatures and Equatorial Temperature

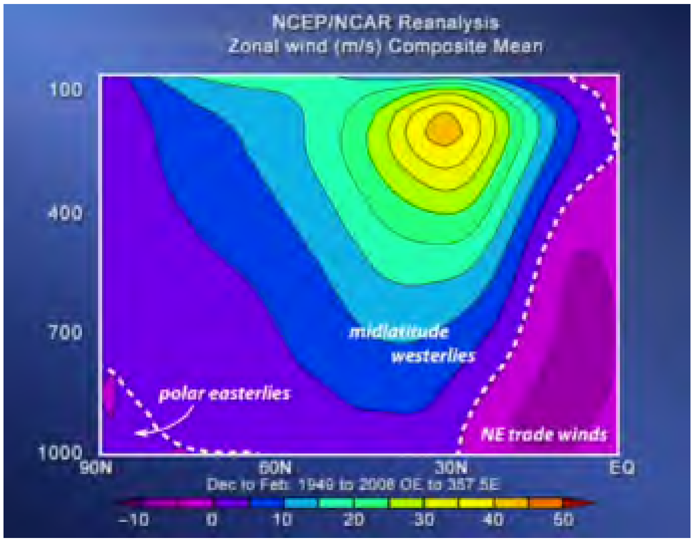
Up to the polar tropopause the equatorial temperature remains higher than the polar temperature at all altitudes, creating a lower polar tropopause and resulting in southerly wind.   But while equatorial temperature continues to decline up to the tropical tropopause, polar temperature remains steady through the stratosphere. At an altitude between the polar and tropical tropopauses the polar temperature exceeds the equatorial temperature and polar air's pressure declines relative to tropical air; as noted above, this raises the 200mb line at the pole. Up to that point the wind remains southerly but wind speed declines. At higher altitudes the wind direction reverses to become a northerly.

The maximum wind speed is reached at the *subtropical jet level* at about 30° latitude north. The subtropical jet stream is a high-speed band of westerlies that crosses the Caribbean Sea and the very southern U. S. at about 12km altitude (the tropopause).

In short, as we go up in the stratosphere the polar air warms relative to the tropical air and the PGF decreases, bringing wind speed down. At the subtropical jet level the wind speed is highest and above that it slows and eventually reverses to become a northerly wind.

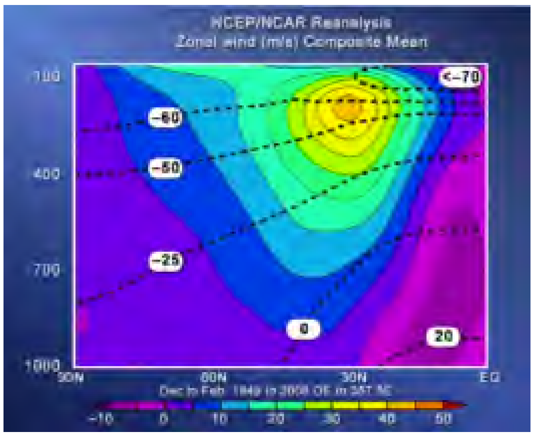
Zonal Winds in the Northern Hemisphere

The chart below shows the general wind flows in the northern hemisphere, color-coded for wind speed. Declining air pressure (increasing altitude) is on the vertical axis and the horizontal axis goes from the North Pole to the Equator. At the Equator the northeast trade winds occur, diminishing with altitude. At the subtropical latitude of 30°N there are slow midlatitude westerlies near the surface and fast subtropical jet stream winds at higher altitudes. These slower surface wind speeds prevail up through the polar easterlies.

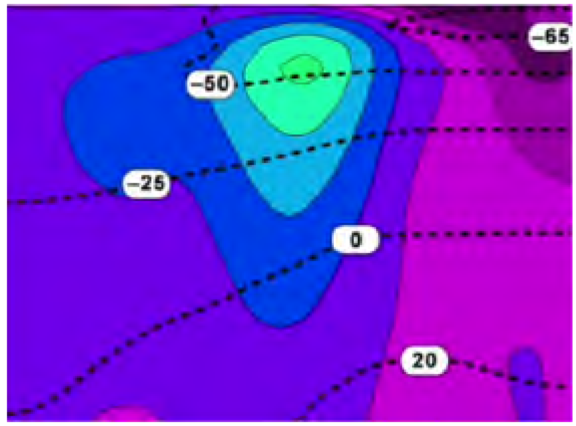


Northern Hemisphere Altitude and Wind Speed

The charts below add vertical wind shear and temperature (°C). The dashed black lines are *isotherms*, along which the same temperature prevails. A steep isotherm slope indicates more rapid temperature change with altitude and, typically, higher wind speeds.



Vertical Wind Shear, Winter



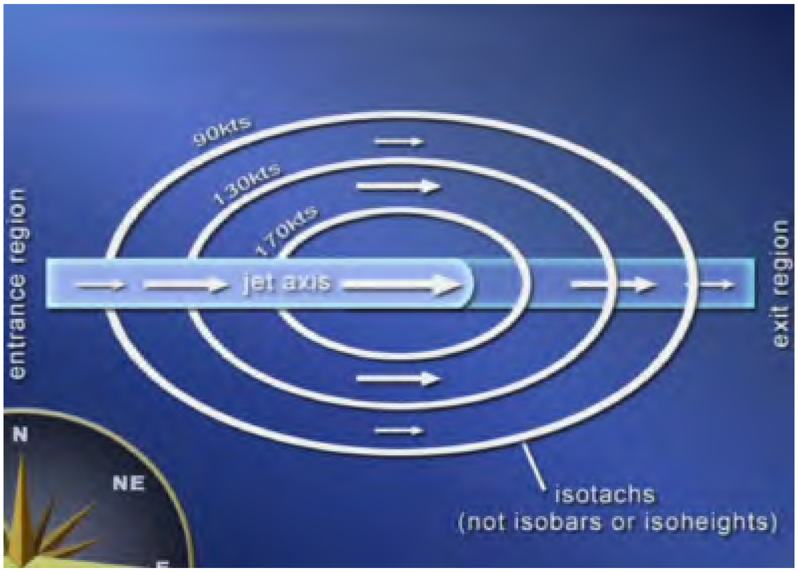
Vertical Wind Shear, Summer

The winter averages on the left show a fairly flat line at high altitudes, steepening as the surface is approached in the midlatitudes. The subtropical jet stream develops in the higher altitudes where yellow is the predominant color. Those rapid westerlies die away as the midlatitude westerlies emerge at about 500mb.

The summer averages on the right (ignore the incorrect Dec-Feb identification) show much flatter dashed lines and, therefore, smaller temperature differentials and lighter vertical wind shear.

Horizontal Wind Shear: Jet Streaks and Jet Streams

Jet Streaks are tubes of fast moving easterly winds that are typically transitory. *Jet Streams* are permanent tubes of fast moving air that circle the globe but frequently change their shape and latitude as they encounter Highs and Lows. Each hemisphere has two prominent jet steams—the *polar jet stream* and the *subtropical jet stream*. Within these tubes wind speeds of 100 knots are common, and winds up to 250 knots have been measured. The polar jet stream is what we in the U. S. consider "the" jet stream.

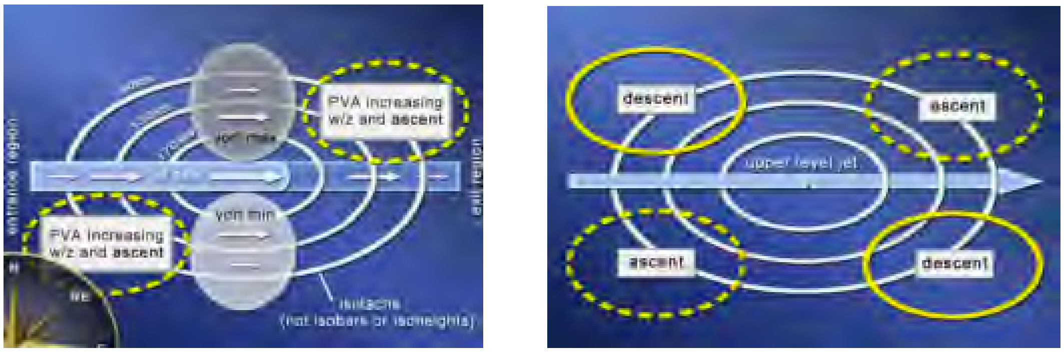


A Jet Stream

The figure above shows a cross-section of a jet stream. It flows west to east with the prevailing winds. The axis, or center of the wind tube, is the area of highest wind speed, in this case 170 knots (195mph). As the *isotachs*—lines of constant wind speed—show, wind speeds outside the axis are slower. Thus, jet stream winds behave like river currents—highest in the center where frictional forces are least, lowest at the edges where water is shallower and riverbanks obstruct the flow.

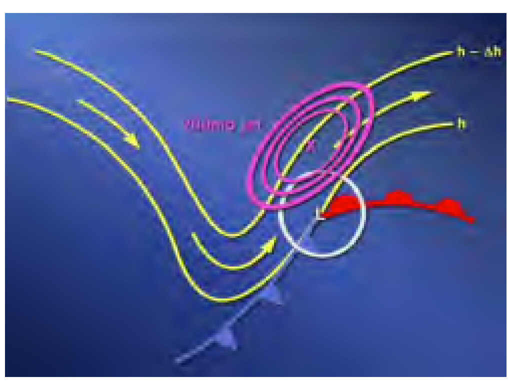
The jet stream has strong regional characteristics: the polar jet stream comes from our northwest over the Pacific and western Canada, then it typically turns northward over the United States as it approaches the high pressure band near 30° north latitude. Variations from this are pronounced as local Highs and Lows direct the stream; these latitudinal shifts in the jet stream have an important effect on weather: cold Canadian air masses are driven south to do combat with warm fronts when the jet Stream drives through low latitudes; warm air is brought north to fight cold Canadian air when the jet stream is at high latitudes.

Because the winds are slower on the sides of the jet axis, the wind directions diverge toward the northeast and southeast as the slower winds tug at the faster central winds. This divergence creates areas of positive and negative vorticity, giving the jet stream some rotation.



The Jet Stream, Vorticity, and Vertical Wind

The charts above show a more complex structure of a jet stream. In the left figure we see a *vort max* (vortical maximum) above the axis and a *vort min* (vortical minimum) below the axis. Recall that areas of higher positive vorticity create air ascent. Positive vertical advection (PVA) develops to the east of the vort max in response to the higher vorticity to its west, while another area of PVA develops to the west of the vort min, again in response to the higher vorticity toward the west. The result is shown in the right figure: a vertical rotation of air occurs in the west section of the tube, and an opposite vertical rotation occurs in the east section.



Cyclonic Trough and Jet Stream

Above we see a jet stream at 200mb pressure combining with a trough formed by the collision of a cold front and a warm front. The higher altitude stream adds to the positive vorticity of the cyclone, deepening the trough and increasing the wind speed.

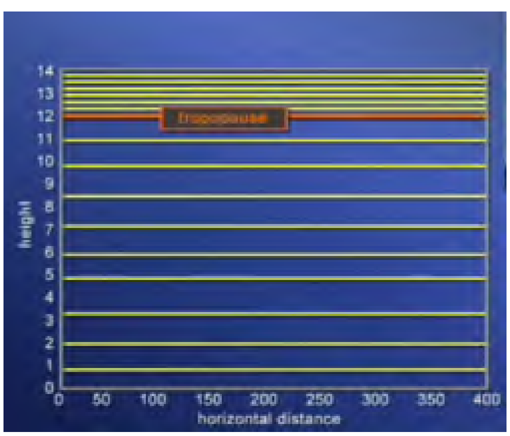
11. Terrain and Weather

Terrain alters meteorological conditions, sometimes dramatically and rapidly. We see this in the western U. S. where westerly winds encounter the Rocky Mountains, creating a variety of effects both near and far away. The mountains do this by destroying atmospheric stability—a condition in which air parcels dislodged from equilibrium tend to return to that equilibrium—and creating chaotic conditions..

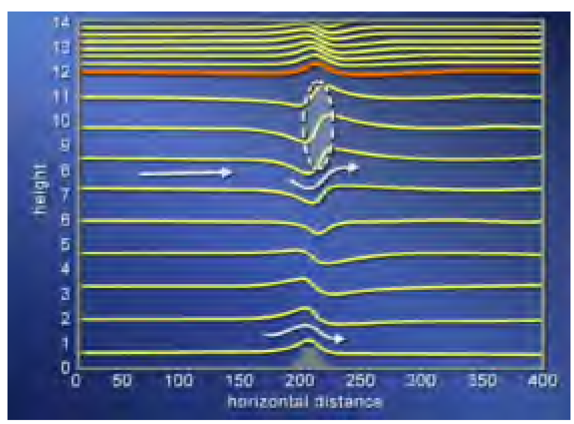
Isentropes and Meteorological Instability

We know from the second law of thermodynamics that nature tends toward increasing entropy, that is, toward increasing disorder or randomness: hot and cold areas mix until the temperature is uniform—structure (hot vs. cold) has become lack of structure (lukewarm throughout); living objects go from a highly ordered body state to complete disorder through death and decay.

The entropy principle applies to all natural phenomenon, including weather. One of the tools used to explain mountain weather is the *isentropic map* showing lines of equal entropy. In this context, entropy is measured by *potential temperature*, defined as the temperature that would prevail in a dry air parcel if it were under a standard pressure, say 1000mb. [5]



Flat Isentropic Lines



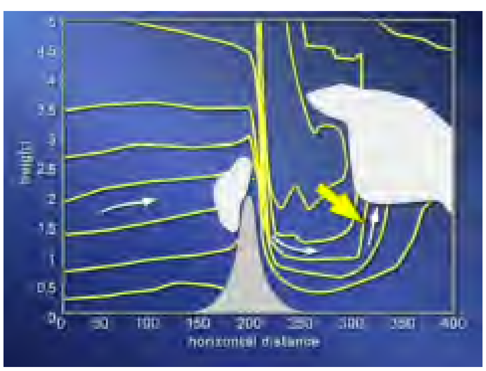
Isentropic Lines around a Mountain

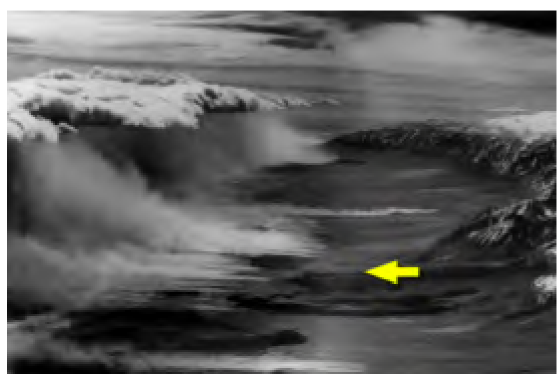
[5] Potential temperature or entropy (θ) is defined as $\theta = T(P_0/P)^{R/c}$: it is directly related to the actual temperature (T) and inversely related to the actual air pressure (P); P_0 is the standard pressure, R is the gas constant, and c is the gas's heat capacity.

While actual temperature falls with altitude in the troposphere, potential temperature rises with altitude. In the left chart there is a uniform atmosphere (constant entropy at each altitude) with no effects of terrain. The flat lines are *isentropes*, points of equal potential temperature. A rise in an isentrope indicates warming air; a fall means cooling air.

Air flows along isentropes, as shown in the right chart where the flow is a westerly. But when the air encounters a mountain and is thrust upward, each isentrope above the mountain rises (indicating cooling air at each altitude). Eventually the isentropes return to their original altitudes as the mountain is left behind. In the upper atmosphere the potential temperature increase is especially high, and the gray oval suggests an area where moisture absorption might create a cloud.

The air around mountains can become very unstable, as shown below. Here the isentropes are seriously distorted as air undergoes chaotic changes in potential temperature.





Downslope Wind and Hydraulic Jump

Here the potential temperatures rise on the lee side of the mountain. The colder dryer and denser air then plunges down the mountain in a heat-creating katabatic wind that reverses dramatically in an *hydraulic jack* about 300 miles to the lee. That reversal is an equally dramatic surge in potential temperature with the air expanding and absorbing moisture, creating a cloud.

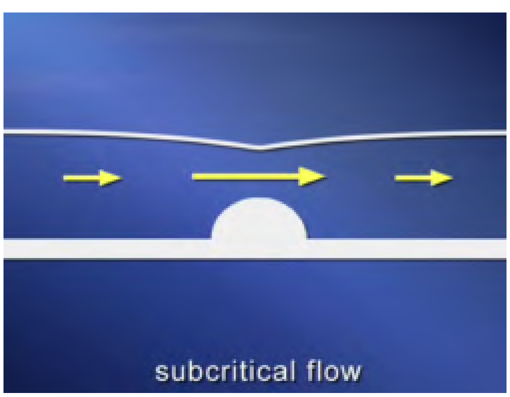
On the right we see an actual example: the katabatic wind on the lee side of the mountain creates a hydraulic jack at a distance, pushing the clouds well back from the mountain. In this famous photograph, taken in 1951 in California, the cloud was filled with the dust shown blowing to the left.

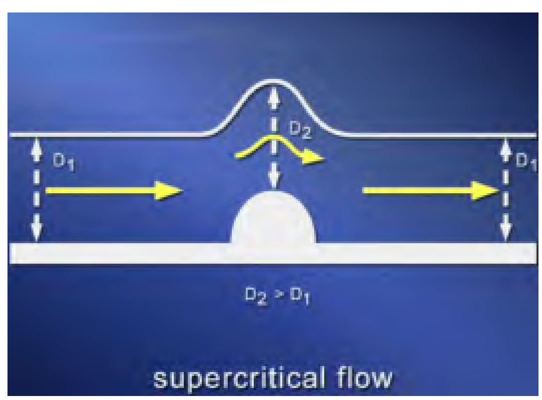
Downstream Consequences of Instability

The atmospheric instability around mountainous areas can certainly make life in the local area—and in airplanes above it—interesting. But it also is the basis of Lows far to the to the east of the mountains: Lows created at the Rockies are an important cause of eastern thunderstorms. To understand this we need to explore *supercritical and subcritical flows*.

A medium like liquid or air thickens in the presence of turbulence. For example, let water run into a sink at high pressure. At first the water will flow away from the spot it hits in a thin stream bordered by bubbles where it slows down. That thin flow to the exterior is a subcritical flow in which water density is unchanged.

But eventually the water density will increase as the congestion around the edges creeps in the center—the water will thicken. This is a *supercritical flow*—the medium has thickened





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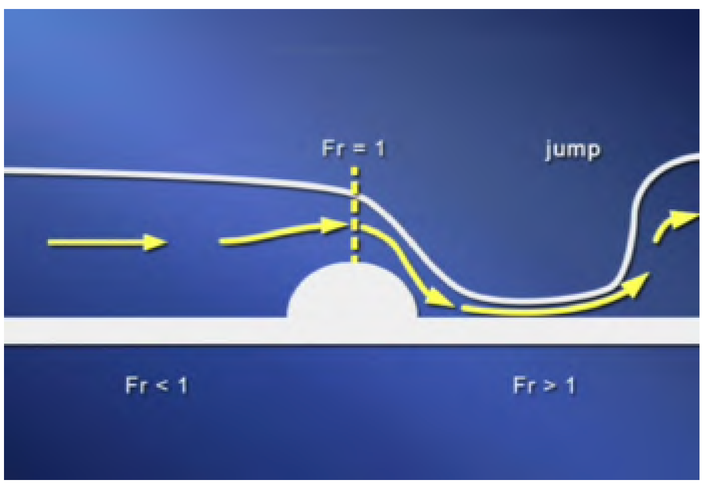
On the left we have air passing over a mountain with an isentrope above it. The wind speed is slow so the only effect is that the isentrope dips slightly, reducing potential temperature, warming the air above the mountain, and increasing the wind speed. On the lee side the original potential temperature returns, as does the original wind speed. As air passes over an obstacle in a subcritical flow it thins, speeds up, then returns to its original speed.

But if the arriving wind speed is sufficiently high, the airflow is directed upward, raising the isentrope and reducing temperature above the mountain. Wind speed is slowed at the mountain, then air falls in the lee and the original wind speed is restored. pulling the air with it; the gap between the mountain and the isentrope widens so the rising air is more vertically directed. The wind speed falls over the mountain and returns to its original path and speed in the lee.

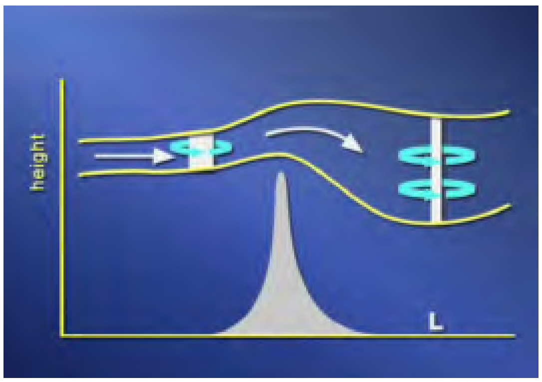
But the effects can be more dramatic. Below is a subcritical flow that becomes supercritical as it reaches the mountain, sharply reducing the isentrope in the lee.



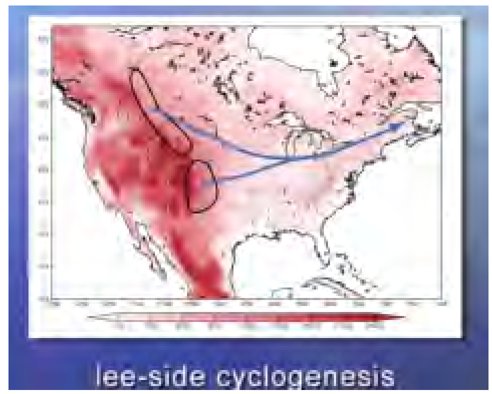
Subcritical to Supercritical Transition

This is, in fact, another view of the previous case of downslope wind with hydraulic jump. The chart shows that the crucial characteristic for criticality is the Froude Number (denoted as Fr). A subcritical airflow has Fr <1, a supercritical flow has Fr >1. The Froude Number is a concept from fluid mechanics: it is directly proportional to the fluid's flow speed and inversely proportional to the fluid's depth; in this case the depth is the distance from the isentrope to the surface—the drop in the isentrope (narrower depth) and increase in wind speed have created highly supercritical conditions, i.e. turbulence as air thickens and gets in its own way.

The development of Lows in the Midwest and eastern U. S. begins at the Rocky Mountains. The wind over the Rockies arrives with some internal vorticity. As the isentropes widen in the lee, positive vorticity increases and air ascends, creating a Low.



Vorticity in the Lee



West-to-East Flow of Lows

The right chart shows a common pattern in the U.S. The oval around the upper Rockies shows an area of low pressure in the lee that travels southeast toward northern Indiana; this is the "Alberta Clipper." Another area of Low development is in the lower Rockies around Colorado. That Low travels to the northeast, again toward northern Indiana. In the Ohio Valley the paths of the Lows converge so the east coast gets a double chance for lows and storms.

12. Thunderstorms, Squall Lines and Radar

Next to hurricanes the most exciting weather is the thunderstorm and its associated squall lines. Thunderstorms are a collection of individual thunderstorm cells, often long-lived. Squall lines are leading sections of thunderstorms that occure when horizontal wind shear and other conditions are right.

Radar

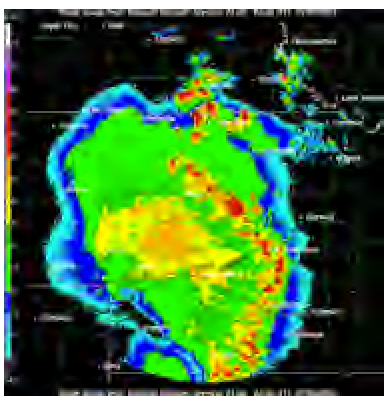
Radio Detection and Ranging (Radar) was a World War II development that is now an essential source of weather analysis. Radars come in two basic sizes: large-dish units called *precipitation radar* with a resolution of 3cm (1 inch), capable of identifying large raindrops, and 10cm small-dish units called *cloud radar,* able to identify larger objects like birds, bats, and large hailstones.

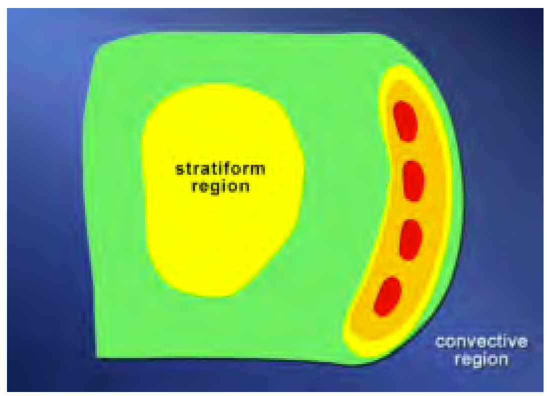
Radar operates by transmitting pulses of electromagnetic radiation at specific microwave frequencies, then receiving the reflected radiation (*backscatter*) and recording it on a monitor. The intensity of the backscatter's reflectivity is measured in decibels (dBZ, a logarithmic scale ranging from -30dBZ to +75dBZ)—a doubling of decibels is a ten-fold increase in reflectivity. Because air is non-reflective while moisture is highly reflective, the backscatter shows the intensity of moisture content.

The figure below shows the locations of NEXRAD (Next Generation Radar) stations in the U. S. The U.S. radar stations are uniformly distributed around the continental U. S. and its noncontiguous states and territories; South Korean stations under U. S. control are also shown.

Below the Doppler station chart are two T-storm images. The first is an actual image of a 2003 thunderstorm over Lincoln, Nebraska. The second is a typical structure of a thunderstorm.

U. S. NEXTRAD Doppler Stations

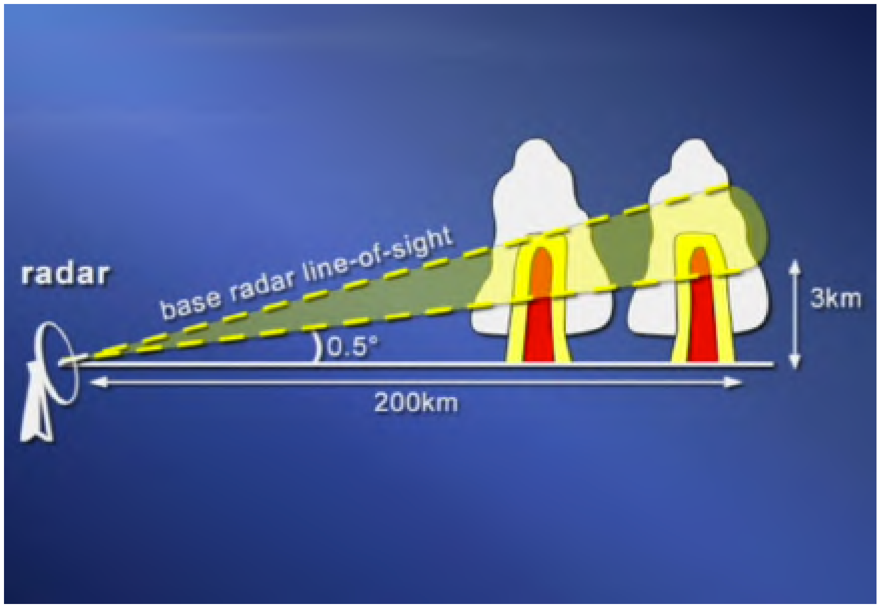




T-Storm Image, Lincoln NE **Typical T-Storm Structure**

The radar image on the left shows a green-blue boundary with minimal reflectivity, an internal green-yellow area of highly reflective moisture, and red cells along the eastern front with the most intense reflectivity, marking a squall line.

This conforms to the standard T-storm formation shown on the right (as seen from above): the storm front is the *convective region* where cold winds from the rear meet warm winds from the front, creating gusts. Just behind the storm's front is a squall line with especially high moisture, and behind that is a large area of moderate moisture through which a vertical column of higher moisture called the *stratiform region* passes; this is a central column of rising warm air.

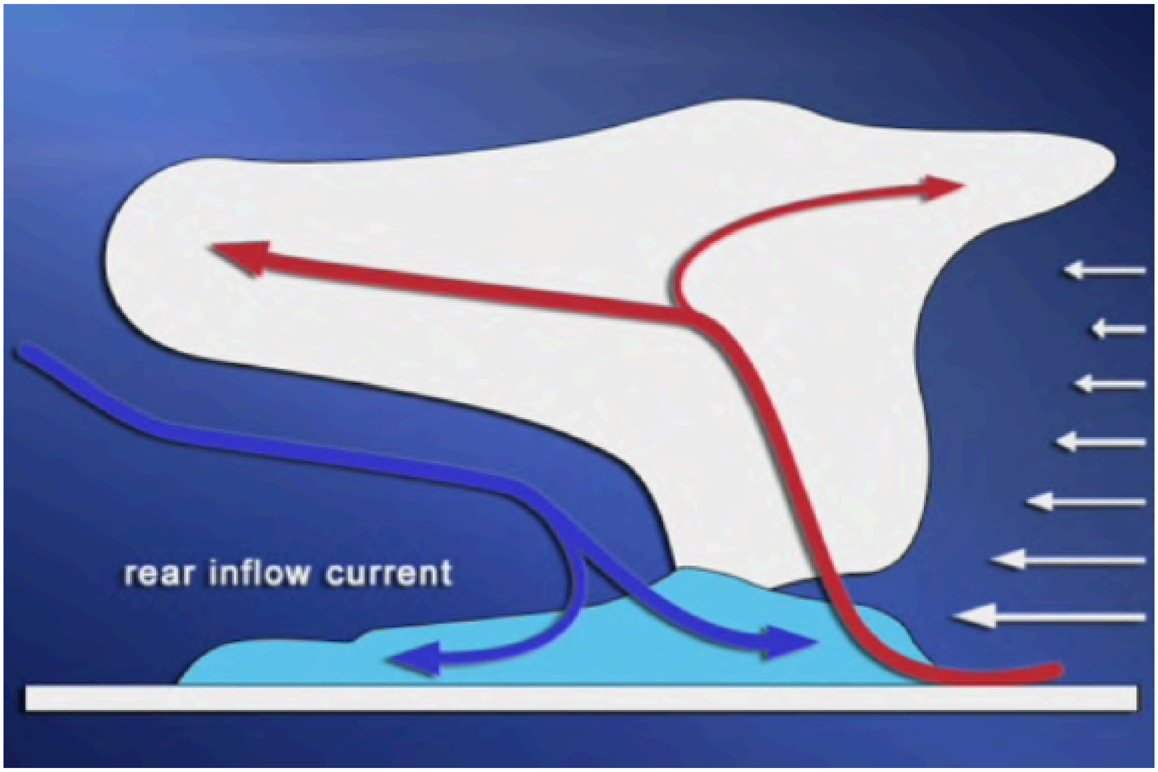


Above we see a radar scan of the altitudes. Typically, radar is set to scan at a baseline from 0.5° to 19.5° above horizontal. This traces out a vertical cross-section of clouds and moisture. Nearby storms are severely clipped at top and bottom, distant storms less so.

Thunderstorm Formation

The prevailing westerlies over the U. S. suggest that thunderstorms have the wind at their backs. But thunderstorms travel eastward faster than the winds pushing them—the storm outruns the westerly wind. So the *relative wind direction* around a T-storm is easterly. The result is vertical wind shear that declines as altitude increases; that decline with altitude is due to westerlies with greater speed at high altitudes, creating lower-speed relative easterlies as altitude increases.

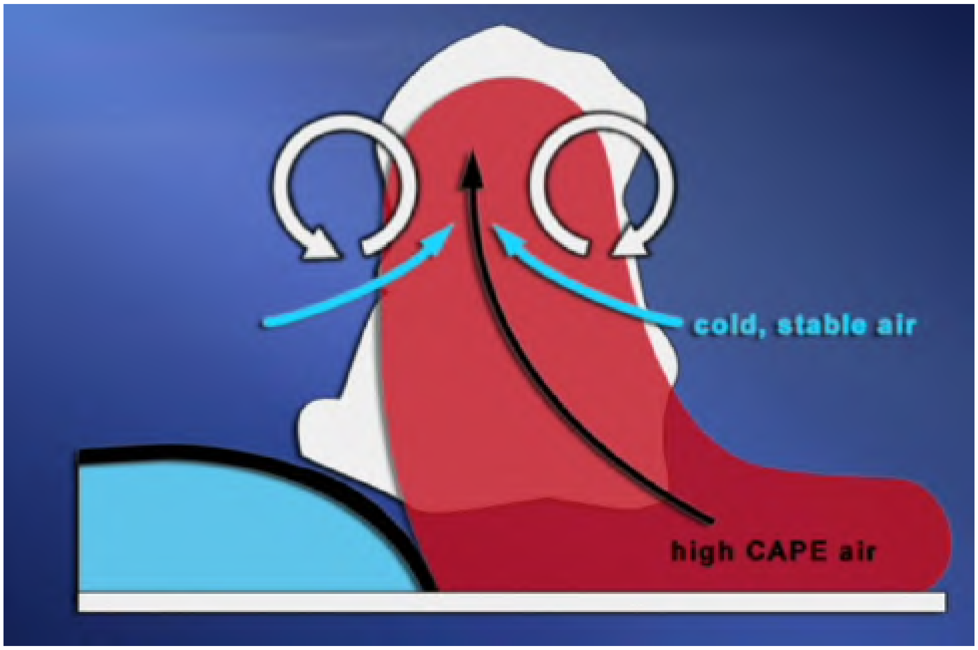
In the chart below, relative vertical wind shear is shown on the right; as noted, it is lower at high altitudes. The airflows are the frontal inflow of warm and positively buoyant air (red) and the rear inflow of cold and negatively buoyant air (blue). The front boundary of the cold front is an area of high wind gusts called the *gust region*.



Wind Shear and Air Circulation in a Thunderstorm

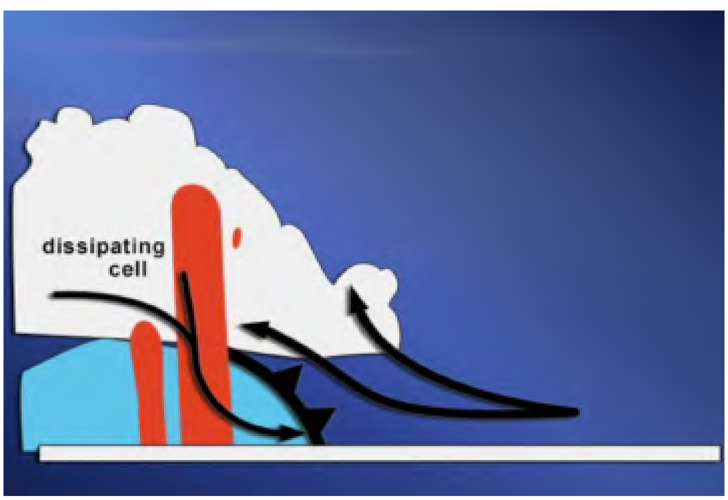
The decreasing vertical wind shear (seen above) at the front creates an area of clockwise circulation with positive vorticity and buoyancy. Behind the cloud is an area of negative vorticity and negative buoyancy, created by the rear inflow.

A cross-section of the typical storm is shown below. A cold front is shown at the left. The boundary of that front is the gust region. The cold front is meeting warm, moist air, which is positively buoyant, developing CAPE as it rises and forms a cloud. The clockwise flow at the front has warm air with positive buoyancy, while the counterclockwise flow behind the cloud has cold air with negative buoyancy. The circular flows draw cold wind in to meet the column of rising warm CAPE air. The result is strong updrafts and downdrafts within the column.



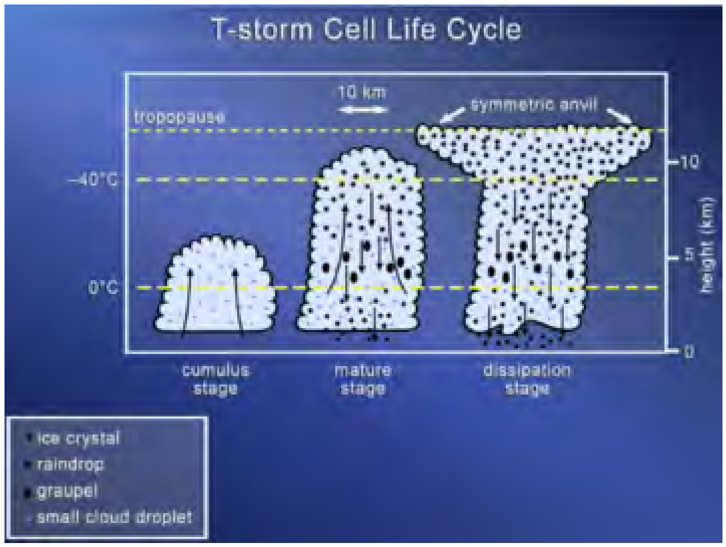
Horizontal Cross Section of a T-Storm

The figure below shows a diagram of the typical horizontal radar image of a T-storm. The warm CAPE air is coming from ahead of the T-storm cell, and the cold negatively buoyant air arrives from behind the storm. Gusts are created near the surface where they meet, and the column of CAPE air (red) shows the core of the storm's moisture. Ahead of it is a small red area showing the remains of the CAPE air from a previous time, and at the left is a medium-size CAPE air column—the next-to-be-created high column.



Radar Image of a Thunderstorm (in red)

T-storm cells are born and soon dies, but constantly replenishing themselves. So one cell's cycle is not the cycle of an entire T-storm. The life cycle of a T-cloud cell is shown below.



T-Storm Life Cycle

The thunderstorm begins with a simple cumulus cloud created by positively buoyant air that keeps the cloud's small water droplets from falling via evaporation. As the storm matures it extends to higher altitudes with the inflow of positively buoyant warm air and outflow of cooler air creating internal updrafts and downdrafts; light precipitation develops. Finally, the rear cold air inflow begins to dominate and condensation releases heavy rain and latent heat. The late-stage thundercloud develops a wide anvil top at the tropopause where the air begins to warm again.

Squall Line Formation

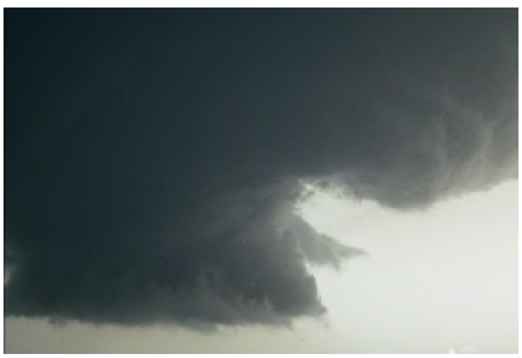
As noted above, squall lines are areas of intense instability with high winds and driving rain. They consist of multiple cells, each short-lived but rapidly replaced by new cells. An essential characteristic is that they have *sub-cloud cold pools produced by evaporation*.

The evaporation at the cloud base arises from the warm air into which a cold front moves. The warm air removes moisture from the cool air, which releases its latent heat and adds to the warm air's temperature.

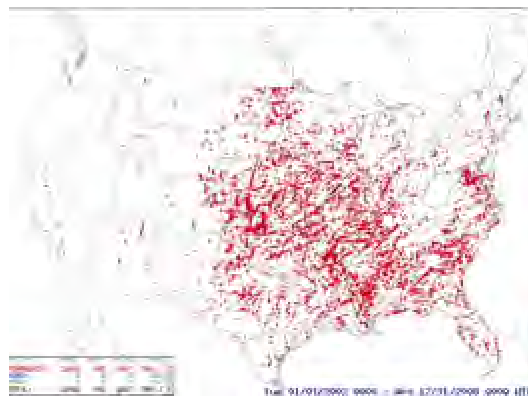
13. Supercells, Drylines and Tornados

Tornados are rapidly rotating *supercells* that can have winds over 300mph. They can be short or long lasting; They might or might not touch the ground; they are typically rainless because falling water drops are evaporated by the warm updrafts. Tornados can occur in sequence and their very dark ominous color is due to debris and condensation.

An early indication of a tornado is a *wall cloud*, shown below, with a flat bottom near the surface. Note the hook shape at the bottom; we will see this again.



Tornadic Wall Cloud



Tornados Sighted, 2002-2008

The right chart above shows that tornados are primarily found in the eastern half of the U. S., often in sparsely populated areas. There are few tornados in California but they are in very populated areas and do more damage that an eastern tornado.

Drylines and Supercells

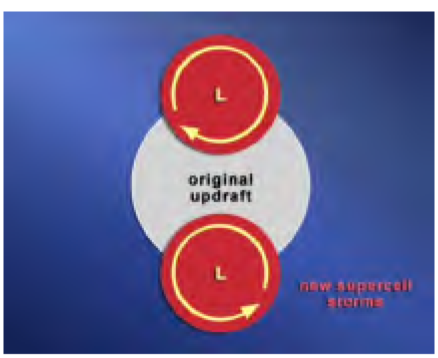
Tornados are born in regions with sharply different dew point temperatures; a *dryline* is the line of demarcation. In the U. S. drylines have higher dew points to the east and lower dew points to the west. The left chart below shows a dryline marked by orange circles, extending from northern to southern Texas. To the east is maritime tropical air (mT); to the west continental polar air (cP) prevails.

The right charts shows the most common dryline states.

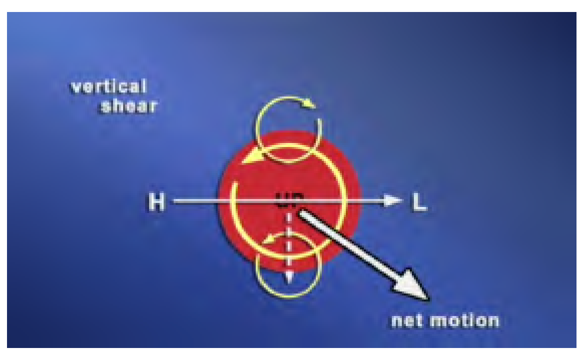
A Dryline

Common Dryline Areas

A dryline is a spawning ground for *supercells*. The formation of supercells is shown below, seen from above. An updraft of air encounters rising vertical wind shear to its north and falling vertical winds shear to its south. The north cell (a high but mislabeled as a low) gets a clockwise rotation and the south cell gets a counterclockwise rotation; the opposite rotations tend to push the supercells apart. The lower supercell has two forces on it: a west-to-east force from the prevailing wind, and a downward force from negative vorticity. The net force is to the southeast, sending the cell in that direction. Thus one updraft spawns two supercells with the top cell moving leftward and the bottom cell moving rightward.



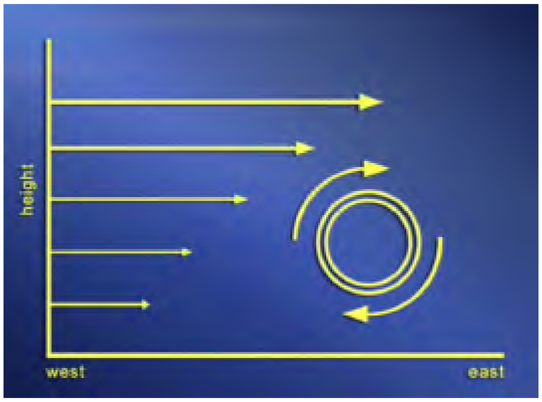
Supercell Formation



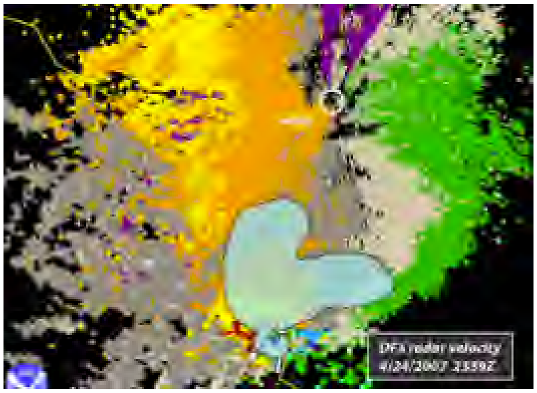
A Right-Moving Supercell

Tornados

The vertical wind shear across a dryline (shown left below) initiates a clockwise vertical rotation with negative vorticity. This pulls the warm air in and up, with the cold air rotating over the top.



Vertical Wind Shear and Vorticity

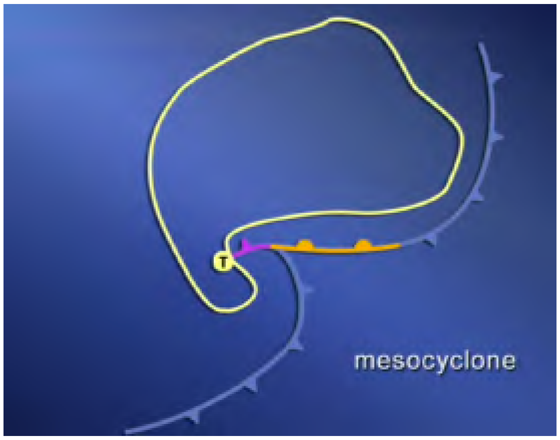


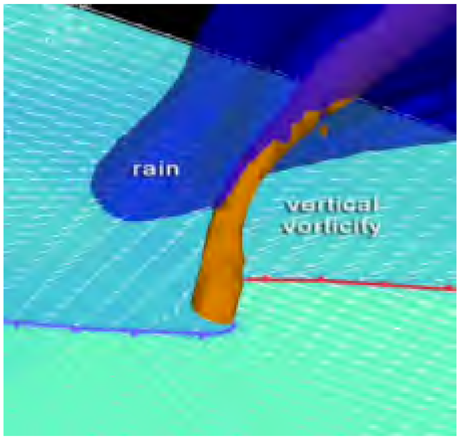
A Tornadic Radar Echo

A tornado's radar echo is shown above right. This is a wind velocity chart: green indicates wind coming toward the station, yellow indicates movement away. The radar station, located at the circle, experiences wind coming toward it from the east and going away from it to the west. The blue area is a probable tornado. Note the hook at the southern edge: this is a common sign of a tornado. The hook area has rapid counterclockwise rotation due to downdrafts located just west of updrafts.

Below left is a tornadic area with the telltale hook. The eastward moving cold air is shown as the blue front. The orange line is westward moving warm air, which has blown through the cold front. There is a local occluded front emanating from the top of the hook. The air circulation is counter clockwise as negatively buoyant cold air meets positively buoyant warm air.

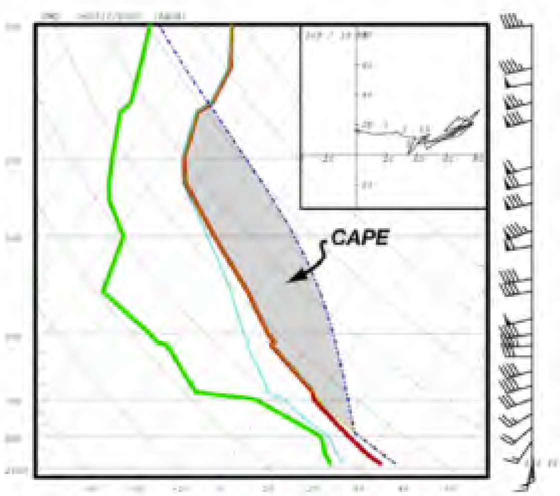
The right chart below is another view of a tornado, shown as the orange funnel with vertical vorticity. The cold air from the west and the warm air from the right rotate counterclockwise, setting up the region of strong updrafts and downdrafts. Note the hook at the bottom.





Vertical Cross Section of Tornado **Structure of Tornado**

A tornado's *Skew-T Chart* is shown below. The vertical axis is pressure in millibars, the horizontal axis is temperature in centigrade. The green line isows the relationship between dew point temperature and pressure; the red line is the actual temperature vs. pressure relationship. The straight dashed lines from lower left to upper right are isotherms, showing constant temperature-pressure relationships. Note the wind speed symbols to the right; the maximum wind speed is 80 knots (92mph).



***Skew-T Chart* of a Tornado**

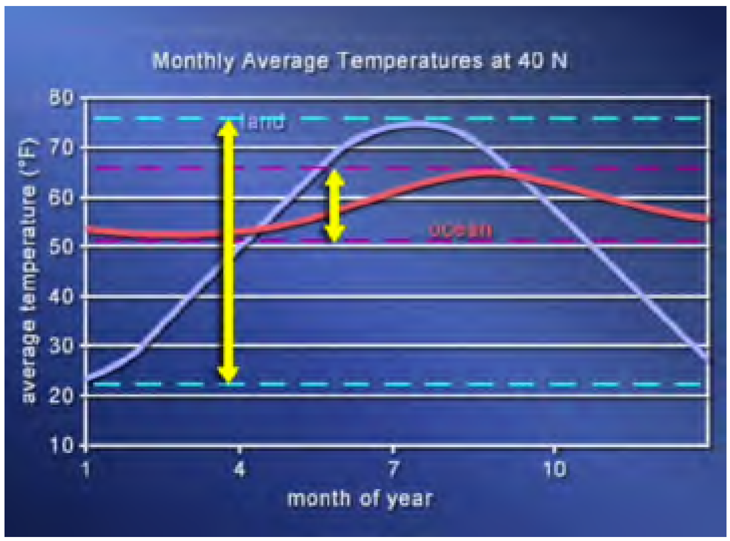
The blue dashed line is the moist adiabatic temperature-pressure curve. The gray area is the CAPE, the accumulated potential energy associated with the tornado.. The CAPE is a very high 6,500 joules of energy. This is a very powerful supercell.

Measuring Tornado Intensity

Tornado intensity is measured on the Fujita Scale, ranging from F0 to F5. An F0 tornado has winds roughly equivalent to a low-level hurricane F5 tornados are extremely strong, higher than a CAT 10 hurricane.

14. The Ocean and Weather

Earth's oceans have a profound effect on the weather, in large part because they are a major store of the energy that feeds storms. This is due to the relatively high thermal inertia of water—temperatures mix slowly and retain their heat longer.



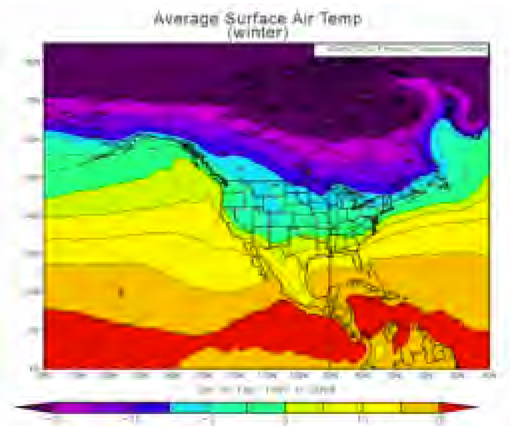
Land and Sea Surface Temperatures

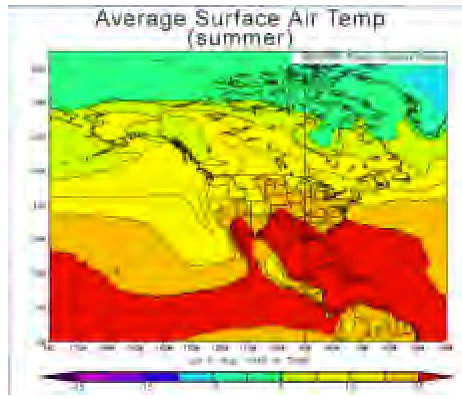
The chart above shows the monthly average sea surface and land temperatures at 40°N latitude. The land temperature varies far more dramatically than sea temperature due to the difference in thermal inertia. This is, of course, the cause of the daily shift between land and sea breezes discussed earlier.

<u>Ocean Temperature and Pressure</u>

The charts below show the winter and summer ocean temperatures in the northern hemisphere. The most pronounced temperature change is in the Gulf of Mexico and the Western Atlantic, where the very warm Caribbean temperature creeps up to the mid-Atlantic states, bringing tropical storms and hurricanes with it. Canada and Alaska also warm significantly, as does the northeastern U. S. But on the

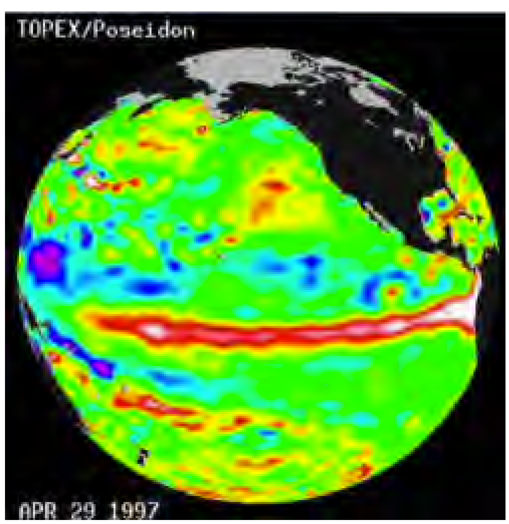
Pacific coast the temperature is fairly steady throughout the year except at higher latitudes; this is due largely to the Pacific High and to the warm western pacific waters of southeast Asia.





Perhaps the best-known area of warming is in the eastern Atlantic at equatorial latitudes, where *El Nino* occasionally warms the waters off of northwestern South America , especially Ecuador. This warming reduces water density and brings fewer nutrients to the surface, thus affecting the fish and bird populations (including, famously, the anchovies used for livestock feed).

El Nino is a cyclical phenomenon, occurring every 3-5 years. It arises from an ocean wave that travels first westward to southeast Asia, then echoes back to South America. A snapshot of El Nino at a peak time is shown below.

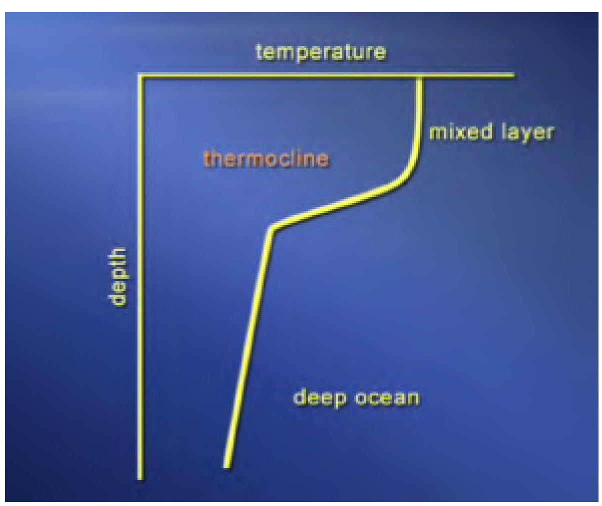


El Nino at its Peak

The red band is the current that carries El Nino (and its colder sibling, *La Nina*, not shown), back and forth. The white area at the right is extremely warm water off Ecuador; that is El Nino. A time lapse video of this red band shows it moving westward, then eastward again, in an endless but imprecisely timed cycle.

We know that water temperature is colder and pressure is higher as water depth increases. The effects on water circulation and currents is akin to those we've discussed for air circulation: water moves from cold and dense toward warm and less dense, and as it moves it warms and draws colder water in behind it.

What is different is the rate at which pressure and temperature change as depth increases. The Lapse rate of air is fairly constant, with temperature dropping steadily as altitude increases. But the lapse rate of water is quite different. Below we have an average temperature-depth relationship. The temperature remains fairly steady through a *mixed layer*, where it mixes well and creates an even temperature. Below that is a narrow region of sharp temperature decrease called the *thermoincline*, and below that is another region of fairly stable temperature. The thermoincline is at roughly 100-300meters (300 – 1000 feet).

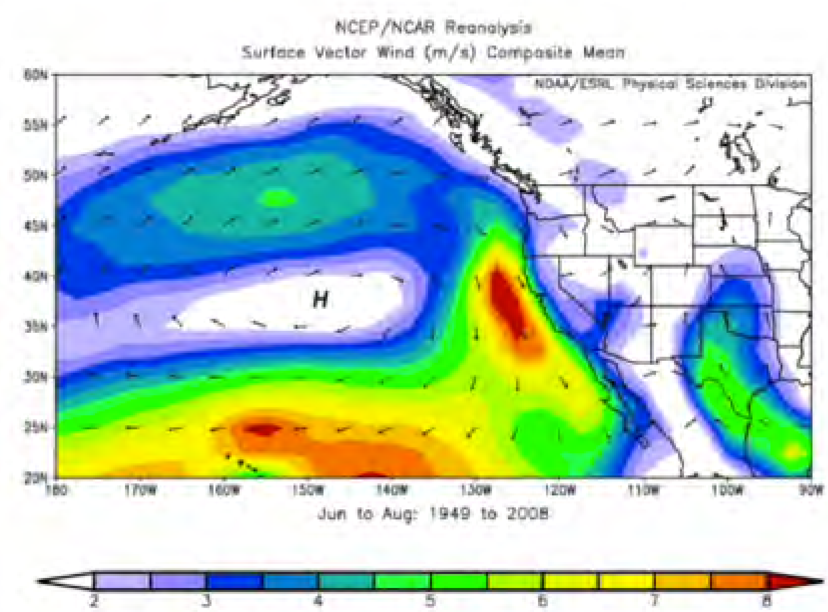


Ocean Depth and Ocean Temperature

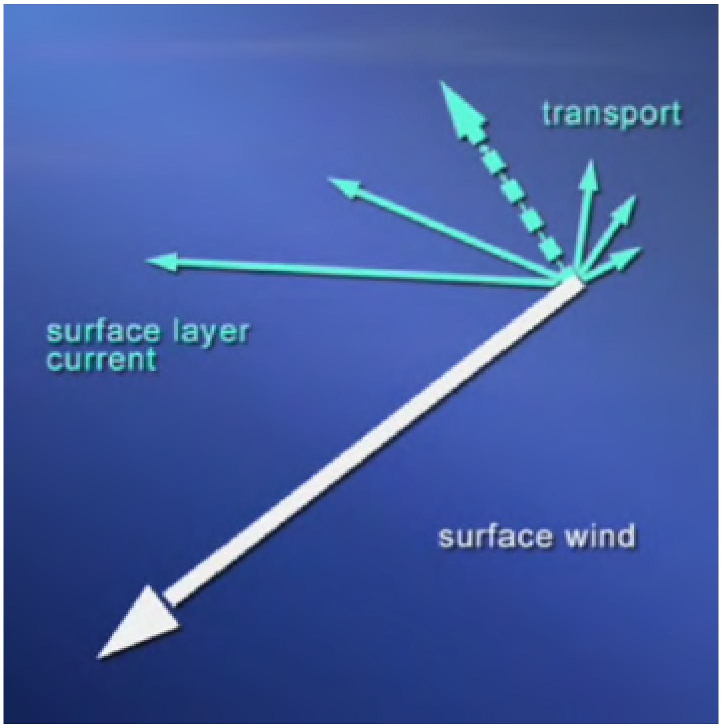
Perhaps the most dramatic change with ocean depth is in pressure. Air pressure at the surface—the pressure from all air in the atmosphere—is about 1000mb, but air plus water pressure jumps to about 2000mb at only 10 meters (33 feet) of water depth. As we know, deep ocean depth has crushing water pressure.

<u>Ocean Wind and Currents</u>

The chart below shows average summer wind vectors in the north Pacific. A clockwise circulation around the Pacific High sets the pattern for that ocean: southeasterly wind north of the high, easterly wind south of the High, southerly wind west of the High, and northerly wind along our west coast.



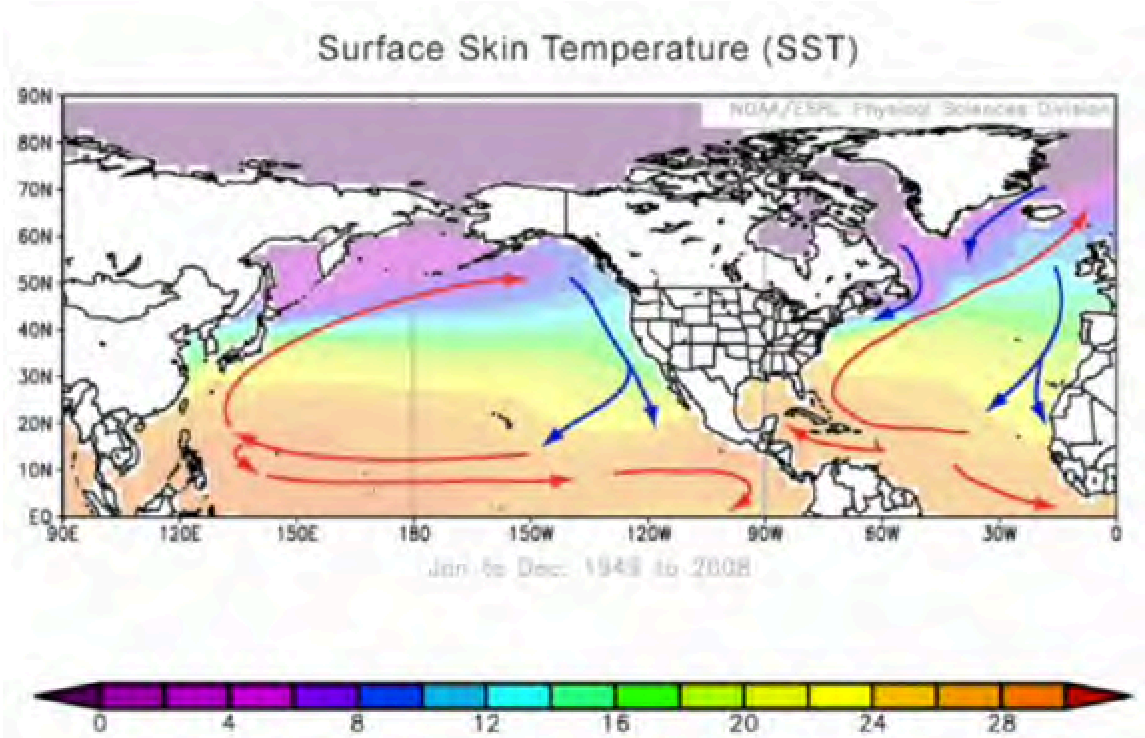
The relationship between prevailing surface wind and ocean currents is complex. The chart below demonstrates the general pattern. The large white arrow shows a northeasterly surface wind. The density of the water makes it difficult to move, and the Coriolis effect pushes the surface current to the right of the wind, along the first long blue arrow. This, in turn, pushes the adjacent water even further to the right. Ultimately, a small current flows in exactly the opposite of the wind direction.



Surface Wind and Current

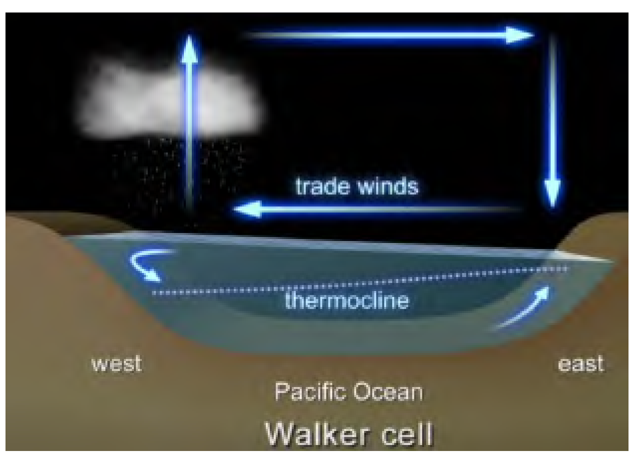
Thus, a strong surface wind sets off currents in all directions, decreasing in speed as they diverge from the wind direction. The average current is called the *transport*, and is shown by the heavy dashed arrow at 90° from the wind. This demonstrates an important result: a steady prevailing wind sets up currents in all directions to the right of the wind, but the predominant current is at right angles to the wind vector.

The chart below shows the prevailing sea surface currents in the northern hemisphere. The pattern described above does not fit precisely, but you can see it between California and Hawaii, where an easterly current accompanies a northerly wind.



The Pacific Tilt

The Pacific Tilt arises from the easterly Pacific trade winds from California toward Southeast Asia. The wind pushes the water to the west, creating a higher water level on the Asian side; the difference is about 1½ feet. At the same time, the thermoincline deepens on the Asian coasts. The result is a *Walker Cell*.



The Pacific Tilt's effect is analogous to the tropopause tilt in northern latitudes that drives winds, only now we have a tilt that drives currents. As shown below, on any horizontal line across the Pacific the water is deeper and denser on the west (Asia) than on the east (California). The result is a subsurface current traveling eastward, accompanied by a surface current traveling westward. This is the main Pacific circulation.

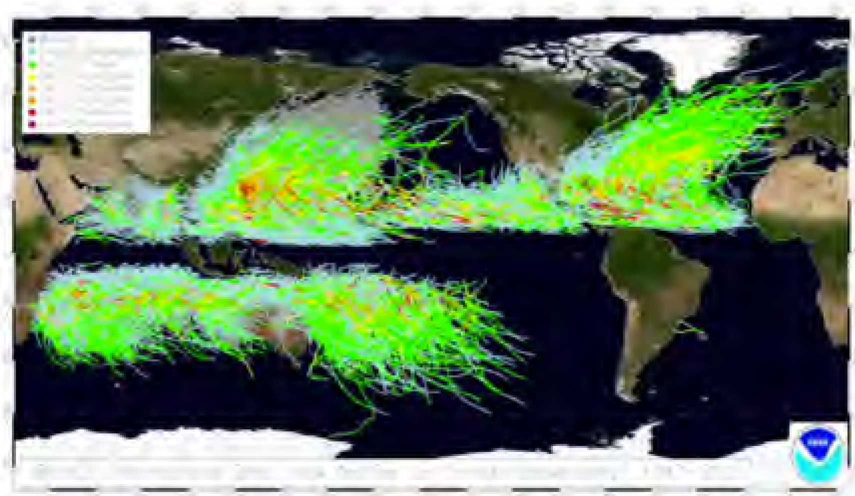
15. Tropical Cyclones: Hurricanes and Typhoons

We've looked at *extratropical* or *mid-latitude cyclones*—the thunderstorms and milder events common in the 30° – 60° latitudes. Now we turn to the tropical cyclones that originate in the lower latitudes and occasionally enter the middle latitudes or higher.

Development of a mature hurricane requires a number of atmospheric and oceanic characteristics. Chief among them are: there must be pre-existing thunderstorms to start the hurricane; the ocean temperature must be very warm, at least 26°C (79°F); there must be low vertical wind shear—high wind shear tears hurricanes apart. Unlike mid-latitude cyclones hurricanes are not associated with collisions of warm and cold fronts.

<u>Overview</u>

Hurricanes are non-equatorial phenomenon, as shown below. Of the hundreds of hurricanes in the 61-year period, none have occurred near the equator. There are two reasons. First, equatorial water is not warm enough—the equator is a band of uniformly temperate seasonal temperatures. Second, the counterclockwise rotation in the northern hemisphere would become clockwise just across the equator, and this acts to spin hurricanes away from the equator.



Hurricane/Typhoon Paths, 1947-2007

Hurricanes are also most abundant in the northern hemisphere, where they stretch from eastern Asia to western Africa; in the southern hemisphere they are isolated to the western Pacific, avoiding South America and Africa. The western Atlantic hurricanes that follow the eastern U. S. coast and veer out to sea are unusual; that veer is associated with the clockwise rotation of the Bermuda High and with *beta drift* (discussed below). There is no equivalent near the Pacific High.

Hurricane intensity is measured on the *Saffir-Simpson Scale* shown below. The most intense hurricanes—CAT 5—have sustained surface wind speed over 156mph and storm surges above 18 feet.[6]


Saffir–Simpson Hurricane Scale

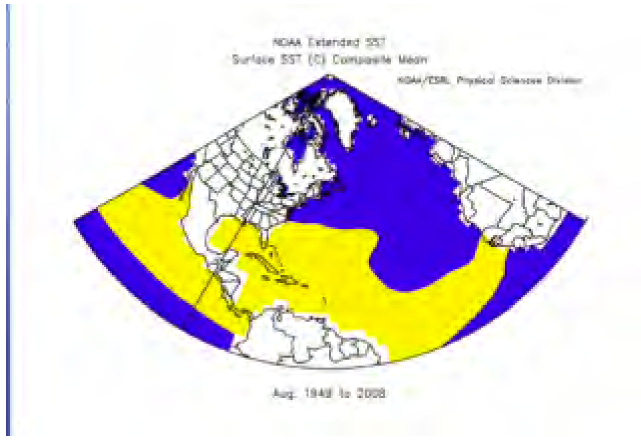
Category	Wind Speed mph (km/h)	Storm Surge ft (m)
5	≥ 156 (≥ 250)	> 18 (> 5.5)
4	131 – 155 (210 – 249)	13 – 18 (4.0 – 5.5)
3	111 – 130 (178 – 209)	9 – 12 (2.7 – 3.7)
2	96 – 110 (154 – 177)	6 – 8 (1.8 – 2.4)
1	74 – 95 (119 – 153)	4 – 5 (1.2 – 1.5)
Additional Classifications		
tropical storm	39 – 73 (63 – 117)	0 – 3 (0 – 0.9)
tropical depression	0 – 38 (0 – 62)	0 (0)

Hurricanes and Ocean Temperatures

The fuel for a hurricane is very warm and moist seawater drawn upward through evaporation and becoming buoyant with high CAPE. As noted above, a minimum temperature of about 79°F is needed to generate the energy required of a

[6] The scale measures wind speed at a height of 10 meters (33 feet) above the surface)

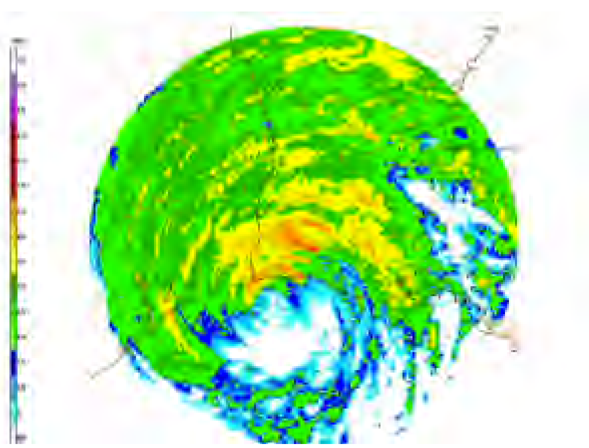
hurricane. This makes the hurricane a tropical phenomenon, as is shown in the chart below where yellow indicates the region with sea surface temperatures at least 79°F. The mid-Atlantic is the northern limit of the required temperatures.



Sea Surface Temperatures
(yellow above 79°C)

<u>Hurricane Anatomy</u>

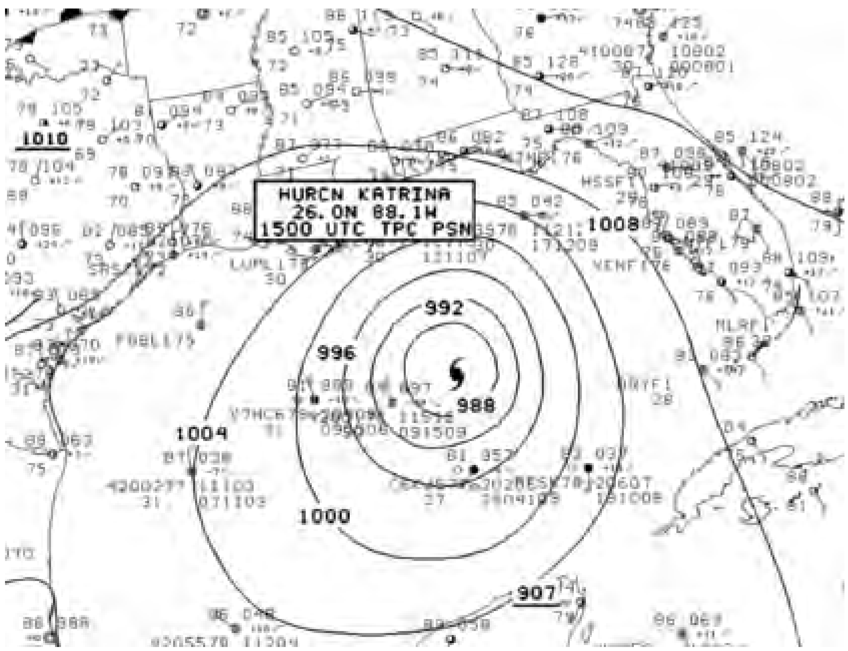
A radar echo of Hurricane Rita is shown below as it strikes the Texas-Louisiana coast in September, 2005, as a CAT 5 hurricane.



Hurricane Rita, September 2005

The warm colors are the areas of greatest moisture, and the green areas are the rain bands that encircle the outer portions of hurricanes.

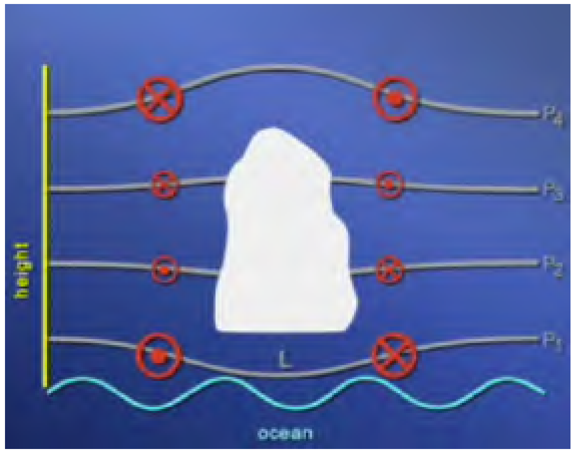
Below is a chart of Hurricane Katrina, a CAT 5 storm that devastated New Orleans in late August of 2005, only three weeks before the slightly stronger Rita headed for Houston.



Hurricane Katrina, 2005

Katrina's eye had an extremely low pressure trough of 988mb and pressure at her outer fringes at 1010mb. This steep pressure gradient created ferocious winds with high storm surge on its east side.

The formation of an early-stage hurricane is shown below. It begins with the remains of a thunder storm with counterclockwise rotation; this distorts the isobars: at the cloud base a downward bend in isobar P1 shows low pressure, while at the top of the cloud the upward bend indicates high pressure. Near the surface, positive vorticity is created and rotation is counterclockwise; the rotation is marked by the red circled bullet showing direction toward you and the red circled X showing direction away from you. The counterclockwise rotation at the cloud base draws warm surface air into the cloud creating an updraft. The air flows out cooler and dryer at the top, leaving energy behind as latent heat. The large size of the red circles at the base indicates a high wind speed.

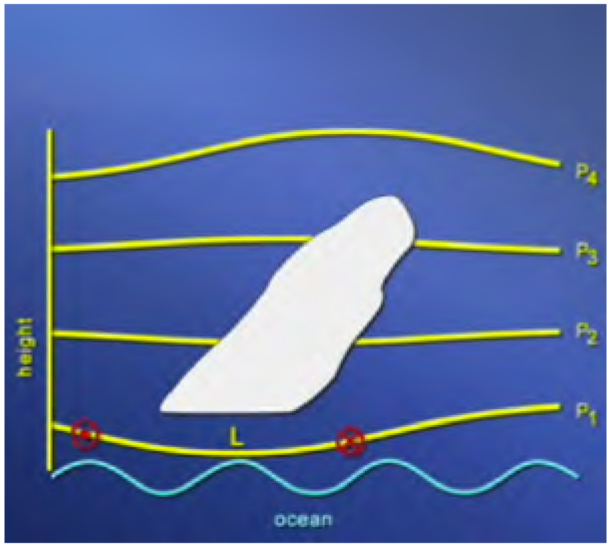


Early Rotation

As the altitude increases, the isobar distortions (pressure differentials) diminish up to the middle of the cloud column so wind speed falls, but above the mid-cloud altitude wind speeds increase and the rotation becomes clockwise. Thus, the usual rule applies: the high pressure at the top of the cloud has anti-cyclonic winds and the low pressure at the cloud bottom has cyclonic winds.

The winds are rotating in opposite directions at the top and bottom of the cloud. There is positive vorticity at lower altitudes with negative vorticity at higher altitudes. The lower air is rising and the upper air is falling, creating strong updrafts and downdrafts. A hurricane is building.

Recall that tropical cyclones, unlike mid-latitude cyclones, require low vertical wind shear. The reason is shown below. Vertical wind shear causes the early hurricane to tilt downwind. This reduces the lifting force that brings warm winds up and reduces the pressure differentials that start the counterclockwise rotation. The result is lower near-surface wind speeds, and less likelihood that the cell will become a hurricane.



Effect of Vertical Wind Shear

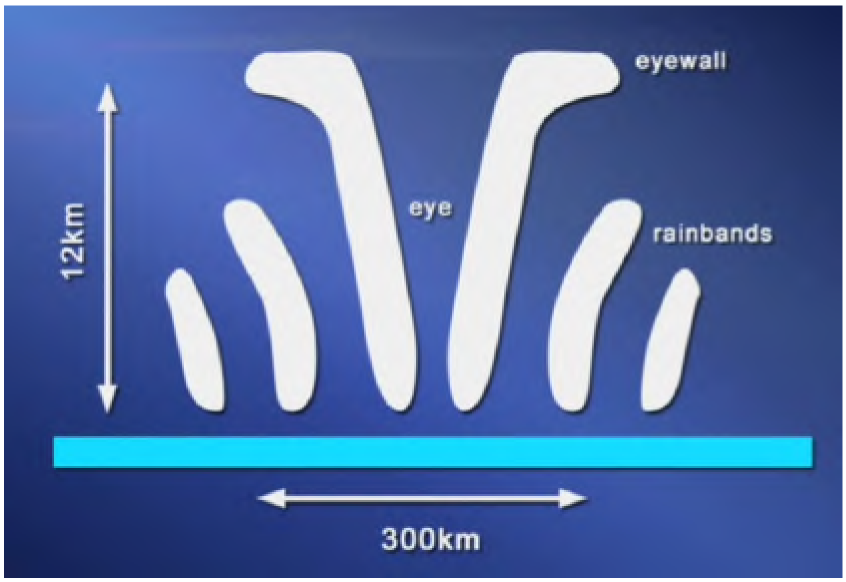
The figure below shows a vertical cross-section of a mature hurricane. At the outer edges are vertical walls of high precipitation called *rainbands*. At the center is the *eye*, flanked by a high cloud *eyewall* and topped with an anvil canopy. The eye is a center of very rapidly rotating highly buoyant warm moist air that warms even more as it rises until it compresses and condenses at the clockwise rotating canopy.

Note that the vertical eyewall and rainband clouds are narrow at the base and wide at the top, giving them an outward tilt. This is due to a *the law of conservation of angular momentum*.[7] As the radius of rotation increases the lair's inear velocity decreases in proportion: higher radius at the top means slower wind speeds. This is the same phenomenon observed when a figure skater pulls the arms in, creating a shorter spin radius and rotating faster.

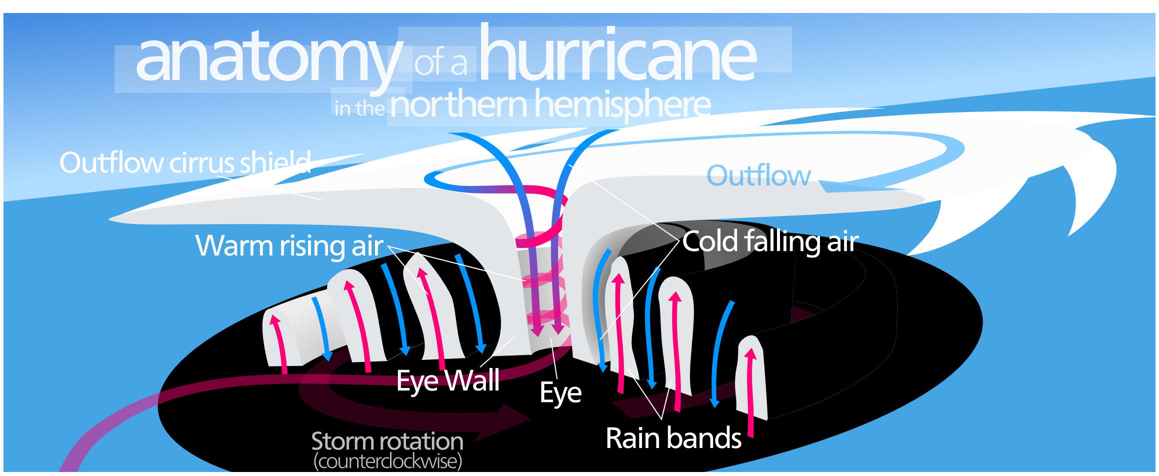
The last figure is an image of a mature hurricane. Warm air under the hurricane's canopy feeds the hurricane as it rises into the rainbands and eyewall, rapidly rotating and gaining CAPE as it rises in the eyewalls. Once at the hurricane canopy it condenses and falls as rain, most heavily between the rainbands. Note

[7] An object rotating in a circle has angular momentum measured as m**rv** where **r** is the radius of the circle, v is the linear velocity of the object, and m is the object's mass. Conservation of angular momentum means that a doubling of the radius implies a halving of the velocity.

that while the storm rotates counterclockwise around the low trough, the outflowing cold air at the high-pressure top is rotating clockwise. Thus, we sometimes see hurricane images as rotating clockwise, but that is the canopy, not the main storm.



Vertical Cross-Section of A Mature Hurricane



Anatomy of a Hurricane

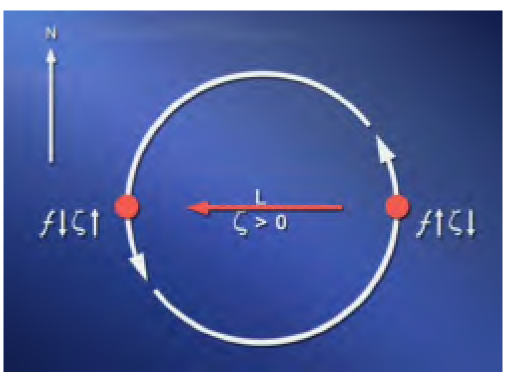
The end of a hurricane is typically near when it reaches cold water or land. The cold water reduces the temperature and moisture of the incoming air, starving the storm of fuel. Landfall also deprives it of fuel, but in addition the surface friction slows the storm's rotation and reduces its CAPE.

Beta Drift

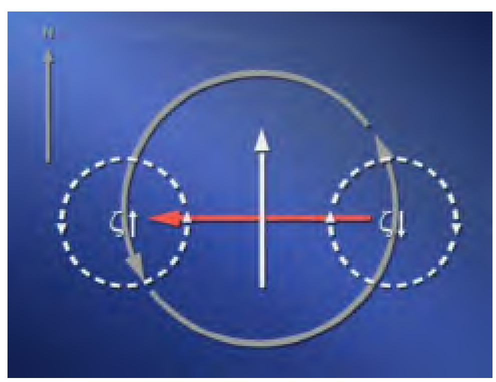
As shown above, Atlantic hurricanes develop in the tropics and move over the Caribbean Sea. Then they either enter the Gulf of Mexico, perhaps crossing land to go up the Atlantic coast, or go directly up the Atlantic coast. As the hurricane's latitude increases it tends to veer to the right, first to the northwest then eventually to the northeast. This rightward veer comes from several sources. Among them is the Bermuda High, whose clockwise rotation pulls the storm rightward. This is compounded by the Coriolis Effect and by *Beta Drift*.

Beta Drift is the result of positive vorticity that develops at the boundary of the hurricane. Recall that total vorticity is the sum of relative vorticity (rotation relative to the Earth's surface) and absolute vorticity (rotation of Earth relative to space). Absolute vorticity (denote **f**) is zero at the equator and highest at the poles. Relative vorticity (denoted ζ) depends on the position within the storm.

The interplay between absolute and relative vorticity determines the storm's track.



Effect of Absolute Vorticity



Effect of Relative Vorticity

In the charts above the hurricane's center has positive relative vorticity, a hurricane requirement. The left chart shows the directions of absolute and relative vorticity: absolute vorticity increases as a storm moves north (decreases as it moves south), and the storm has negative relative vorticity on its east side (positive vorticity on its west side).

The right chart shows the implications for the storm's path. Relative vorticity creates a high on the east side and a low on the west side, inducing an easterly cross wind (the red arrow). The storm is also moving north (it can't move southward into the "no hurricane" region), so its absolute vorticity becomes more positive, deepening the Low.

The combination of northward motion from absolute vorticity and westward motion due to relative vorticities causes the storm to move toward the northwest. This is the motion we see when the storm is a tropical depression coming across the Atlantic: it arrives from due east, then drifts northwest into higher lattitudes.

Increasing latitudes also increase the Coriolis Effect, which pushes the storm to the right. Ultimately, the Coriolis Effect pushes the hurricane into the Gulf of Mexico while it is still moving to the northwest, or into the Atlantic, where it continues to veer rightward as it parallels the east coast..

16. Light and Lightning

In a very real sense, light is what we see. This sounds crushingly trivial, but there is a deeper meaning: we do not always see what is really there: a rainbow is an example of an "optical illusion." In addition, regardless of the "true" physical characteristics of light, such as wavelength and path, what we see can be different for each of us: colorblindness is an extreme example,.

In this section we discuss some characteristics of light, then move on to the dramatic phenomenon of lightning.

The Nature of Light

Light is electromagnetic radiation in the visible spectrum as interpreted by the human eye. Our eyes have three basic retinal sensors called *retinal cones*: red cones, green cones, and blue cones. The red cones are sensitive to long-wavelength radiation, the green cones to middle-wavelength radiation, and the blue to short wavelength radiation. Blends of the three primary colors are what we see as the visible light spectrum.

The sunlight we receive is subject to both *differential scattering* (called *Rayleigh Scattering*) and *differential refraction* as it enters our atmosphere. The scattering occurs when photons hit objects like microparticles that form the condensation nuclei for water vapor. When the size of the particle is larger than the wavelength of the light, *Rayleigh Scattering* occurs and the light is reflected in all directions. This gives a tint to the sky at the wavelength of the reflected light. Thus, the clear mid-day sky appears blue because the blues are the visible wavelengths most reflected by a clear sky.

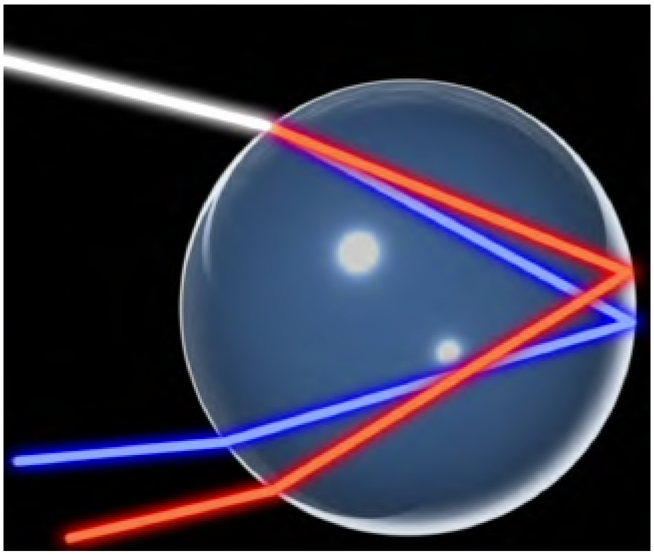
Differential refraction occurs because the path of light is bent when it enters a different medium—space to dry air, dry air to moist air, moist air to water, water to ice. The *index of refraction* measures the bending of light when it enters a medium: a vacuum has an index of 1.000, dry air has an index of 1.0003, water has an index of 1.333, ice has an index of 1.310. Refraction is greatest when the sun's light is entering the atmosphere at a low *angle of incidence*, or, what is the same thing, when the path through the atmosphere to the eye is longest: at noon the angle is highest and we see the sun as white or yellow because there is minimal refraction. But at dawn or dusk, when light's path is longest, we see red because longer wavelengths like the reds are more readily seen (we have an abundance of red cones). The shorter wavelengths—blue and greens—refract most thus disappear from view last, but we only see them briefly in a green or blue flash. The dawn or dusk sky's red tint is enhanced by Rayleigh scattering of light by air particles; this is greater the longer the light's path through the atmosphere.

When light is refracted, the bent rays alter our view of an object's position in the sky. Because refraction increases when wavelength shortens, we see reds as lower than greens, and greens as lower than blues. This creates a variety of optical illusions that we think of as physical facts. One of these is the setting sun's *green flash*, or, less frequently, *blue flash*. This occurs when the horizon is clear—as over an ocean—and its manifestation is a tint of green light on the sun's upper rim just as it completes its descent into the ocean. It varies in intensity from a slight tint to a strong green dot that might be called a flash, but it very rarely lights the sky—it only lights the sun's last edge.

There is a lot of myth surrounding the green flash—some believe it is a sign of upcoming good luck, some believe that is a true myth, some think it is a flash rather than a subtle color shift. But it is real—I've seen it many times on boats—and its explanation is simple. Green light refracts more than red light, so when the sun is at its lowest on the horizon its red light has disappeared below the horizon, leaving the shorter wavelengths like green or blue to the eye; stated differently, we see green light as higher in the sky—hence the last to disappear—because it refracts more than red light. It should be a blue flash, but a blue flash is rare because blue light is easily scattered and lost in the sky's dark blue tint at sunset.

Rainbows are another phenomenon. When white sunlight passes through moist air it is refracted, separating the red, green and blue lights. As shown below, when light strikes a drop of water the red wavelength refracts least, so it arrives at the back of the droplet as red on top of blue. But the reflection of the light to our eyes creates another refraction as light exits the drop and moves toward our eyes.

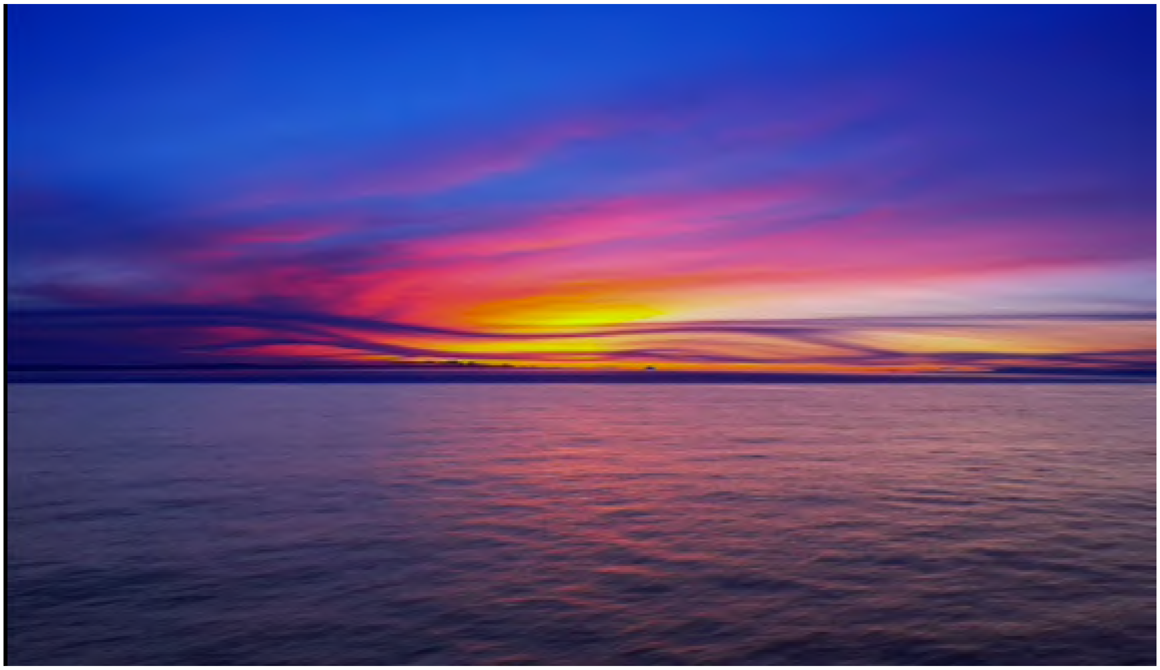
This second refraction increases the angle at which red light strikes our eyes, making the red appear to be on top of the blue even though it all started as the same light ray. What we see as a tricolored band of light is a phenomenon of refraction, not separate light bands.



Seeing a Rainbow

The *Aurora Borealis* is a much-admired light show common to the far north: it is seen as waves of undulating green light wafting over the arctic circle. The light show occurs at night, so it is not a visual manifestation of sunlight. Rather, it is an electrical phenomenon from the sun's cosmic ray's striking the Earth's magnetic field, which is strongest at the poles. This induces an electric current that excites the air particles, causing them to vibrate and emit radiation in the green wavelength. The undulating motion is due to variations in the intensity of the rays, in the intensity of the magnetic field, and in the density of the air.

Another light phenomenon is the deep blue sky often seen during and after sunsets, as seen below. This is not refraction. Rather, it occurs because stratospheric ozone (O3) absorbs the longer wavelength of light, leaving the blue wavelength dominant.



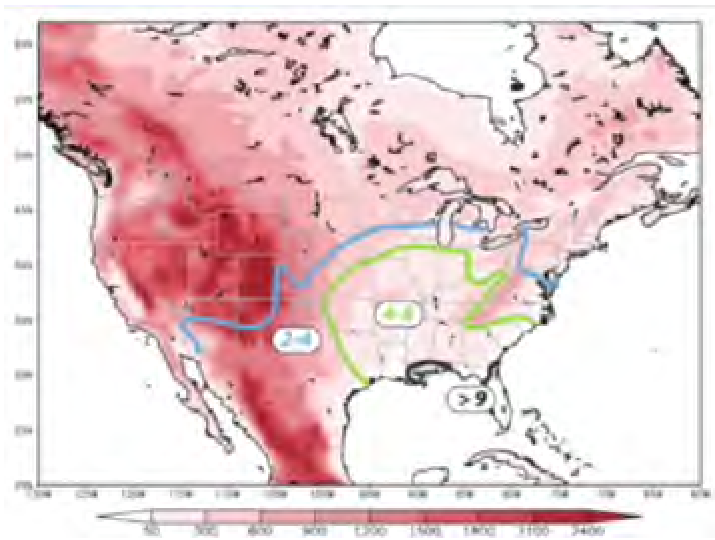
A Magnificent Sunset

Yet another phenomenon is haze—a blueish cloud-like tinting of the sky. The Smokey Mountains are an example of *wet haze*, in which sunlight is scattered by the condensation nuclei of water vapor. Over the ocean we see *dry haze*, in which salt particles provide the condensation nuclei that scatter the sunlight. Whatever the source, the scattering of the light creates a fog-like blue-white cast to the sky.

Lightning and Thunder

Lightning occurs when a *charge separation* between positive and negative ions (electrons stripped from their atoms) builds up to overcome the very low conductivity of air. A path develops between the charged areas and ions of opposite charge collide and mutually annihilate. The ensuing thunder comes from the superheating of the lightning's pathway forcing air molecules apart so rapidly that they exceed the speed of sound and run into each other. Thunder is simply a primitive sonic boom.

As seen below, lightning strikes are most frequent (over 9 per day) in the Gulf of Mexico and Florida, where thunderstorms are frequent. They are less frequent (4-8 per day) in the abutting southern states, rare (2-4 per day) in the eastern four-corner states, and very rare west of the Rockies. Clearly, they are associated with tropical weather.

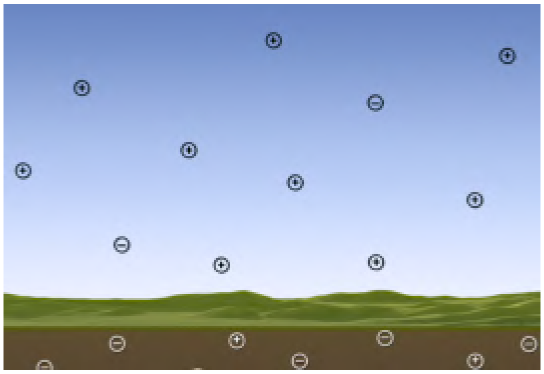


Lightning Strike Frequency (per day)

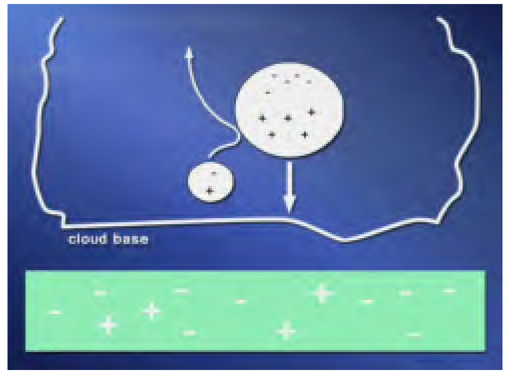
<u>Charge Separations and Lightning Strokes</u>

The charge separation that gives rise to lightning is a quick event, but it starts with a chronic low-energy charge separation between earth and sky. The earth has an abundance of negative ions from previous lightning strikes, leaving the atmosphere with an abundance of positive ions. Lightning comes when this small natural separation is increased by storm activity, reaching the point where nature tries to restore balance.

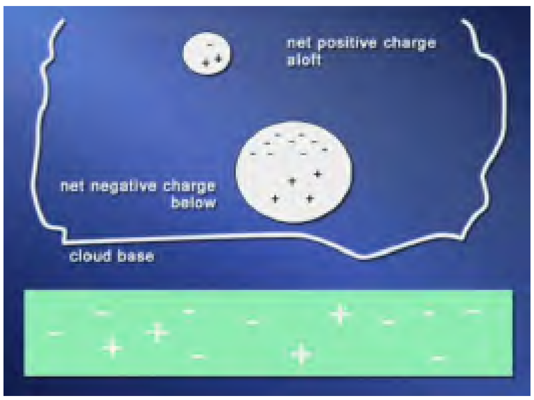
The process is shown in the four figures below. F1gure 1 shows the natural earth-atmosphere charging separation with more negative ions in the ground and more positive ions in the sky. This is called a *fair weather electric field*.



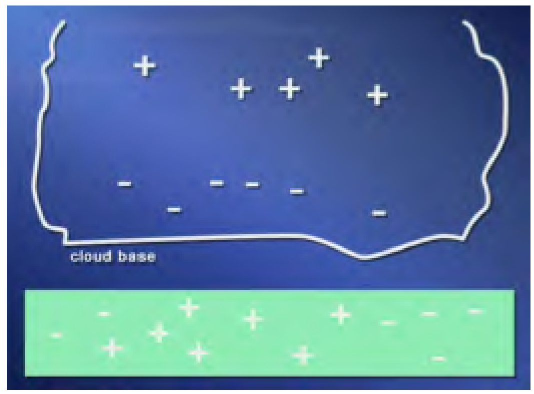
1. Fair Weather Electric Field



2. Stage 1 Electrical Induction



3. Stage 2 Electrical Induction

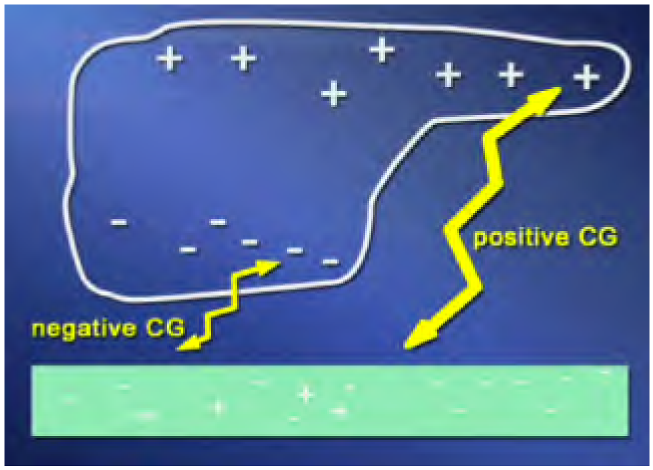


4. Stage 4 Electrical Induction

In Figure 2 a thundercloud has arrived with water vapor condensed around both small nuclei (the small ball, a water droplet) and large nuclei (the large ball, perhaps hail). Both are positively charged at the bottom because the attraction of the negatively charged ground pulls positive ions toward the ground and repels negative ions. The water droplet is lighter and rises in the updraft of warm air; the hail is heavier and falls. When they collide there is a transfer of charge and the water droplet gains some positive ions from the hail; the water now has a net positive charge, and the hail has a net negative charge, as seen in Figure 3. The surface has lost a negative ion to the cloud but remains negatively charged.

Figure four shows the final stage of induction. The ground has lost enough negative ions to the cloud that the cloud base is negatively charged and the ground is positively charged. When the voltage difference in this charge separation becomes sufficiently high to overcome air's low conductivity there will be a cloud-ga lightning strike between cloud base and ground: cloud-to-ground (negative CG) lightning.

If the charge separation is particularly great we might have an exchange of ions between the upper cloud and the ground (positive CG lightning), as seen below.



Positive Cloud-Ground Lightning

We see the lightning as a single stroke, usually cloud toward ground. In fact it is two strokes, each consisting of two phases. In the first phase a *step leader* descends from the cloud—this is a tentative attempt by the cloud to exchange ions with the ground. If the ground accepts, there is a *return stroke* from ground to cloud; the return stroke is what we see as the lightning, and it is by far the most dangerous phase because it completes the circuit between cloud and ground.. Note that the exchange is positive ions rising to meet negative ions—the flow is from positive to negative.

The first return stroke has now eased a pathway for a second stroke. This starts with a *dart leader*, which is a second-stroke step leader, and culminates with a second return stroke. Because both strokes happen so quickly, and a downward leader precedes each, we see this as a single cloud-to-ground stroke.

We are familiar with a cloud-to-cloud lightning called *heat lightning* or *sheet lightning*. This is another visual illusion—actual cloud to cloud transfers are extremely rare, and what we see is normal cloud-ground lightning strokes in storms at such a great distance that we don't see the vertical strokes.

17. Weather Prediction

As we know, weather forecasts are subject to considerable error. This, of course, is not always true. In the early 1980s I spent a couple of days in Moab, Utah, a town noted for uranium mining and Arches National Park. The weather channel on Moab' television was automated, at least by the standards of the day: a camera was trained on an array of analog instruments showing wind speed, humidity, temperature, and barometric pressure. This 24/7 display showed absolutely no change—each needle stayed just where it started (except for temperature). There was no need to hire a meteorologist when weather never changed. Automation is an excellent way to deal with routine events that won't challenge the brain!

Reliability of Weather Forecasts

But in areas where meteorological conditions are dynamic, weather predictions can be far off base. Even so, they have improved dramatically, as the chart below demonstrates.

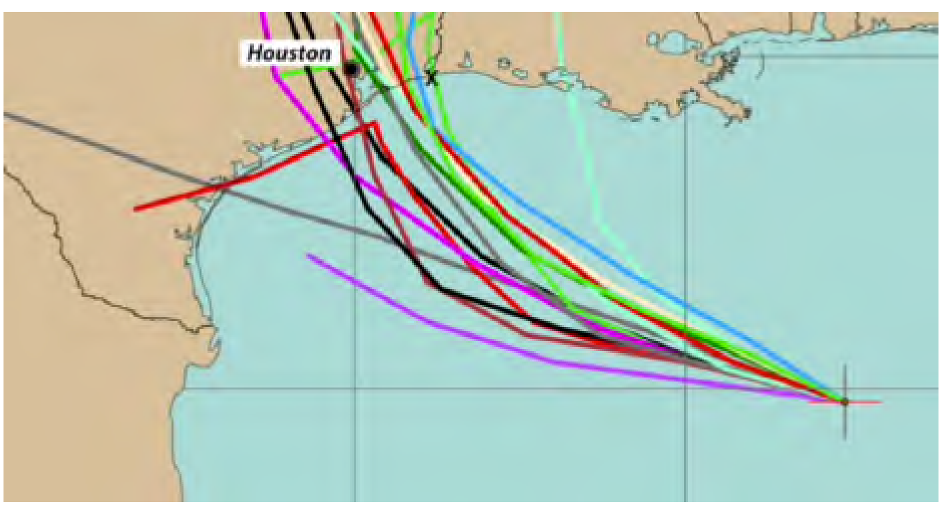
Average Forecast Errors of Hurricane Tracks, 1970-2008

The chart shows the average hurricane track errors (measured as percentage error in landfall location in nautical miles) for one-day, two-day, three-day, and five-day forecast horizons. Clearly, the shorter the horizon, the better the forecast. But what is most interesting is that at all horizons there is a strong decline in average forecast errors: we have greatly improved our ability to forecast hurricane tracks into the near future, and the improvements are greater the further ahead is the forecast.

This result is due to the improved technology of weather forecasting. Forecasting models have become far more complex, including hundreds of meteorological conditions; in the early 1960's models had few variables. This allows forecasters to consider a wide variety of scenarios and interactions.

In addition, computing technology has advanced remarkably. Today's supercomputers can crunch more numbers at faster-than-ever speeds. Not only does this allow more frequent forecasts, it also has allowed the implementation of increasingly rich models and the use of iterative simulation methods.

A happy example of hurricane track forecasting error is Hurricane Rita's landfall. Below we see the tracks predicted by a dozen forecasting models.



Hurricane Rita Track Forecasts

Hurricane Rita was a CAT 5 hurricane that was even more intense than Hurricane Katrina (also CAT 5) that struck New Orleans three weeks earlier. Rita's average forecasted track passed over Galveston and Houston, where its 180mph winds would do untold damage. While some tracks were well away from Houston, the model average predicted another New Orleans-level disaster!

But instead Rita made landfall on the Texas-Louisiana border, at the spot marked X. This was only 100 miles from Galveston, but it was a very important 100 miles. Even so, the forecast would be considered fairly accurate—as we will see, 100 miles is a very short distance for hurricane forecasting.

The Structure of Forecasting Models

Every forecasting model shares the same general structure. First, it requires *data to provide the initial conditions for the forecast* (temperature, humidity, cloud formation, pressure gradients, and so on). These are real numbers that the computer takes as the starting values for a simulation; the better the input data, the better the forecast, as the acronym GIGO (Garbage In, Garbage Out) suggests. The more accurate the input data, and the broader the range of conditions measured, the more refined and (in principle) more accurate the forecast will be.

Second, every weather system evolves as meteorological conditions change and interact. A forecasting model requires information on that evolution. This is in the form of *differential equations that capture the laws of nature*. When solid laws of motion and thermodynamics are not known, the forecaster can make up a "reasonable" representation (this is called *parameterization*).

So now you have your data (starting values) and your equations. Your model building is far from over. Now you have to make some important practical decisions. First is the *grid choice*, which determines the *model's resolution*. The grid is the fundamental distance between spots on your weather map.; it is the size of the weather cells that you will forecast. Conditions smaller than a single cell in the grid cannot be incorporated in your model; they are called *subgrid*; only larger cloud formations (for example) can be seen or simulated. The grid must be three-dimensional because the vertical displacement of weather is as important as the horizontal displacement: a cloud 100 meters high is different from a thundercloud one kilometer high. If you choose a grid with 100 miles horizontal and one mile vertical., you will have very low resolution and will miss important weather events, but you can quickly make a forecast. But a cube grid with, say, 100 feet on a side will have such high resolution that the forecasts should be more accurate but will take an inordinate amount of time; Perhaps they will even have more error because our knowledge of the laws of nature is not that refined.

So you want as small a grid as possible to avoid losing the details of the weather. But a small grid requires more input data (its grid should match) and requires far more numbers to be crunched, interactions to be measured, and time to simulate. High resolution is expensive, low resolution can miss important meteorological factors.

Once you have your initial conditions, equations, and grid you have another choice—the *step time*. A forecast is done in baby steps: at time t_0 you want to determine the weather at time $t_0 + \Delta t$: Δt is the *step time*. If you choose a very short step time, say one minute, then a 24-hour forecast requires 1,440 steps; the model must be run for 1,440 iterations before you have your 24-hour forecast. If you choose a 6-hour step time, it only needs to be run for four iterations. You have the grid choice problem in another form: smaller step times are better for the forecast quality because many of your equations will be wave-like, and small step times will capture the waves better than large steps. But small step times require more computing power and input data at higher frequency.

Now you are set to go. You start your model off by applying the initial values to your equations and computing the weather one step ahead. That provides your starting conditions for the second step ahead, and the two-step forecasts provide starting values for the third step, and so on. You will have a forecast for the horizon of your choice: 24-hours, 48-hours, forever. Simple, huh?

What Can Go Wrong: The Butterfly Effect

We can see how different models make different forecasts—the individual forecasters make different choices about starting conditions, equations, grid size, and step size. These differences might be traced to the budget, to the meteorologist's view-of-the-science, or to just about anything that differentiates between forecasters and the models they build.

One very important factor causing errors is that *the laws of nature—the equations of the model—are nonlinear*. For example, Z = aX +bY is a linear equation with the effects of X and Y on Z completely separated, but Z = aX +bY + cXY is nonlinear because the effect of X on Z depends on the value of Y.

Nonlinearity can create a number of problems. One is extreme *sensitivity to initial conditions*: Suppose you make a very small change in the initial value of a variable, say sea surface temperature, by starting sea surface temperature at 20.5°C instead of 20°C. In a linear equation this has little effect on the forecast of Z, but in a nonlinear equation the small difference in X is magnified because X is multiplied by Y; if Y is large enough, the small change in X makes for a large change in XY, hence a large change in initial conditions. This problem is magnified because almost all equations in physics are nonlinear in complicated ways.

The sensitivity to initial conditions is sometimes called The Butterfly Effect, because, the story goes, a small change (a butterfly in Tokyo flapping its wings faster) might create extreme consequences (a tropical storm in California). The butterfly effect is at the heart of *Chaos Theory*, a mathematical theory of how small initiating causes can have dramatic effects even in a deterministic system without random shocks. Chaos theory can explain how a small electrical disturbance in the heart can escalate to a fatal ventrical fibrillation,

So when nonlinearities exist, your choice of initial conditions can have a dramatic effect on the forecasts—a sea surface temperature difference of .5°C might, in a day or two, create a hurricane instead of a nice day at the beach This was first observed by Edward Lorentz who, in the early 1960s, simulated a weather system with just three variables, two of which interacted. The result was that a small change in one variable's initial value led to virtually no alteration of the forecast—well past the time you would have thought a difference would appear—but then, suddenly and out of the blue, the forecasts began to diverge dramatically. To his credit, Lorentz did not just reject the result as a fluke, he recognized that it arose from nonlinearity in his model and that it posed a very real question for forecasting.

<u>Dealing with Forecast Uncertainty</u>

The best way to deal with forecasting uncertainty is to improve forecasting models. This means getting better data (improving the initial conditions), improving understanding of weather dynamics (improving the equations), using smaller grids, and using smaller step times.

In the short term, however, accepting uncertainty and engaging in repeated forecasts can obtain improvements. We saw one example above—the use of several different models to forecast Hurricane Rita's track. No one model will be reliably more accurate than another, so the merging of different forecasts is useful in obtaining an average forecast, and in showing the dispersion among forecasts.

A similar approach is to use the same model repeatedly, each time changing initial conditions by a small amount. Each forecast will be different, and an average might be meaningful, but in addition you can gain information on the sensitivity to initial conditions and the range of possible results consistent with the same model. This is called a *Monte Carlo simulation*.

18. Summary

If anything, this course has demonstrated why the weather is so difficult to predict. Everything matters: the shape of the terrain and its heat distribution, the air's tendency to positive or negative buoyancy, the rotation of the earth, the pressure and density differences between air parcels and masses, the air's moisture content, the precise phase transitions of water vapor between ice, liquid and gas, the prevalence of greenhouse gases, the collisions of front–on and on and on.

But all of the complexity makes the weather fascinating, and I found this course just that. I understand the moving parts of weather much better, but I doubt my ability to put them all together into a coherent story about a particular weather event. That must come with practice and experience. Perhaps if I use the concepts found here in my cruising life I'll get that practice.

Robert Fovell is an outstanding lecturer. I Googled his UCLA website and found student reviews of his introductory course on meteorology, from which this material was drawn. He was widely praised as engaging, articulate, supportive, and generally outstanding. I'm not surprised.

We all face weather in all its varieties. This course will not change your everyday experience—your reaction to 15 foot seas will not be different just because you might better understand their genesis—but it will change your understanding and appreciation of it.

Glossary

Adiabatic	The process of heat exchange arising from compression or expansion of air rather than input or extraction of heat energy
Anabatic Wind	An "upslope wind" rising up a steep slope, expanding and becoming warmer as it rises and releases latent heat. See *Katabatic Wind* and *Froude Number*
Advection	Air motion due to differences in *temperature*, not *pressure*
Air Mass	A collection of air parcels covering a large area
Air Parcel	A self-contained area of air with temperature different from the ambient air temperature; a source of clouds
Anticyclone	An area with wind rotating clockwise, creating negative *buoyancy* and a High pressure dome
Buoyancy	The tendency of warm air, once above the *LCF*, to rise (positive buoyancy) and cold air below the *LCL* to fall (negative buoyancy)
Buys Law	Pronounced *Bwah Balloh*. A rule-of-thumb for determining the direction of a low pressure center: with the wind at your back, turn 30° toward your right. The Low is on your left, the High on your right
CAPE	Convective Available Potential Energy, the energy accumulated in clouds in the form of latent heat
Centripetal Force	The inward force associated with spinning objects
Condensation	The phase transition of water from vapor to liquid when a warm moist air parcel meets cold air; condensation releases latent heat
Condensation Nucleus	A particle like dust, soot, or salt around which condensation forms
Convection	Motion of air, primarily vertical, to equalize pressure
Coriolis Effect	Also called Coriolis Force; the effect of Earth's rotation on the path of wind or storms
Cyclone	An air mass with wind rotating counterclockwise, creating positive air *buoyancy* and a low pressure trough

Glossary

Diabatic	The process of heat exchange by direct transmission of energy from an external source
Dryline	A line of particularly dry air that is often associated with *supercell* formation
Fujita Scale	A scale used to measure the intensity of tornados (F0 to F5)
Evaporation	The phase transition of water from liquid to vapor by passing air over a moist air parcel; the vapor absorbs latent heat
Froude Number	In fluid dynamics, the resistance to a fluid's flow. A Froude Number (Fr) less than 1 is *subcritica*l, indicating little resistance and small turbulence; a Froude number greater than one is *supercritical*, indicating high resistance and much turbulence.
Geostrophic Wind	A wind in which the Pressure Gradient Force and Coriolis Force are balanced, causing the wind to travel along isobars. The Frictional Force and the Centripetal Force will cause the wind to deviate from geostrophic balance.
Ideal Gas Law	A physical law that relates air pressure to air temperature and air density
Index/Refraction	A measure of the amount by which light bends when entering a medium from a vacuum. The vacuum has refraction index 1.0, air has an index of 1.003, and water has an index of 1.33
Isentrope	A line on a weather chart showing locations of equal *potential temperature*
Isobar	A line on a weather chart showing locations of equal air *Pressure*
Isopressure	A line on a weather chart showing altitudes associated with equal *pressure levels*

Glossary

Isotach	A line on a weather chart showing locations of equal wind speed
Jet Stream	A Tube of fast moving rotating air that encircles the Earth at the upper latitudes of both the northern and southern hemispheres
Katabadic Wind	A "downslope wind" flowing rapidly down a steep slope with supercritical F*roude number*. The air compresses as it falls, creating *adiabatic* heating and releasing latent heat. An example is Southern California's strong and hot Santa Ana Winds. See *Anabatic Wind*.
Lapse Rate	The rate of decrease in temperature as altitude increases. There are several lapse rates: the Environmental Lapse Rate (ELR) is the lapse rate in the ambient air, the Dry Adiabatic Lapse Rate (DALR) is the lapse rate for a subsaturated air parcel; the Moist Adiabatic Lapse Rate (MALR) is the lapse rate for a saturated air parcel.
Latent Heat	The heat released or absorbed when water undergoes a *phase transition* (ice to liquid to vapor or the reverse)
LCL	The *Lifting Condensation Level*; the altitude at which temperature equals dew point and above which a cloud base develops.
LFC	The *Level of Free Convection*; the altitude at which air becomes buoyant and a cloud rises vertically
Millibar	A measure of pressure, denoted as mb (1 mb = .0145 psi)
Occlusion	A Front composed of mixed Cold and Warm fronts moving in the same direction
Pressure Gradient	The difference in air pressure between two locations (per Km of distance), a major determinant of wind velocity
Phase Transition	The transition of water between the three phases solid (ice), water (liquid) and vapor (gas). Phase transitions release or absorb latent heat

Glossary

Potential Temperature	The temperature that an air parcel would have if it were moved to an area of standard pressure (e.g. 500mb)
Pressure	The force placed on an object, measured in millibars (mb) or pounds-per-square-inch (psi)
Refraction	The bending of light rays when they enter a medium with a different index of refraction. Normally the angle of refraction is less than the angle of refraction
Relative Humidity	The ratio of vapor supply to vapor capacity; relative humidity ranges from 0 (no vapor suppy) to 100 (saturation, or vapor supply = vapor capacity)
Ridge	Local region of high pressure within a large area of low or high pressure
Saffir-Simpson Scale	The scale describing hurricane intensity (CAT 1 to CAT 5)
Saturation	The state when an air parcel's vapor supply equals its vapor capacity, making its Relative Humidity 100; the edge of a phase transition from vapor to liquid
Scattering	Also called *Rayleigh Scattering*. The reflection of light rays when they meet an object larger than the light's wavelength
SKEW-T Chart	A charts showing the relationship between temperature and pressure in an air mass
Subsaturation	The state when an air parcel's vapor supply is less than its vapor capacity, making its Relative Humidity less than 100
Temperature	The energy in the form of heat generated within a space due to the vibration of interacting air molecules; there are several definitions, among them *dry bulb*, *moist bulb* and *dew point*

Glossary

Thermal Inertia	The resistance of a medium to transfer of heat. Air has high thermal inertia, land has low thermal inertia
Thermoincline	A region of water depth in which temperature declines strongly with depth
Transport	The average direction of water current relative to surface wind direction, normally at 90° to the wind direction
Trough	A local region of low pressure within a large area of low or high pressure
Vapor Capacity	The maximum water vapor that a unit of air (cubic meter) can hold without becoming supersaturated
Vapor Supply	The actual water vapor per unit of air (cubic meter)
Vorticity	Also called *vorticality*. The creation of pressure differences due to spin. Counter-clockwise (cyclonic) spin creates positive vorticity and positive air buoyancy, resulting in low pressure near the surface. Clockwise (anticyclonic) spin creates negative vorticity and negative buoyancy, resulting in high pressure near the surface
Wind Shear	The difference in wind speeds at different altitudes (vertical wind shear) or at different locations (horizontal wind shear)