Fellow glider pilots with and interest in the weather,

I have written these notes to give some context to the slides from my Pilot's Night presentation, for those that were not able to attend in person and those that were the victims of a rather hurried presentation of the slides. I have taken some of this from the web, specifically Wikipedia and the American Meteorological Society Glossary. I have included a few URL's that I urge you to follow if you have an interest in knowing more. I have also included what I view to be a simple explanation of mountain wave and an additional section some information from the American Meteorological Society Glossary for those who fancy Greek symbols. Finally, I'm happy to answer any questions in person or via e-mail.

So here it is … warts and all.

Keith

Slide 1 Coriolis `Force'

(From Wikipedia)

The **Coriolis effect (force)** is the apparent deflection of objects from a straight path if the objects are viewed from a rotating frame of reference. One of the most notable examples is the deflection of winds moving along the surface of the Earth to the right of the direction of travel in the Northern hemisphere and to the left of the direction of travel in the Southern hemisphere. This effect is caused by the rotation of the Earth and is responsible for the direction of the rotation of cyclones. As a consequence, winds around the center of a cyclone rotate counterclockwise on the northern hemisphere and clockwise on the southern hemisphere.

http://en.wikipedia.org/wiki/Coriolis_force

This is a manifestation of conservation of angular momentum. As seen on the slide, if a parcel of air moves toward either of the poles, it must accelerate in the direction of the rotation of the earth. On the surface of the Earth, a rotating reference frame, this is seen as a deflection of parcels to the left in the southern hemisphere and to the right in the northern hemisphere. The Coriolis "force" is the fictitious force we use to explain this acceleration as we treat what is really a rotating reference frame as one that is fixed.

Slide 2-4 Geostrophic Wind

(From Wikipedia)

The **geostrophic wind** is defined as the wind resulting from what is called the *geostrophic balance* between the Coriolis force and the pressure gradient force acting on a parcel of air, causing the wind to blow parallel to isobars in the earth's atmosphere. However, this balance is rarely found exactly in nature, due to other forces acting on the wind, such as friction from the ground, or the centrifugal force from curved fluid flow. Thus, the isobars must be straight in pure geostrophic flow. Despite this, much of the atmosphere outside the tropics is close to geostrophic flow much of the time and it is a valuable first approximation.

Air naturally moves from areas of high pressure to areas of low pressure, due to the pressure gradient force. As soon as the air starts to move, however, the coriolis force deflects it due to the rotation of the earth. The deflection is to the right in the northern hemisphere, and to the left in the southern hemisphere. As the air moves from the high pressure area, its speed increases, and so does the deflection from the coriolis force. The deflection increases until the coriolis and pressure gradient forces are in geostrophic balance, at which point the air is no longer moving from high to low pressure, but instead moves along an isobar, a line of equal pressure (note that this explanation assumes that the atmosphere starts in a geostrophically unbalanced state and describes how such a state would evolve into a balanced flow. In practice, the flow is nearly always balanced. The geostrophic approximation has no predictive value since it does not contain any expression for change: it is purely diagnostic). The geostrophic balance helps to explain why low pressure systems spin counterclockwise and high pressure systems spin clockwise in the northern hemisphere (and the opposite in the southern hemisphere).

http://en.wikipedia.org/wiki/Geostrophic_wind

Slide 5 Surface Pressure

The geostrophic wind is shown by the red arrows. The real surface wind will have a cross-isobar component toward low pressure.

Slide 6 and 7 850 mb and 700 mb Charts

Height contours of constant pressure surfaces, for example the 850 mb pressure surface, are useful and can be interpreted in a way similar to the surface pressure chart. That is, the geostrophic wind (shown as red arrows) moves clockwise around a low pressure centre roughly parallel to the height contours. At upper levels like these, the wind will be nearly geostrophic as there is little, if any, frictional drag.

Slide 8 Parcel Concepts

Assume a parcel (3-D box) of air can move vertically within the atmosphere. Assume that no heat can move into or out of the box (adiabatic). The amount of moisture in the box remains constant – hence its dew point and mixing ratio (the amount of moisture in the air, normally measured in g/kg) remain constant in unsaturated ascent or descent. Once the parcel cools to its dew point, moisture can be removed from the box as it condenses as the parcel continues to cool past its dew point.

Slide 9 Dry and Moist Adiabatic Lapse Rates

By the ideal gas law, a parcel of air will warm if compressed and cool if expanded. Atmospheric pressure, which is due to the weight of the overlying column of air, decreases with height. If no heat is added to or removed from a parcel of unsaturated air (an adiabatic process), it will cool at 10 degree C / 1000m. Alternatively, if the parcel of air is saturated, moisture in the air parcel will condense as it cools. This condensation gives off sensible heat to the air parcel and acts to reduce the rate at which the air parcel cools as it rises. The rate at which a saturated parcel cools is closer to 6 degrees / 1000m near sea level and increases to approach the dry adiabatic lapse rate at somewhere near 400mb. At this level (temperature) there is very little moisture in the air to condense, hence the latent heating is negligible.

Slides 10 and 11 Static Stability and Instability

A neutrally buoyant parcel at rest will remain at rest unless disturbed. If an unsaturated parcel is displaced vertically upward/downward, it will cool/warm at the dry adiabatic lapse rate. If the environmental lapse rate is such that the parcel is cooler (more dense) than the surrounding air if it rises and warmer (less dense) than the surrounding air if it sinks, the local atmospheric column is referred to as statically stable. If the opposite is the case, the column is unstable. This is not a condition that can persist as the column of air would tend to overturn, mixing heat vertically to remove the instability.

Slide 12 Conditional Instability

One definition of conditional instability is one in which the environmental temperature profile is such that if a parcel is disturbed (moved up or down) from its equilibrium position, it will be stable if the parcel is unsaturated and unstable if the parcel is saturated. An example of this is where air forced to rise over topography is unsaturated at low levels but cools to its dew point as it rises. Given the same atmospheric temperature profile, the parcel is stable at lower levels where it is unsaturated and unstable at levels above the height at which it has cooled to its dew point.

Slides 13 and 14 Thermodynamic Diagrams (Skew-T and Tephigram)

The Skew-T and Tephigram are very similar. Both are area-preserving and therefore are true thermodynamic diagrams. Both contain the same isopleths (lines of constant pressure, temperature, saturated mixing ratio, etc.) but are plotted in slightly different coordinates. The examples given are plotted on a tephigram. Slide 14 identifies each of the lines on a Tephigram.

Slide 15 and 16 Parcel Ascent/Descent

A parcel ascent begins at the surface pressure, in the case shown that's around 1000 mb. This value will decrease with the elevation of the location where the sounding is released. The red line on the right is the temperature trace with the blue line on the left the dew point trace. Note that on the BOM Skew-T's, both the temperature and dew point are red, with yesterday's traces in blue.

A parcel lifted from the surface will cool at the dry adiabatic lapse rate or along a dry adiabat. As the parcel remains unsaturated below about 900 mb, the parcel will lose no moisture and therefore also follow a line of constant saturated mixing ratio on the Tephigram. These are the two black line segments that meet at about 900 mb. At that point the air has cooled to the point of saturation and will thereafter cool at the moist adiabatic lapse rate if lifted further. This is the curved black line above 900 mb. What we notice in this example is that the parcel is at all times cooler than the surrounding air and therefore more dense than the surrounding air. In other words this parcel would need to be mechanically lifted against negative buoyancy.

As the surface temperature rises, heat is transferred to the air through turbulent mixing. Through this process a shallow adiabatic or possibly super-adiabatic (temperature decreasing at slightly greater than the adiabatic lapse rate) develops from the surface. This is shown by the parallel dotted red lines at the base of Slide 16. As the temperature rises and erodes the surface based nocturnal inversion, the layer get deeper and deeper with buoyant plumes (thermals) reaching higher and higher. As in the previous slide, the saturated mixing ratio remains constant. However as parcels rise higher and higher with the increasing surface temperature, they eventually get high enough so that the dry adiabat along which they are cooling intersects the line of constant saturated mixing ratio. This is called the lifted condensation level (LCL). It's also roughly the cloud base. This is 800 mb in this example. Above this pressure (height) the parcel will cool at the most adiabatic lapse rate. In contrast to the previous slide, a parcel of air rising above the LCL will be positively buoyant (to the right of the environmental temperature trace) and will continue to rise. This is instability. The area between the moist adiabat and the temperature trace is the Convective Available Potential Energy (CAPE) and is proportional to the buoyant energy in a convective plume and therefore the thermal strength. Here the parcels will continue to rise, reasonably vigorously until encountering the inversion at 600 mb.

Slide 17 Clear Boundary Layer

In contrast to the previous slide, if the spread between the temperature and the dew point is quite large, no cloud may form (a blue day). As before, the adiabatic layer develops from the surface getting deeper and deeper as the surface temperature rises. Note that as thermals rising from the surface, they will only rise until they reach a level of neutral buoyancy (the intersection between the ground based dry adiabat and the environmental temperature trace). The increase in thermal height for an incremental increase in surface temperature will depend upon the shape of the environmental temperature trace. Here the tops of the thermals will increase in height from 800 mb to about 740 mb with only a single degree of temperature increase at the surface. Contrast this with earlier in the day when the surface temperature moved between 20 and 21 degrees C.

Slide 18 Ground Based Inversion

At the end of the day, the ground begins to cool due to long wave radiative heat loss. Like the growing boundary layer as the surface heats, an inversion grows from the surface up with the loss of heat from the layer next to the ground. As long wave radiative heat loss (say 100-200 watts/m^2) is much less than incoming short wave radiation (up to as much as 1000 watts/m^2) at mid-day, nocturnal surface based inversions grow more slowly and do not get as deep. Also, because the inversion layer tends to damp out turbulence, transport of heat into the ground from the air happens at a modest rate. If the surface temperature drops to the dew point, radiation fog often forms.

Slide 19 Conditional Instability

In this slide, a parcel that is moved slightly upward will cool at the dry adiabatic lapse rate. As such its temperature will remain to the left of the environmental temperature trace. This is a stable sate of affairs. However, if we have mechanical lifting such as air being forced up by topography this can change things markedly. In this example, the parcel reaches the LCL and begins to cool at the moist adiabatic lapse rate at about 850 mb. At this point it is still cooler than the surrounding air, but once raised to just above 700 mb its temperature is warmer than the surrounding air and the parcel is unstable. In the case shown, there is strong convection up to roughly 330 mb. Rapidly rising parcels would likely have significant vertical momentum at that level and over-shoot that level somewhat.

Slides 20 and 21 Saturated Layers

When the dew point and environment temperature trace are equal, the air is saturated. If this occurs through a deep enough layer, cloud will form. This can happen at low levels leading to the formation of boundary layer stratocumulus cloud, often broken to overcast. If this layer is elevated somewhat, perhaps caused by large scale lifting within an area of low pressure area, altocumulus is normally the result.

Slide 22 and 23 Convective Storms

When surface based convection leads to parcels with temperatures significantly warmer than the surrounding air (to the right of the environmental temperature trace) strong convection occurs. Parcels will continue to rise until they reach a point where they cool to the surrounding environmental temperature and have neutral buoyancy. In strong convection (thunderstorms), this normally occurs at the inversion that defines the top of the troposphere (the topopause) causing the anvil structure often observed with thunderstorms. As noted above, the strength of the convection is proportional to the area between the dry/moist adiabat and the environmental temperature trace. In principle, this area of available convective energy is matched by a similar negative area in the stable layer above the convection. This determines how far the parcels will over-shoot their level of neutral buoyancy as momentum carries them past the point of neutral buoyancy.

Slide 24 Example Tephigram: Saturday 17.03.07 Wagga

Canberra/Bunyan Weather: Moderate convection, scattered thunderstorms.

Slide 25 Example Tephigram: Sunday 18.03.07 Wagga

Canberra/Bunyan Weather: Low level cloud persisting most of the day.

Slide 26 Example Tephigram: Monday 19.03.07 Wagga

Canberra/Bunyan Weather: It just rained and thundered most of the day!

Slide 27 Solar and Terrestrial Radiation

All of the atmosphere's motions, vertical and horizontal are driven by the heating of the sun. Short wave radiation arrives at the top of the atmosphere and is variously reflected and absorbed as it makes its way to the surface. As clear air absorbs nearly no short wave radiation, on a clear, mid-summer day, the amount of radiation arriving at the Earth's surface can be as much as 1000 watts/m^2. This energy then acts to heat the ground and is stored there until it is transferred to the air via turbulent motions later in the day as the air begins to cool or is radiated upward overnight. Over time and a range of locations, any of these processes can dominate the surface energy balance. For example even on a very hot summer day, the air is heated very little over a body of water as nearly all of the incoming solar radiation goes to evaporate water from the surface and to a lesser degree warm the surface water. Alternatively, over a dry paddock, that has already been well warmed by the sun, nearly all of that energy will go to warming the air in contact with the ground through turbulent motions.

All objects radiate long wave radiation at a rate proportional to the fourth power of their temperature. The Earth's surface is no different. It is continually radiating upward. During the day, the balance between incoming short wave radiation and outgoing longwave radiation is dominated by the incoming short wave radiation. At night this balance reverses and the balance turns in favour of outgoing long wave radiation. On a clear night this energy goes straight back out into space from where it came. On a cloudy night, this radiation is absorbed by cloud and reradiated downward, keeping the temperatures warmer than they would be on a clear night.

Slide 28 Mountain Wave (A Simplified View)

A parcel disturbed vertically from rest in a stable atmosphere will move back to its original height via buoyant forcing. This happens as flow crosses a topographic feature as shown in Slide 28. As the parcel moves to the lee side of the topographic feature, it moves back to its original and then can overshoot its original height setting up periodic vertical movements. In a strongly stable atmosphere the force acting on the parcel to move it back to its equilibrium position is stronger than in a weakly stable atmosphere. The stronger the restoring force, the higher the natural frequency of the vertical oscillation. This natural frequency is called the Brunt-Vasallai frequency measured in cycles/unit time. If the speed of the air flow is constant, this leads naturally to a wavelength – the distance traveled in the time it take for the parcel to go through one cycle of its vertical displacement. As noted above, this natural frequency is determined by the thermal structure (stability) of the atmosphere. This is in turn somewhat dependent upon the moisture in the atmosphere – remembering what has already been said about conditional instability and such. If the length of the topographic feature matches, in some approximate way, the natural wavelength of the atmospheric oscillations, mountain wave can occur (right Paul?).

In addition, the wave can extract energy from shear in the flow – or more precisely, the vertical gradient of the shear. The shear and stability of the atmospheric column are summarised in the Scorer Parameter:

$$l^2(z) = \frac{N^2}{U^2} - \frac{\frac{\partial^2 U}{\partial z^2}}{U}$$

which is a standard parameter for assessing the likelihood of lee waves occurring. When the Scorer Parameter decreases strongly with height, the likelihood of lee waves is increased. The Scorer Parameter can be calculated using the same information that is used to plot standard atmospheric soundings such as the Skew-T or Tephigram. We should do this – it would be fun.

Slides 29 to 32 Mountain Wave (with Greek symbols)

(From American Meteorological Society Glossary of Meteorology)

mountain wave—An atmospheric gravity wave, formed when stable air flow passes over a mountain or mountain barrier.
Mountain waves are often standing or nearly so, at least to the extent that upstream environmental conditions (and diurnal forcing) are stationary. Two divisions of mountain wave are recognized, vertically propagating and trapped lee waves. Vertically propagating mountain waves over a barrier may have horizontal wavelengths of many tens of kilometers or more, usually extend upward into the lower stratosphere, and in pure form, tilt upwind with height. They can accompany foehn, chinook, or bora wind conditions. They have the capability to concentrate momentum on the lee slopes, sometimes in structures resembling a hydraulic jump, leading to occasionally violent downslope windstorms. When sufficient moisture is present in the upstream flow, vertically propagating mountain waves produce interesting cloud forms, including altocumulus standing lenticular (ACSL) and other foehn clouds. Intense waves can present a significant hazard to aviation by producing severe or even extreme clear air turbulence. Trapped lee waves generally have horizontal wavelengths of 5–35 km. They occur within or beneath a layer of high static stability and moderate wind speeds at low levels of the troposphere (the lowest 1–5 km) lying beneath a layer of low stability and strong winds in the middle and upper troposphere. These conditions are often diagnosed using a vertical profile of the Scorer parameter, a sharp decrease in midtroposphere indicating conditions favorable to trapped lee wave formation. Trapped lee waves assume the form of a series of waves running parallel to the ridges, and the crests of these waves often contain altocumulus, stratocumulus, wave clouds, or rotor clouds in parallel bands that can be very striking in satellite pictures. Because wave energy is trapped within the stable layer, these waves (and accompanying cloud bands) may dissipate only very slowly downwind, and they can continue downstream for many wavelengths spanning many tens of kilometers. Flow beneath the wave crests, occasionally made visible by rotor clouds, is often turbulent, thus

presenting a significant hazard to low-level aviation. Vertically propagating mountain waves and trapped lee waves can coexist, and sometimes lee waves are incompletely trapped or "leaky," leading to a variety of complex rotor interactions. This complexity of rotor patterns often produces interesting variations in cloud forms. As mountain waves propagate upward, the rotor's amplitude can grow to the point that the rotor "breaks," that is, the rotor becomes convectively unstable and overturns. Wave breaking can have an important role in vertically redistributing horizontal atmospheric momentum, as it slows the atmosphere by turbulent transport of the earth's momentum upward.

http://amsglossary.allenpress.com/glossary/browse?s=m&p=60

lee wave—1. Any wave disturbance that is caused by, and is therefore stationary with respect to, some barrier in the fluid flow. Whether the wave is a gravity wave, inertia wave, barotropic wave, etc., will depend on the structure of the fluid and the dimensions of the barrier. Most research has been devoted to the gravity lee wave (mountain wave) in the atmosphere, of wavelength of order

$$2\pi V\left[{T}\big/{g(\gamma_d-\gamma)}\right]^{\frac{1}{2}}$$

where V is the current speed, T the Kelvin temperature, g the acceleration of gravity, and γd and γ the dry-adiabatic and environmental lapse rates, respectively. This is the wave that is evident in lenticular or Moazagotl cloud systems and is strikingly exemplified in the Bishop wave. Dynamically, the lee wave is the sum of the free waves of the system and those wave components forced by the particular shape of the barrier. The disturbance is, in general, negligible at any distance upstream of the barrier, a result that follows from the dynamics when the system is started from rest, but a point that requires special attention when the steady-state assumption is made. The term lee wave is also applied loosely to nonwave disturbances in the lee of obstacles, such as the rotor cloud. 2. A mountain wave occurring to the lee of a mountain or mountain barrier. These waves can become visible in the form of lenticular or trapped lee-wave clouds.

Eliassen, A., and E. Kleinschmidt, 1957: Dynamic Meteorology. *Handbuch der Geophysik*, Vol. XLVIII, 59– 64.

http://amsglossary.allenpress.com/glossary/search?id=lee-wave1

Brunt–Väisälä frequency—1. The frequency N at which a displaced air parcel will oscillate when displaced vertically within a statically stable environment. Given as

$$N=\left(\frac{g}{T_v}\frac{\partial\theta_v}{\partial z}\right)^{\frac{1}{2}}$$

where g = 9.8 m s^{-1} is gravitational acceleration, Tv is the average absolute virtual temperature, and $\partial\theta v/\partial z$ is the vertical gradient of virtual potential temperature. Units are radians per second, although this is usually abbreviated as s^{-1}. This frequency is not defined in statically unstable air and is zero in statically neutral air. The frequency of internal gravity waves in the atmosphere cannot exceed the local Brunt–Väisälä frequency. This frequency is also sometimes used as a measure of the stability within a statically stable environment. 2. *See* buoyancy frequency.

Stull, R. B., 1995: *Meteorology Today for Scientists and Engineers*, 385 pp.

http://amsglossary.allenpress.com/glossary/search?id=brunt-vaisala-frequency1

Scorer parameter—The quantity *l*(*z*) arising from the wave equation for atmospheric gravity waves describing flow over a mountain barrier:

$$l^2(z)=\frac{N^2}{U^2}-\frac{\frac{\partial^2 U}{\partial z^2}}{U}$$

where $N = N(z)$ is the Brunt–Väisälä frequency and $U = U(z)$ is the vertical profile of the horizontal wind, both quantities determined from an atmospheric sounding upstream of the barrier.

The first term on the right-hand side usually dominates, but occasionally the second term, the velocity-profile curvature term, can be of similar magnitude. When l^2 is nearly constant with height, conditions are favorable for vertically propagating mountain waves. This parameter is most often used, however, as an indicator of when trapped lee waves (*see* mountain wave) can be expected; they occur when $l^2(z)$ decreases strongly with height. This is especially true if this decrease occurs suddenly in mid troposphere, dividing the troposphere into two regions, a lower layer of large $l^2(z)$ (high stability) and an upper layer of small $l^2(z)$ (low stability). *l*, the square root of the parameter, has units of wavenumber (inverse length), and the wavenumber of the resonant lee wave lies between *l* of the upper layer and *l* of the lower layer—the equivalent wavelength generally lying between 5 and 25 km in the atmosphere. Mountain ranges wide enough to force wavelengths long relative to *lupper* (the *l* in the upper layer) produce vertically propagating waves with wavenumbers less than *lupper*. Small objects (that force wavenumbers greater than *llower*) produce waves that are evanescent, or vanishing with height.

http://amsglossary.allenpress.com/glossary/search?id=scorer-parameter1

Slide 33 Pincher Creek, Alberta (the home of wind farms and mountain wave)

This is just a photo of wind turbines in Pincher Creek. Pincher Creek is just to the east of the Rocky Mountains in Southern Alberta, Canada.   Other than being quite picturesque, it is also an area where lee waves often develop.

Slide 34 Some Advice Regarding Prediction the Weather

This slide is mostly self-explanatory.

Slide 35 The Beginning

One might say this is the beginning of the required understanding of meteorology. It’s a complicated subject but a fairly interesting one.