Synoptic Meteorology I: Stability Analysis

For Further Reading

Most information contained within these lecture notes is drawn from Chapters 4 and 5 of "The Use of the Skew T, Log P Diagram in Analysis and Forecasting" by the Air Force Weather Agency, a PDF copy of which is available from the course website. Chapter 5 of *Weather Analysis* by D. Djurić provides further details about how stability may be assessed utilizing skew-T/ln-p diagrams.

Why Do We Care About Stability?

Simply put, we care about stability because it exerts a strong control on *buoyantly driven vertical motions* and thus (with ascent) cloud and precipitation formation on the synoptic and other scales. Contrast this with the mechanically driven vertical motions associated with divergence: those are *forced* motions, whereas buoyantly driven vertical motions are *unforced* and occur solely because of a difference in density between an air parcel and its surroundings.

Why else do we care? Rarely is the atmosphere *absolutely stable* or *absolutely unstable*, and thus we need to understand under what *conditions* the atmosphere is stable or unstable. We care about stability because the purpose of instability is to restore stability; atmospheric processes such as latent heat release act to consume the energy provided in the presence of instability. Before we can assess stability, however, we must first introduce a few additional concepts that we will later find beneficial, particularly when evaluating stability using skew-T diagrams.

Stability-Related Concepts

Convection Condensation Level

The convection condensation level, or CCL, is the height or isobaric level to which an air parcel, if sufficiently heated from below, will rise adiabatically until it becomes saturated. The attribute "if sufficiently heated from below" gives rise to the *convection* portion of the CCL. Sufficiently strong heating of the Earth's surface results in dry convection, which generates localized thermals that act to vertically transport energy. The attribute "until it becomes saturated" gives rise to the *condensation* portion of the CCL; at saturation, water vapor condenses.

To identify the CCL, read/interpolate the value of dew point temperature at a desired isobaric level. From this value, ascend parallel to a mixing ratio line until you intersect the observed temperature profile. The isobaric level or height at which this intersection occurs is the CCL. During the warmer months, the CCL can often be approximated from the base of the fair-weather cumulus clouds that dot the afternoon landscape on warm, sunny days.

The lines that we follow to obtain the CCL follow from the underlying physics. Sensible heating locally increases the temperature (but in and of itself does not change the dew point temperature), resulting in dry convection and localized thermals. The thermals vertically mix heat and moisture, conserving mixing ratio and potential temperature as they do so. The CCL is found by ascending along a mixing ratio line from the observed surface dew point temperature because thermals are rooted at the surface, sensible heating alone does not change surface dew point temperature, and mixing ratio is conserved for dry adiabatic ascent in a thermal. It is determined by the level where the mixing ratio line intersects the observed temperature trace since this level defines the potential temperature that a surface air parcel would have if a thermal vertically mixes to that level.

Convective Temperature

The convective temperature (T_c) is the surface temperature that must be reached for an air sample to reach its CCL. In other words, it is the surface temperature that must be reached for fair-weather cumulus clouds resulting from daytime heating of the surface layer to form. The units of convective temperature are °C or K. To identify the convective temperature, first identify the CCL for an air sample originating at the surface. Next, descend parallel to a dry adiabat until you reach the surface. The value of the isotherm intersecting this level gives you the convective temperature.

The process of identifying both CCL and convective temperature is illustrated in Fig. 19 on page 4-14 of "The Use of the Skew T, Log P Diagram in Analysis and Forecasting."

Lifting Condensation Level

The lifting condensation level (LCL) is the height or isobaric level at which an air sample becomes saturated when it is lifted dry adiabatically. Note the similarity to the CCL: both refer to a sample of air that ascends until it becomes saturated. The two differ in the cause of the ascent, however. Ascent to reach the CCL is assumed to result from heating of the underlying surface. Conversely, ascent to reach the LCL is assumed to be forced in nature – driven by synoptic-scale forcing for ascent, topography, local circulations, and so on.

To identify the LCL, first identify the dew point temperature at the desired isobaric level. Draw a line upward along the mixing ratio line that intersects this dew point temperature reading. Next, identify the temperature at the desired isobaric level. Draw a line upward along the dry adiabat that intersects this temperature reading until you intersect the line drawn upward along the mixing ratio line. The isobaric level or height at which this intersection occurs is the LCL. The process of identifying the LCL is illustrated in Fig. 19 on page 4-14 of "The Use of the Skew T, Log P Diagram in Analysis and Forecasting."

Unless the lapse rate between the surface and the CCL is superadiabatic, the LCL will always be found at or below (closer to the surface than) the CCL. Why this is so can be understood from the processes used to identify each on a skew-*T* diagram. Both involve ascending parallel to a mixing ratio line from an observed dew point temperature reading. The LCL also involves ascending parallel to a dry adiabat from an observed temperature reading, whereas the CCL makes use of the lapse rate (which is almost always equal to or less than dry adiabatic). Thus, the LCL is almost always found at or below the CCL. The two are found at equal altitudes when the lapse rate is exactly dry adiabatic between the surface and CCL/LCL.

The lines that we follow to obtain the LCL also follow from the underlying physics. Because we are forcing an air parcel upward, our consideration of the temperature trace begins at the parcel's origination level. For unsaturated conditions, forced lift does not result in heat exchange between an air parcel and its surroundings; thus, potential temperature is conserved as the air parcel ascends, and a dry adiabat is followed. (Note that this does not mean that there is no change in temperature on ascent – as an air parcel ascends to a lower pressure, it expands adiabatically, which results in its cooling.) The relevant physics associated with following a mixing ratio line upward from a dew point observation are identical to those as for the CCL, except that forced lift instead of thermal-induced lift causes the ascent.

Even when strong sensible heating is occurring, most lift from the surface is assumed to be forced rather than convectively driven, and thus the LCL is considered far more frequently than the CCL. Why is this the case? When a near-surface air parcel is locally warmer than its surroundings, it has a higher thickness, which means that its near-surface pressure is lower than its surroundings. Due to friction, this implies near-surface convergence; due to the relationship between divergence and vertical motion, this in turn implies lower-tropospheric ascent. Thus, the sensible heating that gives rise to thermals and the concept of the CCL also gives rise to microscale to mesoscale forced lift. (The LCL also has the advantage of being able to be evaluated from any origination level, whereas the CCL must be evaluated from the surface due to its reliance on heating from below.)

Mixed Layers

The planetary boundary layer, or PBL, is the lowest layer of the tropopause that is in contact with the Earth's surface. It is the layer over which the influence of the surface is directly transmitted to the atmosphere. The PBL is a turbulent layer characterized by small-scale turbulent eddies. There are two causes of such turbulence: buoyancy, driven by strong sensible heating of the underlying surface and resulting in dry convection, and vertical wind shear. Turbulent eddies within the PBL act to mix, or *homogenize*, specific attributes: momentum (wind), moisture (mixing ratio), and heat (potential temperature). Typically, the PBL is deepest in the local afternoon hours and shallowest in the local early morning hours.

To first approximation, mixed layers and their attributes, including their depth, can be identified using skew-T diagrams. The vertical layer extending upward from the surface over which mixing ratio and potential temperature are conserved – i.e., the observed dew point temperature trace is parallel to the mixing ratio lines and the observed temperature trace is parallel to the dry adiabats – reflects a mixed layer associated with the PBL. The altitude or isobaric level at which mixing ratio and potential temperature are no longer conserved provides an estimate for the depth and height of the mixed layer or PBL. Typically, an inversion will exist atop the PBL.

Level of Free Convection

The level of free convection, or LFC, is the height or isobaric level at which a sample of air that is lifted dry adiabatically until it becomes saturated (to its LCL) and then lifted saturated adiabatically thereafter first becomes warmer (or less dense) than the surrounding air. It represents the height at which convection becomes "free," signifying that the sample of air can freely rise solely because it is less dense than the surrounding air. We define this attribute as *positive buoyancy*, or of being *positively buoyant*. An air sample that reaches its LFC will continue to ascend freely until it again becomes colder (or denser) than the surrounding air.

Assuming that saturation occurs as a result of forced lifting (and not by heating), the LFC is found by first finding the LCL for a given air sample. From the LCL, ascend parallel to a saturated adiabat until it intersects and crosses to the *right* of the observed temperature trace. The level or height at which this first occurs is the LFC. The LFC is always at or above LCL height, and the process of identifying the LFC is illustrated in Fig. 22 on page 4-18 of "The Use of the Skew T, Log P Diagram in Analysis and Forecasting."

Equilibrium Level

The equilibrium level, or EL, is the level at which the temperature of a positively buoyant sample of air again becomes equal to (and then colder) than its surroundings. The EL is identified by first finding the LFC. From the LFC, continue to ascend parallel to a saturated adiabat until intersecting the observed temperature profile again, with the EL being the level or height of this intersection. The vertical distance between the LFC and EL provides a measure of the vertical depth over which ascent can occur if an air parcel can reach its LFC.

A Note About Air Samples

In the foregoing, we defined the LFC and LCL in terms of individual air samples that implicitly originate at the surface. The LCL and LFC need not be defined exclusively in terms of individual samples of air that originate at a single isobaric level, however; one can instead define them in terms of, for instance, mixed-layer–averaged air samples. In such a scenario, potential temperature and water-vapor mixing ratio values vertically averaged over a specified layer (e.g., the lowest 100 hPa above the surface) are used to define a mixed-layer parcel that can then be used to identify the LCL and LFC.

Furthermore, one may also define the LCL and LFC in terms of an air sample originating above the surface – e.g., one that may be found above an inversion. Not all ascent that leads to saturation, clouds, and possibly precipitation formation originates at the surface; it can and sometimes does originate above the surface. Clouds and precipitation that result from ascent originating at or very near the surface are known as *surface-based* phenomena whereas clouds and precipitation that result from ascent originating above the surface are known as *elevated* phenomena.

Assessing Stability

The basic premise underlying stability analysis is thus: if an air sample (or air parcel) is displaced upward from its starting location by some means (usually by forced ascent), will it continue to rise freely? Fundamentally, it will do so as long as its density is less than that of the surrounding air. To first approximation, this occurs so long as its temperature is warmer than that of the surrounding air. We thus now wish to examine how stability can be assessed in the context of skew-*T* diagrams.

Parcel Method

Consider a small upward displacement of an air parcel. If the air parcel is unsaturated, it will cool at the dry adiabatic lapse rate as it ascends. Conversely, if the air parcel is or becomes saturated, it will cool at the saturated (or moist) adiabatic lapse rate as it ascends.

If the air parcel's temperature is greater than the environmental temperature after the small upward displacement, the air parcel is said to be *positively buoyant*. In other words, it has a lower density than the surrounding air and will continue to ascend. This represents an *unstable* situation. As the air parcel ascends, it will cool at the dry adiabatic lapse rate until it becomes saturated and at the saturated adiabatic lapse rate thereafter. It will ascend until it is no longer positively buoyant. The level at which this occurs is the equilibrium level that we previously introduced.

Conversely, if the air parcel's temperature is smaller than that the environmental temperature after the small upward displacement, the air parcel is said to be *negatively buoyant*. This parcel thus has a greater density than the surrounding air and will descend back toward its initial position. This represents a *stable* situation.

When the air parcel's temperature after the small upward displacement is equal to the environment temperature, the air parcel is said to be *neutrally buoyant*. In other words, it has equal density to the surrounding air and it neither ascends nor descends.

Mathematically, the resulting stability criteria can be expressed in terms of the environmental, dry adiabatic, and saturated adiabatic lapse rates. Specifically,

	Unsaturated	**Saturated**
Stable	$\Gamma < \Gamma_d$	$\Gamma < \Gamma_s$
Neutral	$\Gamma = \Gamma_d$	$\Gamma = \Gamma_s$

Unstable	$\Gamma > \Gamma_d$	$\Gamma > \Gamma_s$

In the above, Γ_d and Γ_s represent the rates at which unsaturated and saturated air parcels, respectively, cool as they ascend. Γ represents the rate at which the *environmental* temperature (or that of the surrounding air) changes with increasing altitude above the ground.

For an unsaturated air parcel, if the surrounding air cools less rapidly with increasing altitude than does the air parcel (which cools at the dry adiabatic lapse rate Γ_d), the air parcel is said to be stable. If the surrounding air cools more rapidly with increasing altitude than does the air parcel, the air parcel is said to be unstable. The case where the air parcel and surrounding environment both cool at the dry adiabatic lapse rate represents neutral buoyancy or stability. The same principles apply for a saturated air parcel, except replacing Γ_d with Γ_s.

The following basic principles apply whether the air parcel is saturated or unsaturated:

- **Absolutely Stable:** The slowest rate at which an ascending air parcel can cool is the saturated adiabatic lapse rate. As a result, absolute stability is defined for an environmental temperature that decreases with increasing height at a rate less than the saturated adiabatic lapse rate, i.e., $\Gamma < \Gamma_s$. An inversion is a prime example of an absolutely stable situation.

- **Absolutely Unstable:** The greatest rate at which an ascending air parcel can cool is the dry adiabatic lapse rate. Thus, absolute instability is defined for an environmental temperature that decreases with increasing height at a rate greater than the dry adiabatic lapse rate, i.e., $\Gamma > \Gamma_d$. These are typically only found in vertical profiles taken during the warm-season, during the local afternoon or evening hours, and immediately above the surface.

- **Conditionally Stable/Unstable:** An ascending air parcel is said to be conditionally stable or conditionally unstable if the surrounding air cools at a rate between the saturated and dry adiabatic lapse rates, where $\Gamma_s < \Gamma < \Gamma_d$. The condition on stability/instability is whether the ascending air parcel is saturated (if saturated, unstable; if unsaturated, stable).

These concepts are illustrated in Fig. 1 below.

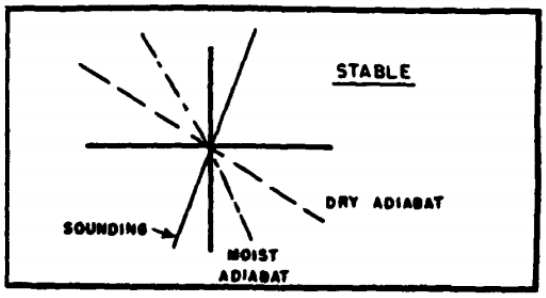

Figure 23a. Environmental (Sounding) Lapse Rate Less Than Lapse Rates of Dry and Moist Adiabats.

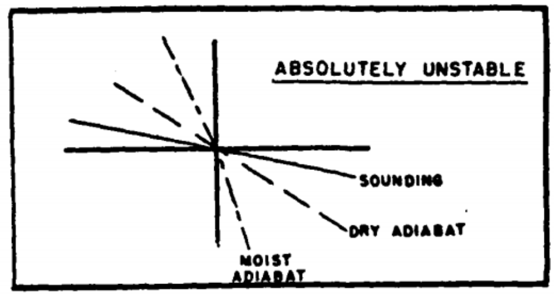

Figure 23b. Environmental (Sounding) Lapse Rate Greater Than Lapse Rates of Dry and Moist Adiabats.

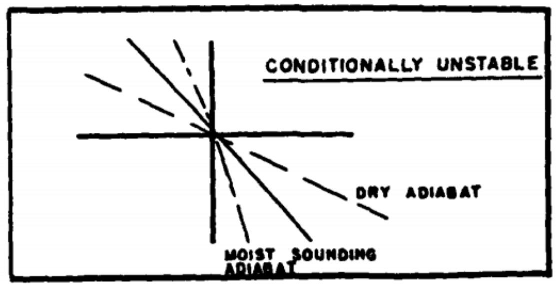

Figure 23c. Environmental (Sounding) Lapse Rate Less than Lapse Rate of Dry Adiabat, but Greater than Lapse Rate of Moist Adiabat

Figure 1. Examples of (a) absolutely stable, (b) absolutely unstable, and (c) conditionally stable or unstable temperature profiles. The observed temperature, representing the environmental lapse rate, is given by the slanted solid line. Dry adiabats, representing the dry adiabatic lapse rate, are given by the long-dashed lines. Saturated adiabats, representing the moist adiabatic lapse rate, are given by the short-long-dashed lines. Figure reproduced from Fig. 23 of "The Use of the Skew T, Log P Diagram in Analysis and Forecasting" by the Air Force Weather Agency.

Layer Method and Potential Instability

Instead of evaluating the stability of a single air parcel, the layer method seeks to evaluate the stability of an entire vertical layer or, more exactly, how the stability of a layer changes in response to ascent. This is done by considering two parcels, one each at the top and bottom of the layer, and evaluating how the lapse rate between them changes as the layer is lifted. As before, air parcels cool at the dry adiabatic lapse rate if they are unsaturated and cool at the saturated adiabatic lapse rate if they are or once they become saturated.

Let us first consider vertical layers that remain unsaturated upon ascent. As air parcels representing the top and bottom of such layers, ascend, they cool at the dry adiabatic lapse rate. We can illustrate this and its impacts to the stability of the layer with two examples: an initially superadiabatic lapse rate (i.e., absolutely unstable; Fig. 2) and an inversion (i.e., absolutely stable; Fig. 3).

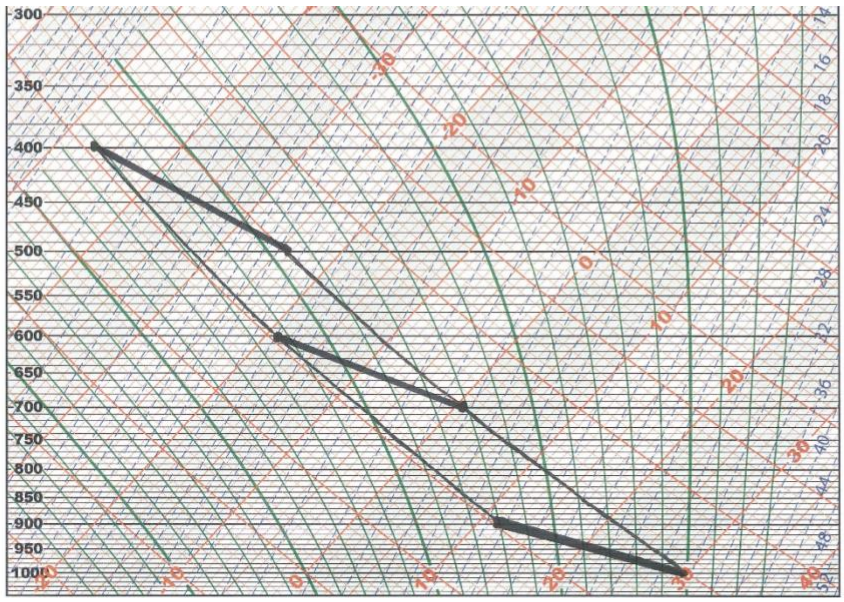

Figure 2. Skew-*T*/ln-*p* diagram depicting the change in lapse rate of an unsaturated, initially superadiabatic layer between 1000-900 hPa as it ascends to 700-600 hPa and, eventually, 500-400 hPa. The layer lapse rates are given by the thick black lines while the dry adiabats followed upon ascent are given by the thin sloped black lines. Dry adiabats are given by the negatively sloped orange lines while saturated adiabats are given by the negatively sloped green lines. Please refer to the text for further details.

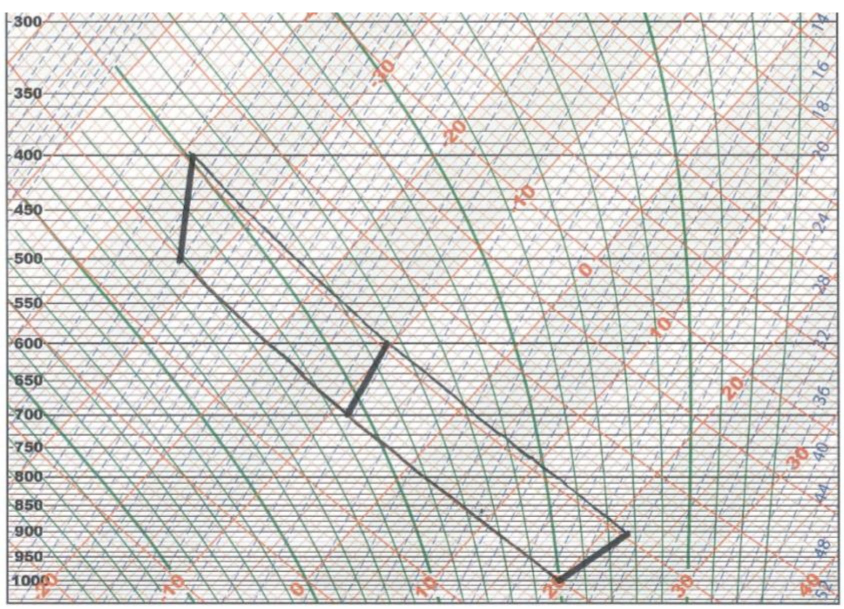

Figure 3. As in Fig. 2, except for the lifting of an inversion layer between 1000-900 hPa. Please refer to the text for further details.

In the superadiabatic lapse rate case, notice how the degree to which the lapse rate is superadiabatic decreases as the layer ascends, first from 1000-900 hPa to 700-600 hPa, then from 700-600 hPa to 500-400 hPa. Likewise, in the inversion case, note how the inversion weakens as it ascends, first from 1000-900 hPa to 700-600 hPa, then from 700-600 hPa to 500-400 hPa.

The former case represents a stabilizing scenario – we initially have $\Gamma >> \Gamma_d$ and end with $\Gamma > \Gamma_d$. The latter case represents a destabilizing scenario – we initially have $\Gamma << \Gamma_s$ and end with $\Gamma < \Gamma_s$. From this, we can state the following general stability criteria *so long as the entire layer is and remains unsaturated*:

- Lifting an initially unstable layer makes it less unstable.
- Lifting an initially stable layer makes it less stable.
- Lifting an initially neutrally buoyant layer does not change its stability.

Let us now consider a more complicated case, as depicted in Fig. 4. In this case, we begin with a stable situation since $\Gamma < \Gamma_s$ over the 1000-900 hPa layer. However, although we are not given dew point temperature data, we are told that the LCL for a parcel originating at A is at 950 hPa (or only 50 hPa above the air parcel's origination level) and the LCL for a parcel originating at B is at 730 hPa (or 170 hPa above the air parcel's origination level). Thus, the bottom of the layer is closer to saturation than is the top of the layer, which has an important influence on how the stability of the layer changes upon ascent.

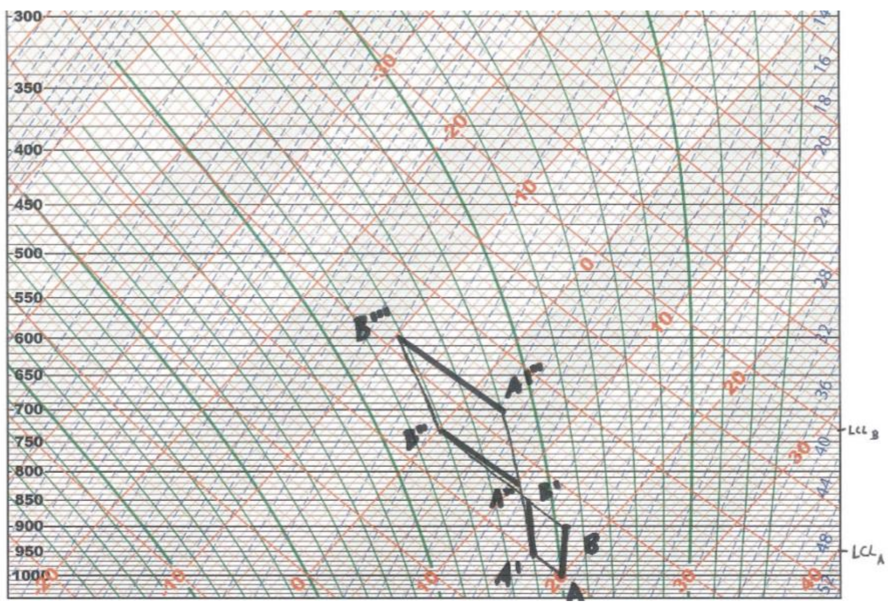

Figure 4. As in Fig. 2, except for the case where the layer's bottom is initially moister (with an LCL that is closer to the ground) than is the layer's top. Please refer to the text for further details.

Let us lift the 1000-900 hPa layer by 50 hPa to 950-850 hPa, or until the bottom of the layer becomes saturated. The layer's lapse rate after doing so is given by A' to B', which is less stable than that from A to B. Let us now lift this layer to 830-730 hPa, or until the top of the layer becomes saturated. Note that to do so, we follow a saturated adiabat upward from A' to 830 hPa (because the bottom of the layer has become saturated) and a dry adiabat upward from B' to 730 hPa (since the top of the layer is still unsaturated). The layer's lapse rate after doing so is given by A'' to B'', which is far less stable than that from either A to B or A' to B'. Indeed, because the top and bottom of the layer are both now saturated and the lapse rate of the layer exceeds the saturated adiabatic lapse rate, the layer has become absolutely unstable. This holds if we lift the layer further to 700-600 hPa, with the lapse rate then given by A''' to B'''.

How did such a drastic change in stability come about? From 950 hPa to 830 hPa, the bottom of the layer cooled at the saturated adiabatic lapse rate. Concurrently, from 850 hPa to 730 hPa, the top of the layer cooled at the dry adiabatic lapse rate. Thus, the top of the layer cooled more rapidly than the bottom of the layer, increasing the lapse rate within the layer.

In using the layer method to evaluate stability, you may have noticed that it follows much like the process for finding the equivalent potential temperature for air parcels at the top and bottom of the layer. In fact, all that is missing is ascending parallel to a saturated adiabat until it parallels a dry adiabat, then following parallel to that dry adiabat downward to 1000 hPa. However, one need not complete this process to qualitatively determine if one air parcel has a higher equivalent potential temperature than another. As the latter stages only involve paralleling saturated and dry adiabats, which are fixed entities on a skew-*T* diagram, the process we followed to determine how the layer's stability changed upon lift is all we need to determine whether the equivalent potential temperature of the top or bottom of the layer is higher. If both the top and bottom of the layer are saturated, the air parcel that is along the saturated adiabat that is *further to the right* has the higher equivalent potential temperature.

In the example in Fig. 4, both the top and bottom of the layer are saturated once reaching 830-730 hPa, or A'' to B''. Here, A'' lies along a saturated adiabat that is further to the right than B'', and thus A has a higher equivalent potential temperature than does B. As we documented that the layer destabilized upon lifting, we can infer that equivalent potential temperature decreasing with height represents a destabilizing scenario. Indeed, this is true. The full layer stability criteria are:

Stabilizing: $\frac{\partial \theta_e}{\partial z} > 0$ **Destabilizing:** $\frac{\partial \theta_e}{\partial z} < 0$ **No Change:** $\frac{\partial \theta_e}{\partial z} = 0$

Let us consider the stable case first. If θ_e increases with increasing height in a layer, it is typically moister (if not also warmer) at the top of the layer than at the bottom. This indicates that a parcel lifted from the top of the layer will reach its LCL quicker than will a parcel lifted from the bottom of the layer, and thus, a parcel lifted from the top of the layer will cool less rapidly with increasing height than will a parcel lifted from the bottom. If the bottom of the layer cools more rapidly than the top, the layer's lapse rate becomes smaller, or more stable, and the layer is not likely to continue to ascend.

Let us consider the unstable case. If θ_e decreases with increasing height in a layer, it is typically moister at the bottom of the layer than at the top. This indicates that a parcel lifted from the bottom of the layer will reach its LCL quicker than will a parcel lifted from the top of the layer. Thus, a parcel lifted from the bottom of the layer will cool less rapidly with increasing height than will a parcel lifted from the top of the layer. This causes the layer's lapse rate to become larger, or less stable, and the layer is likely to continue to ascend. This is the case in Fig. 3. Katafrontal environments are typically characterized by θ_e decreasing with increasing height in a layer as a result of the warm, moister air near the surface being overrun by cooler, drier air aloft.

The neutral case represents the case where parcels lifted from the top and bottom of a given layer reach their LCL after equal lifting. In this case, the lapse rate does not change, and the layer is no more or less likely to continue to ascend than before.

Taken together, these criteria are known as the *potential stability* criteria. Namely, they indicate that a layer has the *potential* to increase or decrease in stability *if* the layer is permitted to ascend.

Neglected Influences

The parcel and layer methods of stability assessment rely on many assumptions and simplifications that we have not yet explicitly stated. Generally speaking, these assumptions and simplifications do not change the qualitative stability evaluation, although they can and do change the quantitative stability evaluation. Here, we wish to briefly consider a couple of these neglected factors. A more complete list, including discussions of why we make such simplifications and the effect that doing so has upon the stability assessment, is provided on pages 5-2 and 5-3 of "The Use of the Skew T, Log P Diagram in Analysis and Forecasting."

To this point, we have stated that an air parcel will rise so long as it has lower density than its surroundings. In the context of the parcel method, we considered this explicitly in the context of temperature. However, because the gas constant R in the ideal gas law (which relates temperature and density) varies with the amount of water vapor in the air, the gas constant and temperature are typically replaced by the dry-air gas constant R_d and the virtual temperature T_v. Thus, density is a function of virtual temperature rather than temperature, and stability assessment should be done in terms of virtual temperature rather than air temperature. In practice, however, stability assessments conducted utilizing temperature produce similar qualitative results to those for virtual temperature.

To this point, we have also considered air parcels or layers that are completely isolated from their surroundings. Consider, for instance, an air parcel that ascends freely through a dry middle to upper troposphere. To this point, we have not considered the effect that the dry environmental air could have upon the parcel's ability to ascend or its stability. In fact, it can have a large detrimental effect upon the parcel's ability to ascend. Horizontal mixing (or *entrainment*) and/or vertical mixing of this dry air with the ascending air parcel can reduce the air parcel's water content and buoyancy.

In addition to entrainment, a full consideration of the vertical momentum equation indicates that buoyancy is not the only contributor to whether an air parcel will accelerate upward. Instead, there is also a vertical pressure gradient term and associated vertical pressure gradient force that controls the rate at which an air parcel moves vertically. The vertical pressure gradient force resulting from the decrease in atmospheric pressure with height acts to accelerate air parcels upward, but gravity counteracts this under hydrostatic balance. When hydrostatic balance does not hold, however, there may exist a perturbation vertical pressure gradient force due to either localized heating or dynamic effects. These processes are primarily of interest on the micro-to-mesoscales, however, and thus are beyond the scope of this class.

Furthermore, keep in mind that the environment changes with time as processes such as advection and diabatic heating occur. Although these generally occur on time scales longer than those of an ascending air parcel (hours vs. minutes), they can influence a stability assessment obtained from an observed skew-*T* trace that includes data from earlier times.

What About Descent?

When air parcels descend, they do so at the dry adiabatic lapse rate. Just as dry adiabatic ascent conserves potential temperature (with a parcel trace following a dry adiabat) and mixing ratio (with a parcel trace following a mixing ratio contour), so too does descent. Because this is so, even an initially saturated air parcel becomes unsaturated immediately upon beginning to descend, such that we need only consider dry adiabatic processes when evaluating descent.

Let us consider how descent impacts the stability of two layers, one that is initially superadiabatic and one that is initially characterized by a temperature inversion. These are depicted in Fig. 5, with both layers starting in the 500-400 hPa layer. As both layers descend to 700-600 hPa and then to 1000-900 hPa, note how both the superadiabatic and inversion layers become further exaggerated – i.e., even more superadiabatic and with an even stronger inversion, respectively. This is the opposite of what we saw upon lifting such layers.

What about a layer that is initially characterized by a saturated adiabatic lapse rate, or one that is stable if the layer is unsaturated and neutral if the layer is saturated? This is depicted in Fig. 6, starting again in the 500-400 hPa layer. As this layer descends to 700-600 hPa and then to 1000-900 hPa, its lapse rate becomes greater than the saturated adiabatic lapse rate.

For this case, it is important to note how the moist adiabats become more upright as you move toward warmer temperatures. This is because the effects of latent heat release upon the dry adiabatic lapse rate are greater when moisture content (controlled by temperature) is greater. As a result, though the slope of the line defining the lapse rate of the descending layer becomes less steep, this layer becomes less stable.

Because the slope of the moist adiabats changes as it does as one descends toward the surface, it is difficult to generalize these examples to a set of stability principles. However, the following three principles always hold and are applicable independent of the layer's initial saturation:

- Causing an initially unstable layer to descend makes it *more* unstable.
- Causing an initially stable layer (here, one with a lapse rate substantially less than the moist adiabatic lapse rate) to descend makes it *more* stable.
- Causing a layer over which potential temperature is constant (i.e., temperature trace follows a dry adiabat) to descend does not change its stability.

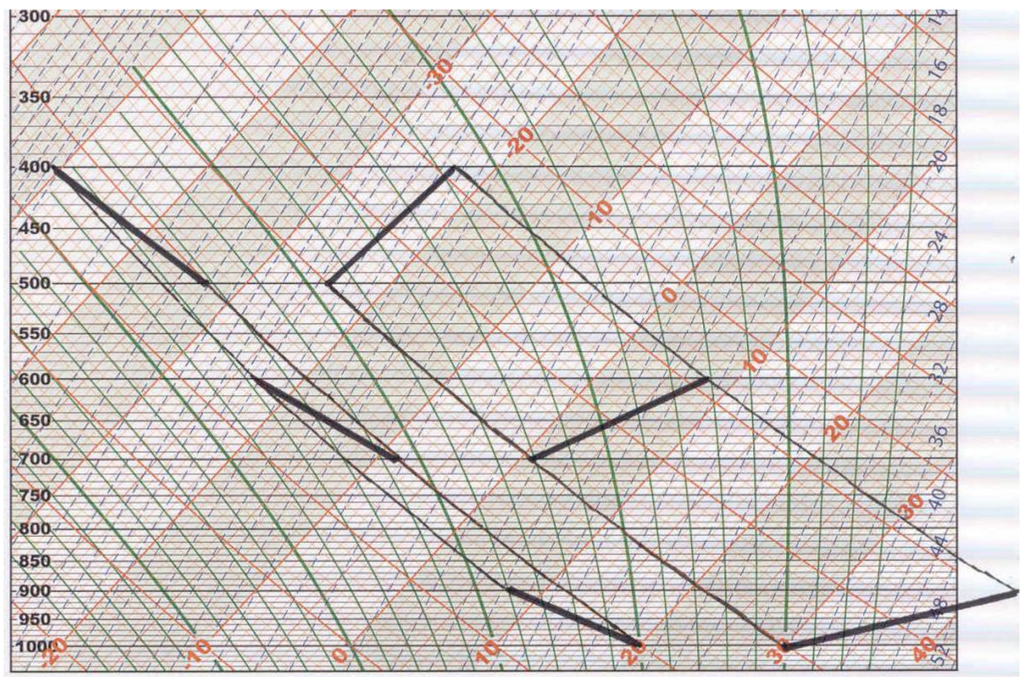

Figure 5. As in Figs. 2 and 3, except for the case in which an initially superadiabatic (at left) or inversion (at right) layer are forced to descend. Please see the text for details.

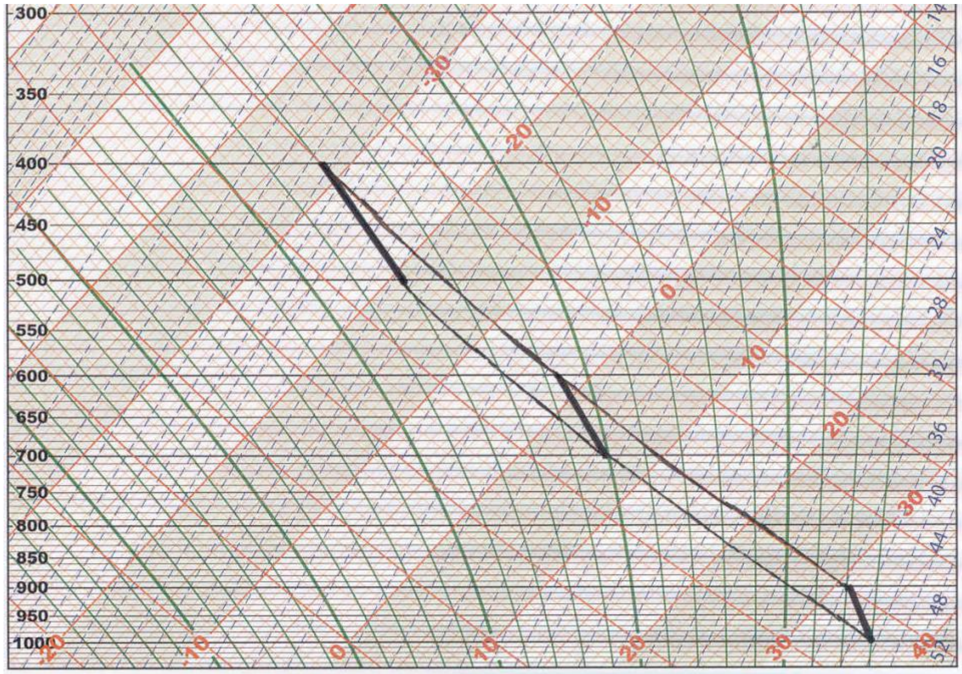

Figure 6. As in Fig. 5, except for a layer initially characterized by a lapse rate equal to the saturated adiabatic lapse rate. Please see the text for details.

Derived Stability Parameters

An air parcel only ascends if it is forced to do so (e.g., by a front, outflow boundary, terrain feature, or other such mechanism) or is positively buoyant. In the former, because the air parcel is cooler (and thus denser) than its surroundings, kinetic energy must be input to the parcel for it to ascend. The amount of kinetic energy that must be input for the air parcel to ascend is *convective inhibition*, or CIN. In the latter, because the air parcel is warmer (and thus less dense) than its surroundings, it may freely ascend. The amount of kinetic energy potentially available to this air parcel because it is warmer than its surroundings is *convective available potential energy*, or CAPE. CAPE can be viewed as a measure of an air parcel's buoyancy.

The amount of CIN and/or CAPE is directly proportional to the area on a skew-*T* diagram between two curves: the observed and parcel temperature traces. Both are typically expressed in units of J kg^{-1}. The area characterizing CIN is known as *negative area*, whereas the area characterizing CAPE is known as *positive area*.

Mathematical formulations for CAPE and CIN follow from these definitions, where:

$$E = -\int_{p_b}^{p_t} R_d \left(T_{vp} - T_{ve}\right) d \ln p \qquad (1)$$

In the above, E is energy (positive for CAPE, negative for CIN), R_d is the dry air gas constant, T_{vp} is the virtual temperature of the ascending parcel, T_{ve} is the environmental virtual temperature, p_b is the pressure at the bottom of the layer over which the integration occurs, and p_t is the pressure at the top of the layer over which the integration occurs. To first approximation, E is proportional to the difference between the virtual temperatures of the parcel and its environment. We often use temperature in place of virtual temperature for simplicity without significant qualitative impact.

For ascent due to surface heating, CAPE is evaluated between the CCL (p_b) and an EL (p_t). CIN is evaluated between the surface (p_b) and the CCL (p_t) as well as for any layers between an EL (p_b) and a subsequent return to a positive area (p_t). For forced ascent, CAPE is evaluated between the LFC for the selected lifted parcel (p_b) and an EL (p_t). CIN is evaluated between the level at which the parcel originates (p_b) and the LFC (p_t) as well as between an EL (p_b) and any LFCs that may exist above it (p_t).

The process of determining CAPE and CIN from skew-T diagrams varies depending upon whether a parcel ascends due to heating or due to forced lifting. As the former is less common than the latter, we start with ascent due to heating and then close with the more common forced ascent. In the following, we assume that we begin at the surface; however, CAPE and CIN can be evaluated under forced ascent for parcels beginning on any isobaric surface and for parcels representing the properties of a layer (e.g., a mixed layer).

Determination of CAPE and CIN for Ascent due to Heating

An illustration of how CAPE and CIN are determined from a skew-*T* diagram when the ascent is due to surface heating is given by Fig. 7. The process begins by finding the CCL. After identifying the CCL, continue upward along a saturated adiabat until the tropopause, identifying any ELs that are encountered in the process of doing so. (The example in Fig. 7 contains only one such EL.)

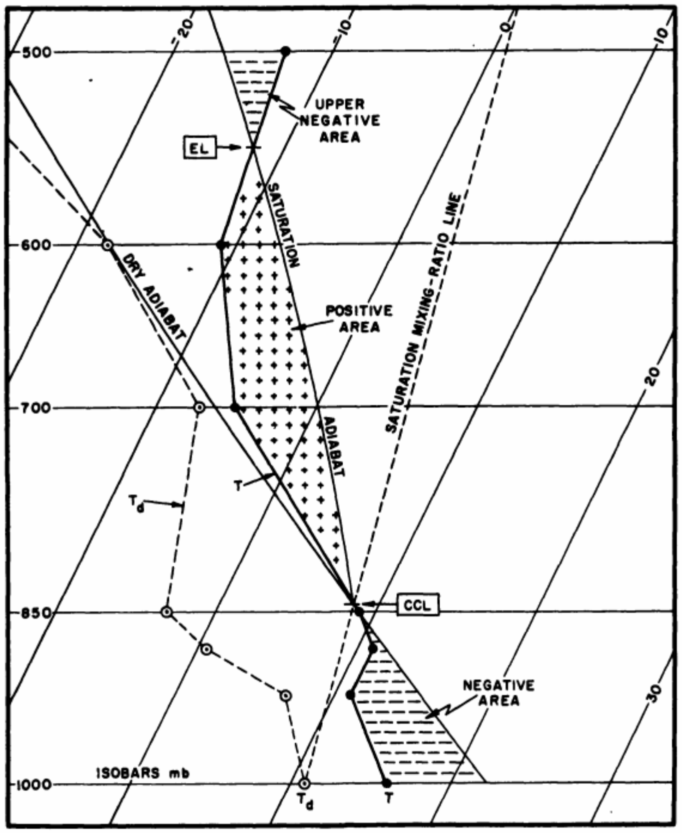

Figure 7. Illustration of how positive and negative areas are determined for an air parcel that ascends due to surface heating. Figure reproduced from Fig. 21 of "The Use of the Skew T, Log P Diagram in Analysis and Forecasting."

Below the CCL, the negative area is characterized by the area between the observed temperature trace and the dry adiabat followed downward from the CCL to the surface. This negative area represents the magnitude of warming (and thus kinetic energy input, as temperature is a measure of the average kinetic energy of the air) needed for a parcel to reach its convective temperature (T_c) and CCL. Stated differently, it represents the warming needed for an air parcel to become positively buoyant.

Above the CCL, we first encounter a positive area between 850-550 hPa. The greatest contribution to CAPE occurs when the parcel trace is furthest to the right of the observed temperature trace. An EL is found at 550 hPa, above which lies a negative area. Greater separation between the parcel and observed temperature traces above the EL results in a larger negative area and greater CIN.

Consider the case of Fig. 7. The negative area between the surface and the CCL represents the CIN that must be overcome by surface heating for an air parcel to reach its CCL. The positive area between the CCL and the EL represents the CAPE available to this air parcel. The EL at 550 hPa is an upper bound on the air parcel's ascent and the negative area above the EL represents CIN that must be overcome by some means for an air parcel to be able to ascend above 550 hPa.

Determination of CAPE and CIN for Ascent due to Forced Lifting

An illustration of how CAPE and CIN are determined from a skew-*T* diagram when the ascent is due to forced lifting (the more common of the two cases) is given by Fig. 8. The process of identifying CAPE and CIN begins by finding the parcel's LCL. After identifying the LCL, continue upward along a saturated adiabat until the tropopause, identifying the LFC (if present) and any ELs encountered in the process.

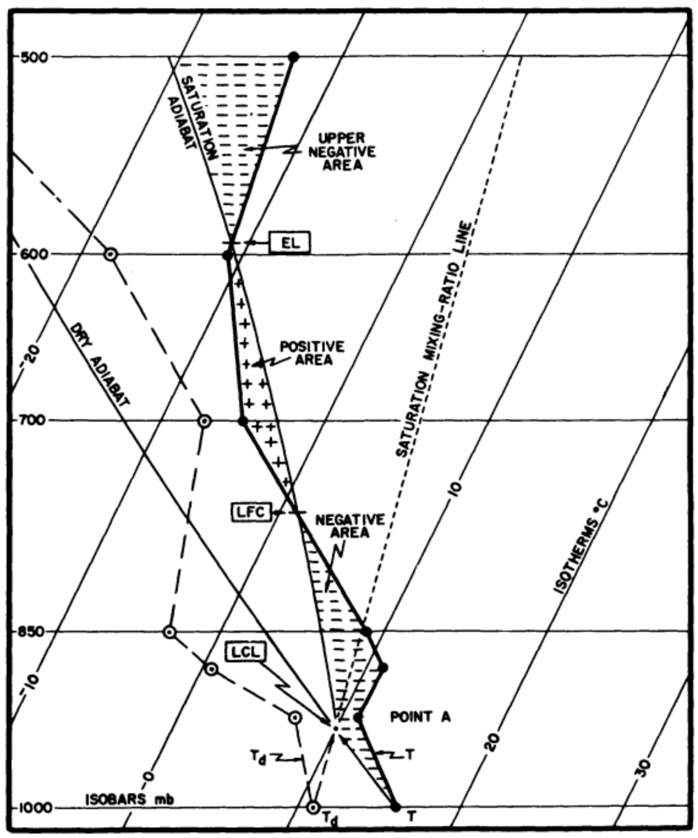

Figure 8. Illustration of how positive and negative areas are determined for an air parcel that ascends due to forced lifting. Figure reproduced from Fig. 22 of "The Use of the Skew T, Log P Diagram in Analysis and Forecasting."

The lower negative area in Fig. 8 represents the CIN that must be overcome for a parcel to reach its LFC; the upper negative area represents the CIN that must be overcome for a parcel to be able to rise beyond its EL. The positive area is the CAPE available to the ascending air parcel.

Graphical Estimation of CAPE and CIN

On the skew-*T* diagrams available through Canvas, isotherms are drawn at 1°C intervals and dry adiabats are drawn at 2°C intervals. The area of the rectangle cut out between two adjacent isotherms and two adjacent dry adiabats represents approximately 7 J kg^{-1} of energy. Thus, summing up the number of filled rectangles in positive and negative areas and multiplying by this 7 J kg^{-1} value (adding fractional amounts of this 7 J kg^{-1} value for any partially filled rectangles present) enables us to graphically estimate the CAPE and CIN present for a given scenario.

How is CIN Overcome to Unlock CAPE?

The observed sounding in Fig. 8 is identical to that in Fig. 7. This allows us to clearly illustrate that the positive and negative areas – and thus CAPE and CIN – differ depending upon how an air parcel initially ascends. In the real atmosphere, forced lifting generally dominates over lifting due to surface heating, but both are generally present (or at least possible).

Consider the case of warm season thunderstorm formation over the Great Plains. Thunderstorms typically initiate during the late afternoon hours, when a surface or near-surface air parcel is closest to its convective temperature. This minimizes the lower negative areas found in both Fig. 7, by bringing the observed temperature trace closer to the dry adiabat followed to find the convective temperature, and Fig. 8, by causing a dry adiabat closer to the observed temperature trace to be followed to find the air parcel's LCL.

Concurrently, one or more forced lifting mechanisms, such as associated with a frontal boundary, dryline, topographic feature, etc., continually lift parcels found within the planetary boundary layer (e.g., at and beneath the inversion seen in Figs. 7 and 8). Recall from before that the dry adiabatic lifting of a layer with a lapse rate that is less than the dry adiabatic lapse rate causes it to become less stable. Consequently, *persistent lifting at and beneath the inversion acts to lift and weaken the inversion, thereby reducing the amount of CIN that must be overcome for an ascending air parcel to rise freely!*