6.3 Determination of Stability and Stability Classifications

In practice, the stability of the atmosphere is determined by the application of parcel theory to atmospheric soundings on thermodynamic diagrams. Although this theory is developed on the concept of insulated parcels of minute dimensions, in current forecasting practice its application is commonly extended to large layers of the atmosphere with a fair degree of success. However, it is important to recognize the deficiencies of parcel theory as a representation of the behavior of cumulus clouds and convective systems in the real atmosphere. Some of the processes observed in real clouds and convective systems not accounted for by the parcel theory are entrainment, internal vertical mixing within the convective cell, evaporative cooling due to precipitation within the cloud, frictional drag with the environment, drag due to falling precipitation, internal viscosity, buoyancy reduction due to condensate in the cloud, large scale environmental subsidence, radiative heating and cooling, and effects of thermodynamic phase changes, especially with regard to ice formation.

Despite the limitations of parcel theory, and coupled with the fact that any thermodynamic diagram is only a graphical representation of parcel processes, we will discuss the determination of atmospheric stability on the Skew T-Log p diagram, first by comparing the slope of the temperature curve (the sounding) with the slopes of the dry and moist adiabats, and subsequently by the vertical variation of static energy.

[1]Specific humidity and mixing ratio differ by a negligible amount, and mixing ratios, which are easily measured, are commonly used instead of specific humidities.

Figure 6.3 shows three sections of schematic soundings (temperature curves) on a Skew T-Log p diagram. The dry and moist adiabats through the point of intersection (0) of the three sections are also shown. We will discuss the stability of the atmosphere in the vicinity of this point by the application of parcel concepts. By convention, the slope of a line through the point of interest is defined by the angle subtended between the upward directed line segment and the isobaric line segment in the direction of increasing temperature (i.e. in the standard trigonometric sense of angular measure). Thus in Figure 6.3, CD has the steepest slope and AB the least slope. The slope of the sounding line segment (temperature curve) in a given layer is a measure of the atmospheric lapse rate in that layer. We will see that the steeper the slope of the sounding, the more unstable the atmosphere.

If a parcel at point 0 on Figure 6.3 is lifted to 580 mb (say) its buoyancy at the end of its rise determined by parcel concepts will be as follows:

(i) For either dry or saturated ascent, the parcel will be positively buoyant ($T_p > T_e$) if the atmospheric sounding (T_e) is represented by CD, and it will continue rising. Thus CD is an absolutely unstable sounding because it is unstable to both saturated and unsaturated ascent.

(ii) The parcel is negatively buoyant ($T_p < T_e$) for dry ascent, and positively buoyant ($T_p > T_e$) for saturated ascent if the atmospheric sounding (T_e) is represented by EF. Thus EF represents a conditional state of stability in the atmosphere because the stability or instability depends on the saturation conditions of the lifted parcel.

(iii) For both dry and saturated ascent, the parcel will be negatively buoyant ($T_p < T_e$) if the atmospheric sounding (T_e) is represented by AB. Under these conditions, a lifted parcel will tend to sink back to its original level, and the atmosphere is said to be absolutely stable.

It is clear from Figure 6.3 and the above discussion that the dry and moist adiabats through a point delimit the region of the conditional state which separates the absolutely stable from the absolutely unstable soundings. Using γ to represent the atmospheric lapse rate, $-\partial T/\partial z$, i.e., the slope of the lines AB, CD, EF in Figure 6.3, and Γ_d and Γ_s to represent the dry and saturated adiabatic lapse rates respectively, we may summarize the above conclusions into the following stability criteria:

i)	$\gamma < \Gamma_s$	absolutely stable	
ii)	$\gamma = \Gamma_s$	neutral w.r. to moist, stable w.r. to dry ascent	conditional state
iii)	$\Gamma_s < \gamma < \Gamma_d$	unstable w.r. to moist, stable w.r. to dry ascent	conditional state
iv)	$\gamma = \Gamma_d$	unstable w.r. to moist, neutral w.r. to dry ascent	conditional state
v)	$\gamma > \Gamma_d$	absolutely unstable	

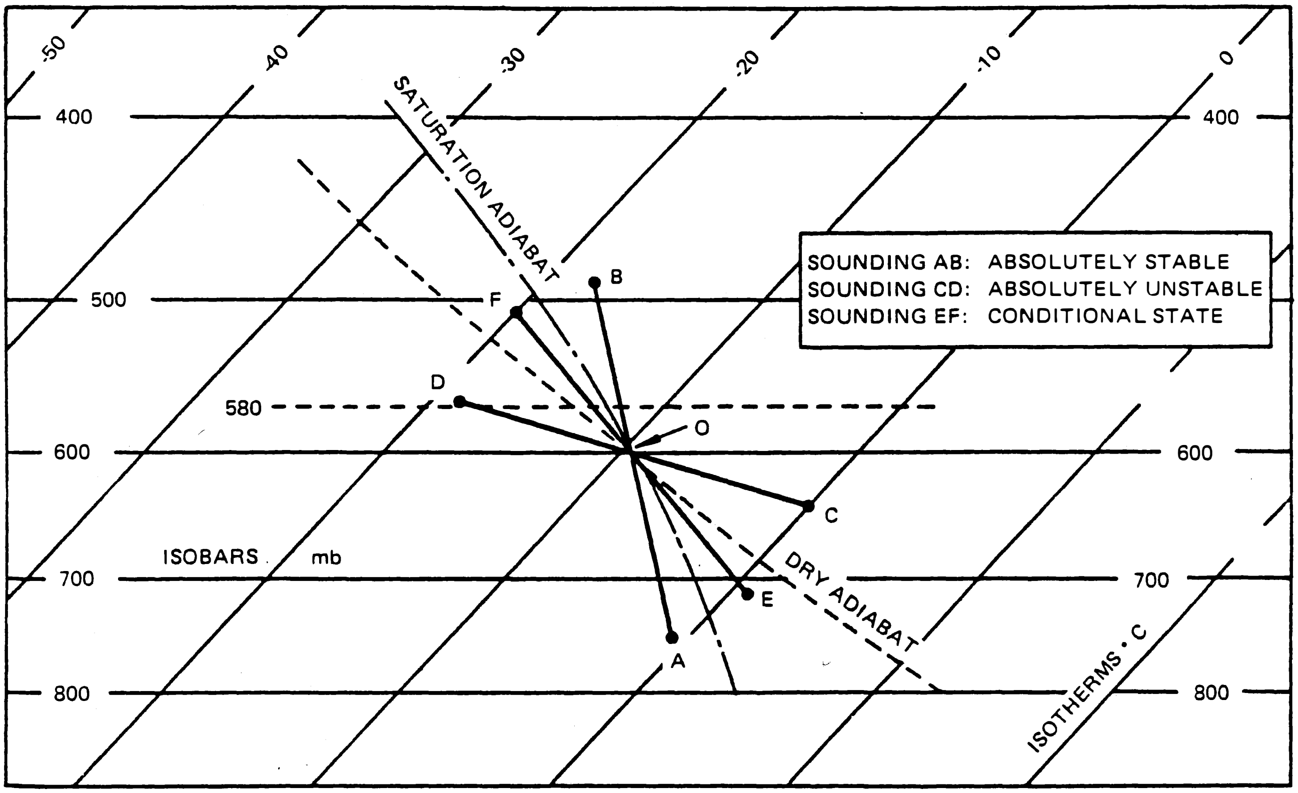

FIGURE 6.3 STABILITY CLASSIFICATIONS.

Alternative classifications of stability criteria are based on the vertical variations of dry static energy, $s(= c_pT + gz)$, and total static energy, $h(=c_pT + gz + L_eq)$ respectively. The former applies only to dry motions and the latter includes the effects of condensation when saturation occurs.

Consider the vertical derivative of dry static energy,

$$\frac{\partial s}{\partial z} = c_p \frac{\partial T}{\partial z} + g \tag{6.11}$$

Using the definition of lapse rate,

$$\gamma = -\frac{\partial T}{\partial z} = \frac{g}{c_p} - \frac{1}{c_p}\frac{\partial s}{\partial z} \tag{6.12}$$

An important relationship can be stated for the change in temperature following a parcel undergoing an adiabatic process. If $\frac{ds}{dt} = 0$, then ${}^{dT}/dt = -w\Gamma_d$ where $\Gamma_d = -\frac{dT}{dz} = \frac{g}{c_p}$, the dry adiabatic lapse rate and $w = {}^{dz}/dt$ the vertical motion of the parcel. Thus, subsiding air ($w < 0$) undergoes a warming, equivalent to the rate of dry adiabatic descent. The reverse is true for rising air.

Since $s \cong c_p\theta$, we may conclude the following:

$$\begin{array}{llll} \text{i)} & \frac{\partial\theta}{\partial z} = 0 & \frac{\partial s}{\partial z} = 0 & \gamma = \frac{g}{c_p} = \Gamma_d \\ \text{ii)} & \frac{\partial\theta}{\partial z} > 0 & \frac{\partial s}{\partial z} > 0 & \gamma < \frac{g}{c_p} \text{ or } \gamma < \Gamma_d \\ \text{iii)} & \frac{\partial\theta}{\partial z} < 0 & \frac{\partial s}{\partial z} < 0 & \gamma > \frac{g}{c_p} \text{ or } \gamma > \Gamma_d \end{array} \tag{6.13}$$

By comparison with the previous classifications, we have the following stability criteria for dry motions:

$$\begin{array}{lll} \frac{\partial\theta}{\partial z} > 0 & \text{stable} & (\gamma < \Gamma_d) \\ \frac{\partial\theta}{\partial z} = 0 & \text{neutral} & (\gamma = \Gamma_d) \\ \frac{\partial\theta}{\partial z} < 0 & \text{unstable} & (\gamma > \Gamma_d) \end{array} \tag{6.14}$$

The atmosphere cannot persist for extended periods of time, or over great depth, in a state of instability ($\frac{\partial\theta}{\partial z} < 0$) because overturning and the generation of turbulence will occur and attempt to return the unstable region to a neutrally stable condition. Such overturning and mixing occurring in the surface layer over heated ground during the day results in the growth of the well mixed boundary layer through the upward transport of heat and moisture, finally manifesting itself in the development of cumulus clouds.

A similar approach to the vertical variation of total static energy yields

$$\gamma = -\frac{\partial T}{\partial z} = \frac{g}{c_p} + \frac{L_e}{c_p}\frac{\partial q}{\partial z} - \frac{1}{c_p}\frac{\partial h}{\partial z} \tag{6.15}$$

In saturated ascent, $\frac{\partial q}{\partial z} = \frac{\partial q_s}{\partial z} < 0$, and since $h \simeq c_p\theta_e$ and θ_w is a measure of θ_e, we have

$$\text{i)} \quad \frac{\partial \theta_e}{\partial z} = 0 \qquad \frac{\partial \theta_w}{\partial z} = 0 \qquad \frac{\partial h}{\partial z} = 0 \qquad \gamma = \frac{g}{c_p} + \frac{L_e}{c_p}\frac{\partial q}{\partial z} = \Gamma_s < \Gamma_d$$

$$\text{ii)} \quad \frac{\partial \theta_e}{\partial z} > 0 \qquad \frac{\partial \theta_w}{\partial z} > 0 \qquad \frac{\partial h}{\partial z} > 0 \qquad \gamma < \Gamma_s \tag{6.16}$$

$$\text{iii)} \quad \frac{\partial \theta_e}{\partial z} < 0 \qquad \frac{\partial \theta_w}{\partial z} < 0 \qquad \frac{\partial h}{\partial z} < 0 \qquad \gamma > \Gamma_s$$

A more rigorous derivation of the saturated adiabatic lapse rate will be found in standard texts on theoretical meteorology; however, the above form illustrates the fact that the saturated adiabatic lapse rate is less than the dry because of the heating due to condensation.

The resulting stability criteria for saturated motion are

$$\frac{\partial \theta_e}{\partial z} > 0 \quad \text{stable} \quad (\gamma < \Gamma_s)$$

$$\frac{\partial \theta_e}{\partial z} = 0 \quad \text{neutral} \quad (\gamma = \Gamma_s) \tag{6.17}$$

$$\frac{\partial \theta_e}{\partial z} < 0 \quad \text{unstable} \quad (\gamma > \Gamma_s)$$

These are the criteria commonly used for determining the stability of the tropical atmosphere. Because θ_w is a direct measure of θ_e, the former may be substituted in these criteria which are referred to as <u>convective</u> stability, <u>convective</u> neutrality and <u>convective</u> instability respectively.