

Chapter 4

ATMOSPHERIC STABILITY

Wildfires are greatly affected by atmospheric motion and the properties of the atmosphere that affect its motion. Most commonly considered in evaluating fire danger are surface winds with their attendant temperatures and humidities, as experienced in everyday living. Less obvious, but equally important, are vertical motions that influence wildfire in many ways. Atmospheric stability may either encourage or suppress vertical air motion. The heat of fire itself generates vertical motion, at least near the surface, but the convective circulation thus established is affected directly by the stability of the air. In turn, the indraft into the fire at low levels is affected, and this has a marked effect on fire intensity.

Also, in many indirect ways, atmospheric stability will affect fire behavior. For example, winds tend to be turbulent and gusty when the atmosphere is unstable, and this type of airflow causes fires to behave erratically. Thunderstorms with strong updrafts and downdrafts develop when the atmosphere is unstable and contains sufficient moisture. Their lightning may set wildfires, and their distinctive winds can have adverse effects on fire behavior.

Subsidence occurs in larger scale vertical circulation as air from high-pressure areas replaces that carried aloft in adjacent low-pressure systems. This often brings very dry air from high altitudes to low levels. If this reaches the surface, going wildfires tend to burn briskly, often as briskly at night as during the day.

From these few examples, we can see that atmospheric stability is closely related to fire behavior, and that a general understanding of stability and its effects is necessary to the successful interpretation of fire-behavior phenomena.

ATMOSPHERIC STABILITY

Atmospheric stability was defined in chapter I as the resistance of the atmosphere to vertical motion. This definition and its explanation were based on the **parcel method** of analysis appropriate to a vertical temperature and moisture sounding through the troposphere.

This method employs some assumptions: (1) The sounding applies to an atmosphere at rest; (2) a small parcel of air in the sampled atmosphere, if caused to rise, does not exchange mass or heat across its boundary; and (3) rise of the parcel does not set its environment in motion. We learned that lifting under these conditions is **adiabatic lifting**.

Three characteristics of the sounding then determine the stability of the atmospheric layer

in which the parcel of air is embedded. These are: (1) The temperature lapse rate through the layer; (2) temperature of the parcel at its initial level; and (3) initial dew point of the parcel.

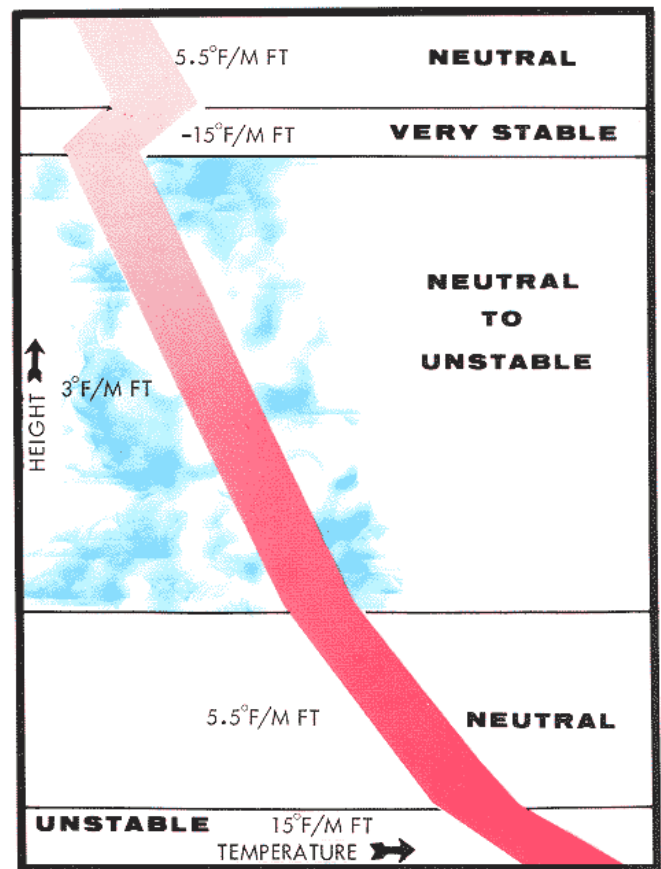
Adiabatically lifted air expands in the lower pressures encountered as it moves upward. This is a cooling process, and the rate of cooling with increase in altitude depends on whether or not the temperature reaches the dew point and consequent saturation. As long as the air remains unsaturated, it cools at the constant dry-adiabatic lapse rate of 5.5°F. per 1,000 feet of rise. Rising saturated air cools at a lesser rate, called the **moist-adiabatic** rate. This rate averages about 3°F. per 1,000 feet, but, as we will see later, it varies considerably.

STABILITY DETERMINATIONS

The degree of stability or instability of an atmospheric layer is determined by comparing its temperature lapse rate, as shown by a sounding, with the appropriate adiabatic rate. A temperature lapse rate less than the dryadiabatic rate of 5.5°F. per 1,000 feet for an unsaturated parcel is considered **stable**, because vertical motion is damped. A lapse rate greater than dry-adiabatic favors vertical motion and is **unstable**. In the absence of saturation, an atmospheric layer is **neutrally stable** if its lapse rate is the same as the dry-adiabatic rate. Under this particular condition, any existing vertical motion is neither damped nor accelerated.

In the case of a saturated parcel, the same stability terms apply. In this case, however, the comparison of atmospheric lapse rate is made with the moist-adiabatic rate appropriate to the temperature encountered.

Layers of different lapse rates of temperature may occur in a single sounding, varying from superadiabatic (unstable), usually found over heated surfaces, to dry-adiabatic (neutral), and on through inversions of temperature (very stable). In a saturated layer with considerable convective motion, the lapse rate tends to become moist-adiabatic.



Atmospheric stability of any layer is determined by the way temperature varies through the layer and whether or not air in the layer is saturated.

The adiabatic process is reversible. Just as air expands and cools when it is lifted, so is it equally compressed and warmed as it is lowered. Hence, adiabatic processes and stability determinations for either upward or downward moving air parcels make use of the appropriate dry- or moist-adiabatic lapse rates. The temperature structure of the atmosphere is always complex. As mentioned above, the moist-adiabatic lapse rate is variable-not constant as is the dry-adiabatic rate.

Adiabatic Chart

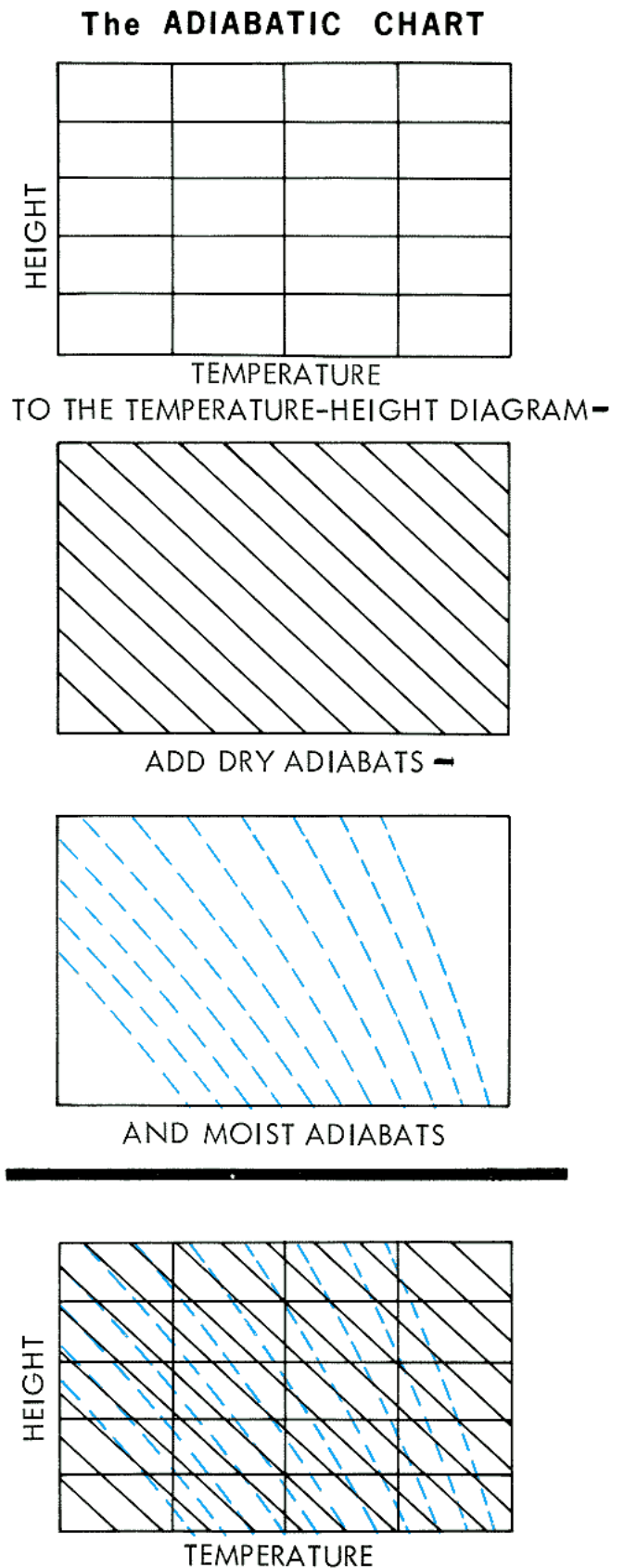
To facilitate making stability determinations, therefore, meteorologists analyzing upper-air observations use a thermodynamic diagram called an **adiabatic chart** as a convenient tool for making stability estimates. The basic portion of the chart is a set of gridlines of temperature and pressure (or height) on which the measured temperature and moisture structure of the atmosphere can be plotted. The moisture is plotted as dew-point temperature. Also printed on the chart is a set of dry-adiabatic and a set of moist-adiabatic lines. By referring to these adiabats, the lapse rates of the various layers or portions of the atmosphere can be compared to the dry-adiabatic rate and the moist-adiabatic rate. In later chapters we will consider other ways in which the adiabatic chart is used.

Stability determinations from soundings in the atmosphere are made to estimate the subsequent motion of an air parcel that has been raised or lowered by an external force. In a stable atmosphere, the parcel will return to its original position when the force is removed; in an unstable atmosphere, the parcel will accelerate in the direction of its forced motion; and in a neutrally stable atmosphere, it will remain at its new position.

Stability of Unsaturated Air

We can illustrate use of the adiabatic chart to indicate these processes by plotting four hypothetical soundings on appropriate segments of a chart. We will first consider unsaturated air to which the constant dry-adiabatic lapse rate applies.

Assume for simplicity, that each of our four soundings has a lapse rate indicated dia-



To determine stability, the meteorologist plots temperature and moisture soundings on an adiabatic chart and compares the lapse rates of various layers to the dry adiabats and moist adiabats.

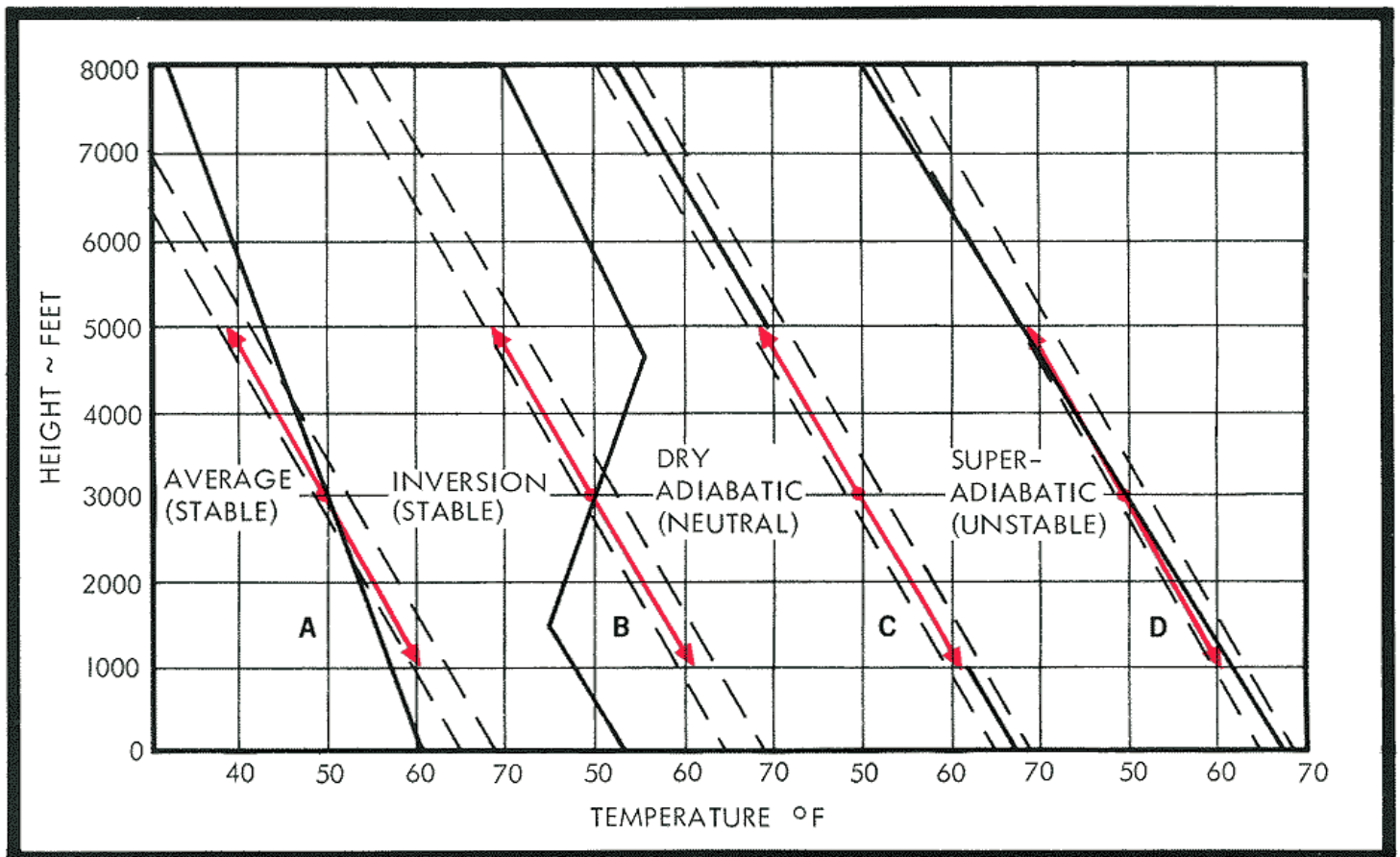
grammatically by a solid black line. Note also in the accompanying illustration that each shows the temperature at 3,000 feet to be 50°F. For our purposes, let us select a parcel of air at this point and compare its temperature with that of its environment as the parcel is raised or lowered by external forces. If it remains unsaturated, the parcel will change in temperature at the dry-adiabatic rate indicated on the chart by red arrows.

The sounding plotted in (A) has a lapse rate of 3.5°F. per 1,000 feet. As the parcel is lifted and cools at its 5.5° rate, it thus becomes progressively colder and more dense than its environment. At 5,000 feet, for example, its temperature would be 39°F., but the temperature of the surrounding air would be 43°F. Gravity thus returns the parcel to its point of origin when the external force is removed. Moved downward, the parcel warms at the dry adiabatic rate and becomes warmer than its

environment. At 1,000 feet, for example, the parcel temperature would be 61°F., but the temperature of the environment would be only 57°F. Buoyancy forces the parcel back up to its original level. The damping action in either case indicates stability.

The parcel in (B) is initially in an inversion layer where the temperature increases at the rate of 3°F. per 1,000 feet of altitude. If the parcel is lifted, say 1,000 feet, its temperature will decrease 5.5°F., while the temperature of the surrounding air will be 3°F. higher. The parcel will then be 8.5°F. colder and will return to its original level as soon as the lifting force is removed. Similarly, a lowered parcel will become warmer than the surrounding air and will also return to its original level. Thus, inversions at any altitude are very stable.

Next, let us consider (C) where the parcel is embedded in a layer that has a measured lapse rate of 5.5°F. per 1,000 feet, the same as



In unsaturated air, the stability can be determined by comparing the measured lapse rate (solid black lines) to the dry-adiabatic lapse rate (dashed black lines). The reaction of a parcel to lifting or lowering may be examined by comparing its temperature (red arrows for parcel initially at 3,000 feet and 50°F.) to the temperature of its environment.

the dry-adiabatic rate. If moved upward or downward in this layer, the parcel will change in temperature at the same rate as that of its environment and, therefore, will always be in temperature equilibrium with the surrounding air. The parcel will come to rest at its new level when external forces are removed. Technically, such a layer is neutrally stable, but we will see, after we consider an unstable case, that a neutrally stable layer is a potentially serious condition in fire weather.

In the last example (D) in unsaturated air, the plotted temperature lapse rate is 6°F. per 1,000 feet, which is greater than the dry adiabatic rate. Again, if our parcel is lifted, it will cool at the dry-adiabatic rate or 0.5° less per 1,000 feet than its surroundings. At an altitude of 5,000 feet, for example, the temperature of the parcel would be 39°F., while that of its surroundings would be 38°F. Thus, the parcel is warmer and less dense than the surrounding air, and buoyancy will cause it to accelerate upward as long as it remains warmer than the surrounding air. Moved downward, the parcel would similarly cool more rapidly than the surrounding air and accelerate downward. Hence, an atmospheric layer having a lapse rate greater than the dry-adiabatic rate is conducive to vertical motion and overturning, and represents an unstable condition.

Lapse rates greater than the dry-adiabatic rate, we learned in chapter 2, are called super-adiabatic. But since they are unstable, the air tends to adjust itself through mixing and overturning to a more stable condition. Super-adiabatic lapse rates are not ordinarily found in the atmosphere except near the surface of the earth on sunny days. When an unsaturated layer of air is mixed thoroughly, its lapse rate tends toward neutral stability.

The term "neutral" stability sounds rather passive, but we should be cautious when such a lapse rate is present. The temperature structure of the atmosphere is not static, but is continually changing. Any warming of the lower portion or cooling of the upper portion of a neutrally stable layer will cause the layer to become unstable, and it will then not only permit, but will assist, vertical motion. Such changes are easily brought about. Thus, we should consider the terms stable, neutral, and unstable in a relative, rather than an absolute,

sense. A stable lapse rate that approaches the dry-adiabatic rate should be considered relatively unstable.

Warming of the lower layers during the daytime by contact with the earth's surface or by heat from a wildfire will make a neutral lapse rate become unstable. In an atmosphere with a dry-adiabatic lapse rate, hot gases rising from a fire will encounter little resistance, will travel upward with ease, and can develop a tall convection column. A neutrally stable atmosphere can be made unstable also by advection; that is, the horizontal movement of colder air into the area aloft or warmer air into the area near the surface. Once the lapse rate becomes unstable, vertical currents are easily initiated. Advection of warm air aloft or cold air near the surface has the reverse effect of making the atmosphere more stable.

So far we have considered adiabatic cooling and warming and the degree of stability of the atmosphere only with respect to air that is not saturated. Rising air, cooling at the dry-adiabatic lapse rate, may eventually reach the dew-point temperature. Further cooling results in the condensation of water vapor into clouds, a change of state process that liberates the latent heat contained in the vapor. This heat is added to the rising air, with the result that the temperature no longer decreases at the dry-adiabatic rate, but at a lesser rate which is called the moist-adiabatic rate. On the average, as mentioned earlier, this rate is around 3°F. per 1,000 feet, but it varies slightly with pressure and considerably with temperature. The variation of the rate due to temperature may range from about 2°F. per 1,000 feet at very warm temperatures to about 5°F. per 1,000 feet at very cold temperatures. In warmer air masses, more water vapor is available for condensation and therefore more heat is released, while in colder air masses, little water vapor is available.

Stability of Saturated Air

Let us now consider a situation in which an air parcel is lifted and cooled until it reaches saturation and condensation. For this, we need to know both the initial temperature of the parcel and its dew-point temperature. This stability analysis of a sounding makes use of both

the dry-adiabatic and moist-adiabatic lines shown on the adiabatic chart. For this example, assume a sounding, plotted on the accompanying chart, showing a temperature lapse rate of 4.5°F . We will start with a parcel at sea level where the temperature is 80°F . and the dew point is 62° .

The 80°F . temperature and 62° dew point indicate that the parcel is initially unsaturated. As the parcel is lifted, it will cool at the dry-adiabatic rate until saturation occurs. The parcel dew-point temperature meanwhile decreases, as we learned in chapter 3, at the rate of 1°F . per 1,000 feet. If we draw a line on the adiabatic chart with a slope of -1°F . starting at the surface 62° dew point, we find that this line intersects the dry-adiabatic path of the parcel. The parcel temperature at this point is therefore at the dew point. The altitude of the point is thus at the **condensation level**.

In our example, condensation occurs at 4,000 feet above sea level at a temperature of 58° . The atmosphere is stable at this point because the parcel temperature is lower than that

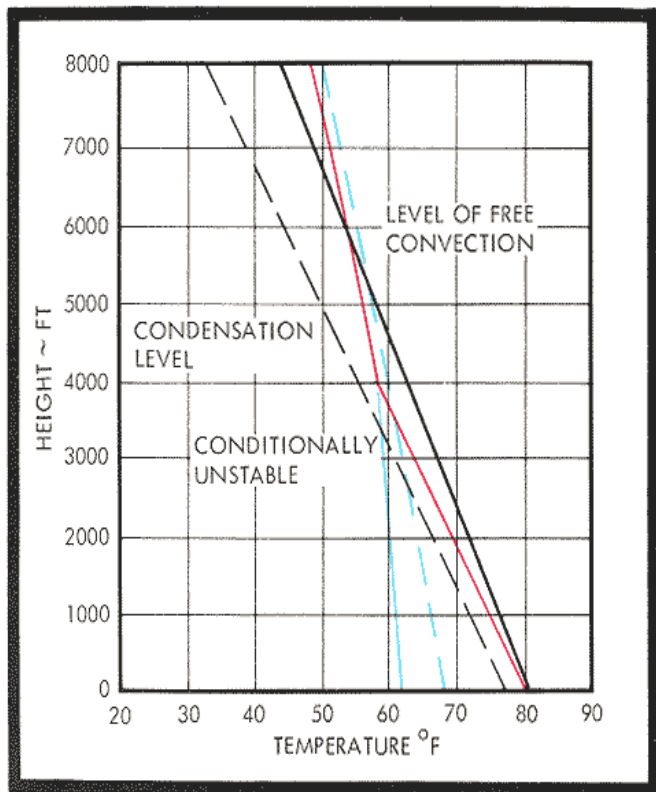
shown by the sounding for the surrounding air. If the parcel is forced to rise above the condensation level, however, it then cools at the moist-adiabatic rate, in this case about 2.5°F . per 1,000 feet. At this rate of change, the parcel temperature will reach the temperature of the surrounding air at 6,000 feet. The level at which the parcel becomes warmer than the surrounding air is called the **level of free convection**. Above this level, the parcel will become buoyant and accelerate upward, continuing to cool at the moist-adiabatic rate, and no longer requiring an external lifting force.

Conditional Instability

The atmosphere illustrated by the above example, which has a lapse rate lying between the dry and moist adiabats, is said to be **conditionally unstable**. It is stable with respect to a lifted air parcel as long as the parcel remains unsaturated, but it is unstable with respect to a lifted parcel that has become saturated. In our example, the measured lapse rate of the layer is 4.5°F . This layer is, therefore, stable with respect to a lifted parcel as long as the parcel temperature follows the dry-adiabatic rate. It is unstable with respect to a lifted saturated parcel, because the temperature of the saturated parcel would follow the lesser moist-adiabatic rate, in this case about 2.5°F . per 1,000 feet.

A saturated parcel in free convection loses additional moisture by condensation as it rises. This, plus the colder temperature aloft, causes the moist-adiabatic lapse rate to increase toward the dry-adiabatic rate. The rising parcel will thus eventually cool to the temperature of the surrounding air where the free convection will cease. This may be in the vicinity of the tropopause or at some lower level, depending on the temperature structure of the air aloft.

Reliance on the parcel method of analyzing atmospheric stability must be tempered with considerable judgment. It is true that from the plotted temperature lapse rates on the adiabatic chart one can read differences between temperatures of parcels and the surrounding air. These are based, however, on the initial assumptions upon which the method is founded. One of these, for example, is that there is no



A lapse rate between the dry- and moist-adiabatic rates is conditionally unstable, because it would be unstable under saturated conditions but stable under unsaturated conditions. The temperature of a parcel raised from near the surface will follow the dry-adiabatic rate until saturation, then follow the moist-adiabatic rate. At the level where the parcel temperature exceeds the environment temperature, the parcel will begin free ascent.

energy exchange between the parcel and the surrounding air. Vertical motion is, however, often accompanied by various degrees of mixing and attendant energy exchange, which makes this assumption only an approximation. The usual practice of plotting the significant turning points from sounding data and connecting them with straight lines

also detracts from precision. These are additional reasons for considering stability in a relative sense rather than in absolute terms.

The temperature of the parcel and the environment, and the dew-point temperature of the parcel used in this example, are summarized below;

Altitude	Environment temperature	Parcel temperature	Dew-point temperature
Sea level	80	80 *	62
2000'	71	69 * Dry-adiabatic lapse rate	60
4000' Condensation level	62	58 *	58
6000' Level of free convection	53	53 ~ Moist-adiabatic laps	53
8000'	44	48 ~	48

LAYER STABILITY

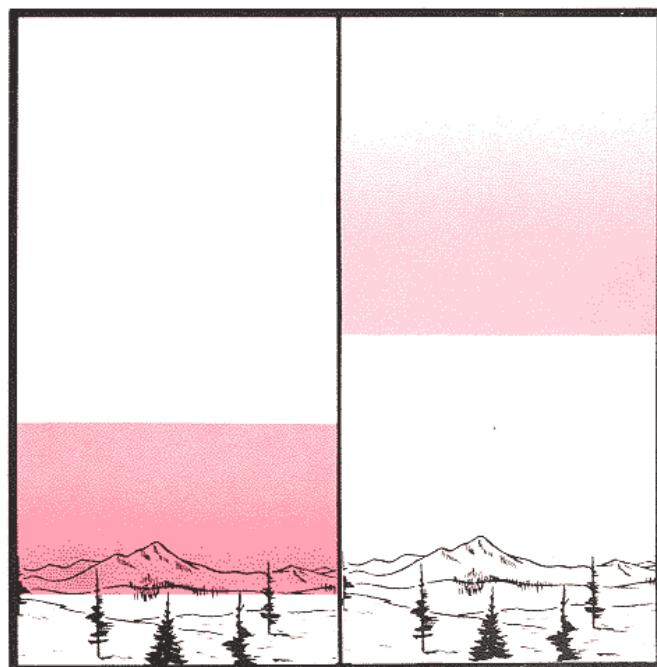
Many local fire-weather phenomena can be related to atmospheric stability judged by the parcel method. Equally important, however, are weather changes that occur when whole layers of the atmosphere of some measurable depth and of considerable horizontal extent are raised or lowered. Here again, it is necessary to employ some assumptions with respect to conservation of mass and energy, and the assumption that the adiabatic processes still apply. However, it is often possible to employ these concepts with somewhat greater confidence here than in the case of parcel-stability analyses. Let us first examine how the stability of an air layer changes internally as the layer is lifted or lowered.

When an entire layer of stable air is **lifted** it becomes increasingly **less stable**. The layer stretches vertically as it is lifted, with the top rising farther and cooling more than the bottom. If no part of the layer reaches condensation, the stable layer will eventually become dry-adiabatic. Let us consider an example:

We will begin with a layer extending from 6,000 to 8,000 feet with a lapse rate of 3.5°F. per 1,000 feet, and raise it until its base is at 17,000 feet. Because of the vertical stretching upon reaching lower pressures, the layer would be about 3,000 feet deep at its new altitude and the top would be at 20,000 feet. If the air in the layer remained unsaturated, its temperature would have decreased at the dry-adiabatic rate. The temperature of the top of the layer would have decreased 5.5×12 , or 66°F. The temperature of the bottom of the layer would have decreased 5.5×11 , or 60.5°F.

Originally, the difference between the bottom and top was 7°F., but after lifting it would be $66 - 60.5 = 5.5^\circ\text{F.}$ greater, or 12.5°F. Whereas the original lapse rate was 3.5°F. per 1,000 feet, it is $12.5 / 3$, or 4.2°F. per 1,000 feet after lifting. The layer has become less stable.

Occasionally, the bottom of a layer of air being lifted is more moist than the top and reaches its condensation level early in the lift-



A lifted layer of air stretches vertically, with the top rising farther and cooling more than the bottom. If the layer is initially stable, it becomes increasingly less stable as it is lifted. Similarly, a subsidizing layer becomes more stable.

ing. Cooling of the bottom takes place at the slower moist-adiabatic rate, while the top continues to cool at the dry-adiabatic rate. The layer then becomes increasingly less stable at a rate faster than if condensation had not taken place.

A **descending** (subsiding) layer of stable air becomes **more stable** as it lowers. The layer compresses, with the top sinking more and

warming more than the bottom. The adiabatic processes involved are just the opposite of those that apply to rising air.

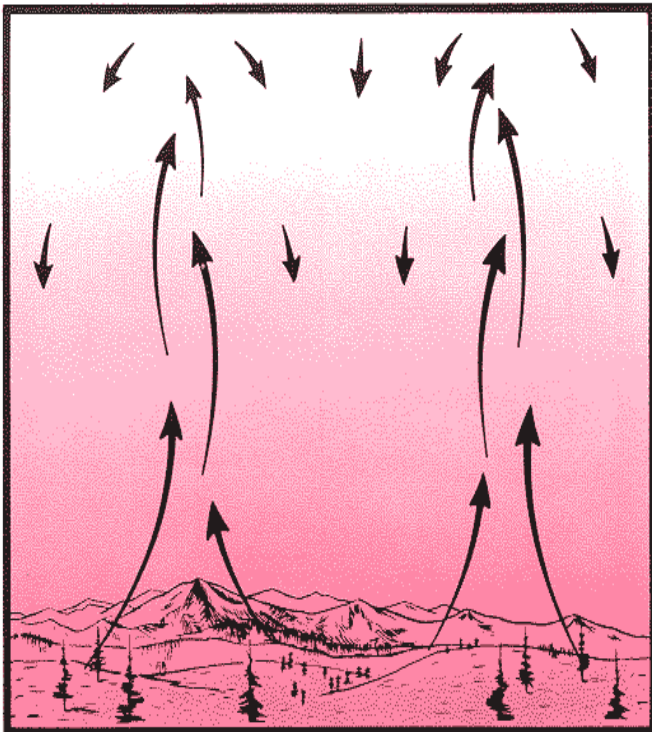
Since the lapse rate of the atmosphere is normally stable, there must be some processes by which air parcels or layers are lifted in spite of the resistance to lifting provided by the atmosphere. We will consider several such processes.

LIFTING PROCESSES

A common process by which air is lifted in the atmosphere, as is explained in detail in the next chapter, is **convection**. If the atmosphere remains stable, convection will be suppressed. But we have seen that surface heating makes the lower layers of the atmosphere unstable during the daytime. Triggering mechanisms are required to begin convective action, and they usually are present. If the unstable layer is deep enough, so that the rising parcels reach their condensation level, cumulus-type clouds will form and may produce showers or thunderstorms if the atmosphere layer above the con-

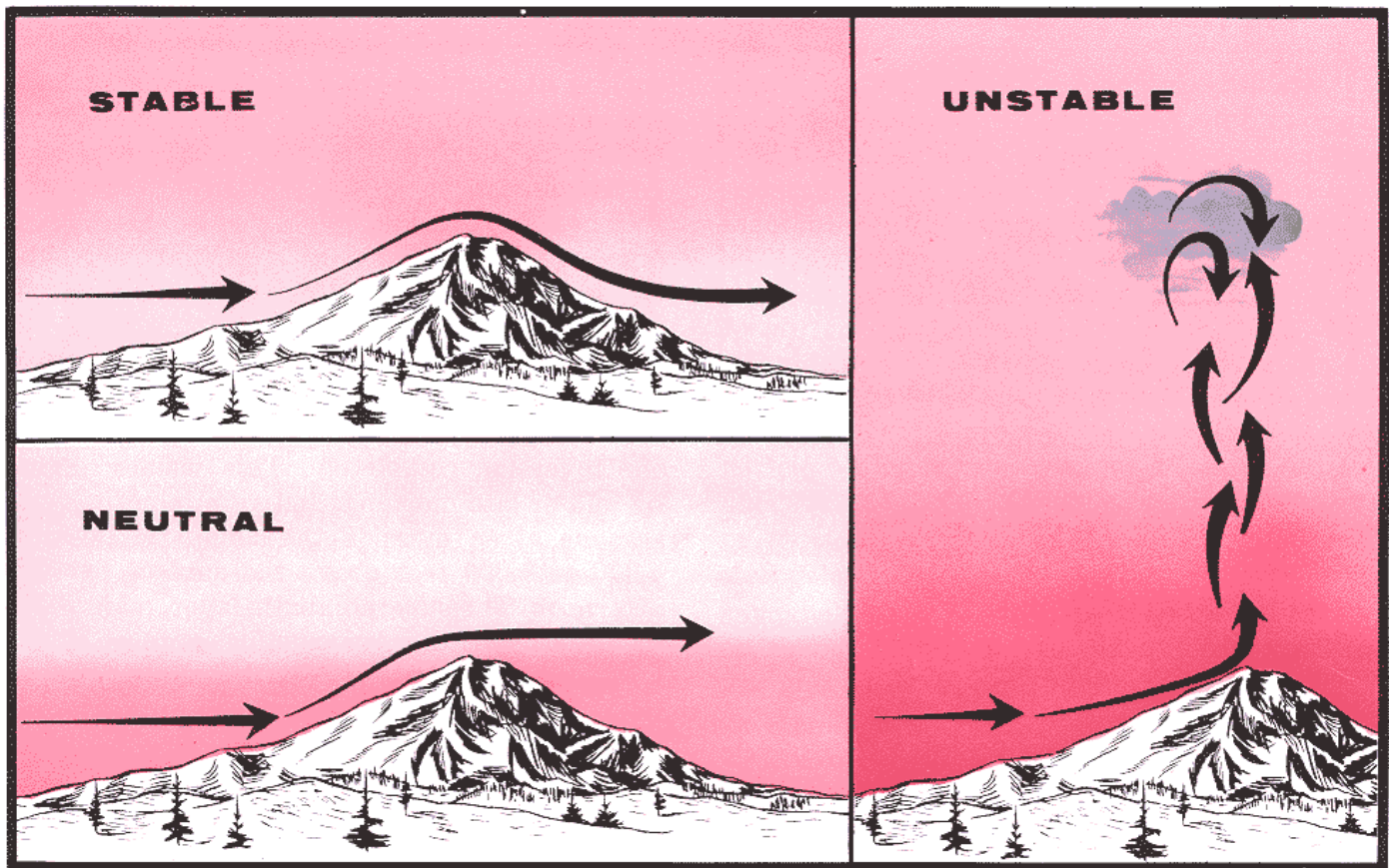
densation level is conditionally unstable. Wildfire also may be a source of heat which will initiate convection. At times, the fire convection column will reach the condensation level and produce clouds. Showers, though rare, have been known to occur.

Layers of air commonly flow in response to pressure gradients. In doing so, if they are lifted up and over mountains, they are subjected to what is called **orographic lifting**. This is a very important process along our north-south mountain ranges in the western regions and the Appalachians in the East, because the general airflow is normally from a westerly direction. If the air is initially stable, and if no condensation takes place, it sinks back to its original level after passing over a ridge. If it is neutrally stable, the air will remain at its new level after crossing the ridge. In an unstable atmosphere, air given an initial uplift in this way keeps on rising, seeking a like temperature level, and is replaced by sinking colder air from above. If the condensation level is reached in the lifting process, and clouds form, initially stable air can become unstable. In each case, the internal depth and lapse rate of the layer will respond as indicated above.



Convection is a process by which air is lifted in the atmosphere. Surface heating during the daytime makes the surface layer of air unstable. After its initial inertia is overcome, the air is forced upward by the more dense surrounding air.

As we will see in the chapter on air masses and fronts, warmer, lighter air layers frequently flow up and over colder, heavier air masses. This is referred to as **frontal lifting** and is similar in effect to orographic lifting. Stable and unstable air masses react the same way regardless of whether they are lifted by the slope of topography or by the slope of a heavier air mass.



As air is lifted over mountain, the resulting airflow depends to some extent upon the stability of the air. These simple airflows may be complicated considerably by daytime heating and, in some cases, by wave motion.

Turbulence associated with strong winds results in mixing of the air through the turbulent layer. In this process, some of the air near the top of the layer is mixed downward, and that near the bottom is mixed upward, resulting in an adiabatic layer topped by an inversion. At times, the resultant cooling near the top of the layer is sufficient to produce condensation and the formation of stratus, or layerlike, clouds.

The airflow around surface **low-pressure** areas in the Northern Hemisphere is counter-clockwise and spirals inward. In the next chapter we will see why this is so, but here we will need to consider the inflow only because it produces upward motion in low-pressure areas. Airflow into a Low from all sides is called convergence. Now, the air must move. It is prevented from going downward by the earth's surface, so it can only go upward. Thus, low-pressure areas on a surface weather map are regions of upward motion in the lower atmosphere.

In surface **high-pressure** areas, the airflow is clockwise and spirals outward. This airflow away from a High is called **divergence**. The air must be replaced, and the only source is from aloft. Thus, surface high-pressure areas are regions of sinking air motion from aloft, or **subsidence**. We will consider subsidence in more detail later in this chapter.

Frequently, two or more of the above processes will act together. For example, the stronger heating of air over ridges during the daytime, compared to the warming of air at the same altitude away from the ridges, can aid orographic lifting in the development of deep convective currents, and frequently cumulus clouds, over ridges and mountain peaks. Similarly, orographic and frontal lifting may act together, and frontal lifting may combine with convergence around a Low to produce more effective upward motion.

DIURNAL AND SEASONAL VARIATIONS IN STABILITY

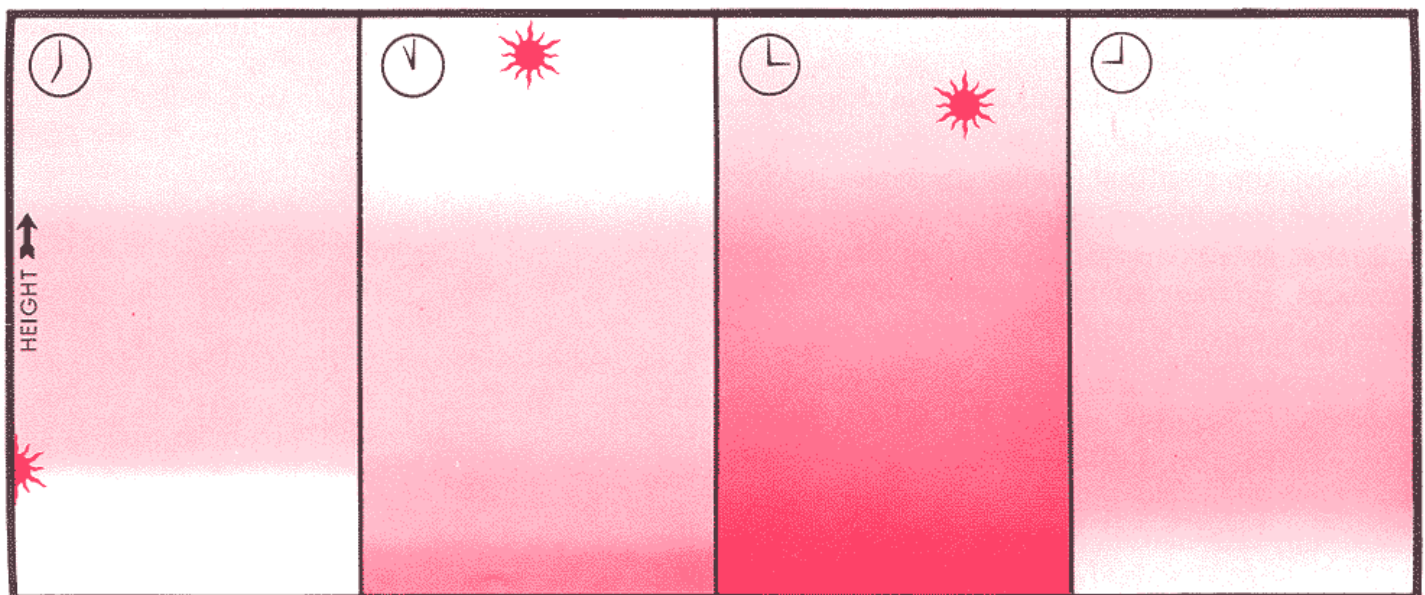
Stability frequently varies through a wide range in different layers of the atmosphere for various reasons. Layering aloft may be due to an air mass of certain source-region characteristics moving above or below another air mass with a different temperature structure. The inflow of warmer (less dense) air at the bottom, or colder (more dense) air at the top of an air mass promotes instability, while the inflow of warmer air at the top or colder air at the surface has a stabilizing effect. The changes in lapse rate of a temperature sounding plotted on an adiabatic chart frequently correspond closely to the layering shown in upper-wind measurements.

At lower levels, stability of the air changes with surface heating and cooling, amount of cloud cover, and surface wind all acting together. We will consider first the changes in stability that take place during a daily cycle and the effects of various factors; then we will consider seasonal variations.

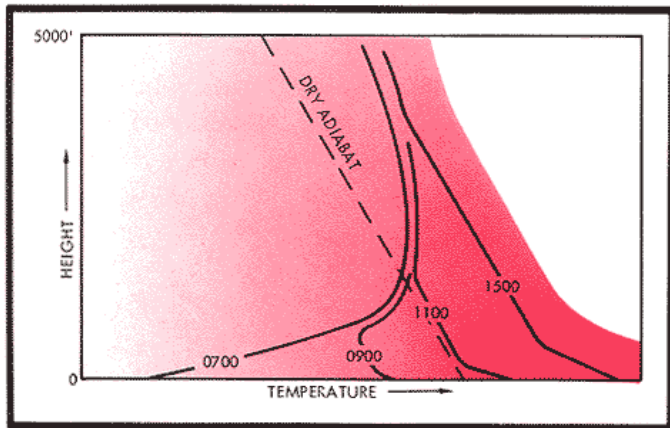
Diurnal changes in surface heating and cooling, discussed in chapter 2, and illustrated in particular on pages 27, 28, produce daily changes in stability, from night inversions to daytime superadiabatic lapse rates, that are common over local land surfaces. During a typical light-wind, fair-weather period, radiation cooling at night forms

a stable inversion near the surface, which deepens until it reaches its maximum development at about daybreak. After sunrise, the earth and air near the surface begin to heat, and a shallow superadiabatic layer is formed. Convective currents and mixing generated in this layer extend up to the barrier created by the inversion. As the day progresses, the unstable superadiabatic layer deepens, and heated air mixing upward creates an adiabatic layer, which eventually eliminates the inversion completely. This usually occurs by mid or late morning. Active mixing in warm seasons often extends the adiabatic layer to 4,000 or 5,000 feet above the surface by midafternoon. The superadiabatic layer, maintained by intense heating, is usually confined to the lowest few hundreds of feet, occasionally reaching 1,000 to 2,000 feet over bare ground in midsummer.

As the sun sets, the ground cools rapidly under clear skies and soon a shallow inversion is formed. The inversion continues to grow from the surface upward throughout the night as surface temperatures fall. The air within the inversion becomes increasingly stable. Vertical motion in the inversion layer is suppressed, though mixing may well continue in the air above the inversion. This mixing allows radia-



On a typical fair-weather summer day, stability in the lower atmosphere goes through a regular cycle. Cooling at night near the surface stabilizes the layer of air next to the ground. Warming during the daytime makes it unstable.



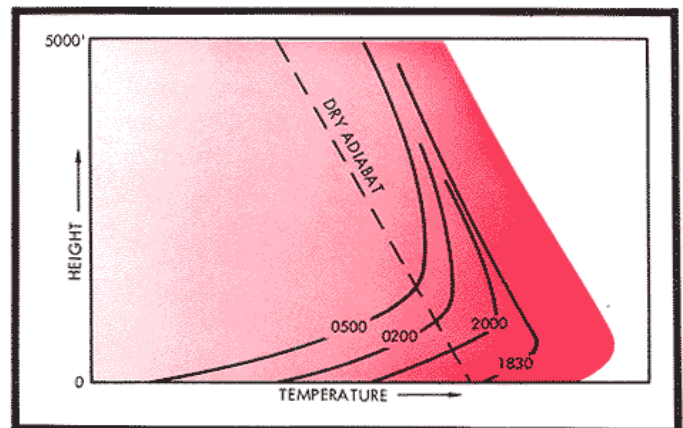
A night surface inversion (0700) is gradually eliminated by surface heating during the forenoon of a typical clear summer day. A surface superadiabatic layer and a dry-adiabatic layer above deepen until they reach their maximum depth about mid afternoon.

tional cooling above the inversion to lower temperatures in that layer only slightly during the night.

This diurnal pattern of nighttime inversions and daytime superadiabatic layers near the surface can be expected to vary considerably. Clear skies and low air moisture permit more intense heating at the surface by day and more intense cooling by radiation at night than do cloudy skies. **The lower atmosphere tends to be more unstable on clear days and more stable on clear nights.**

Strong winds diminish or eliminate diurnal variations in stability near the surface. Turbulence associated with strong wind results in mixing, which tends to produce a dry-adiabatic lapse rate. Mechanical turbulence at night prevents the formation of surface inversions, but it may produce an inversion at the top of the mixed layer. During the day, thermal turbulence adds to the mechanical turbulence to produce effective mixing through a relatively deep layer. Consequently, great instability during the day, and stability at night occur when surface winds are light or absent.

Stability in the lower atmosphere varies locally between surfaces that heat and cool at different rates. Thus, dark-colored, barren, and rocky soils that reach high daytime temperatures contribute to strong daytime instability and, conversely, to strong stability at night. Areas recently blackened by fire are subject to about the maximum diurnal variation in surface temperature



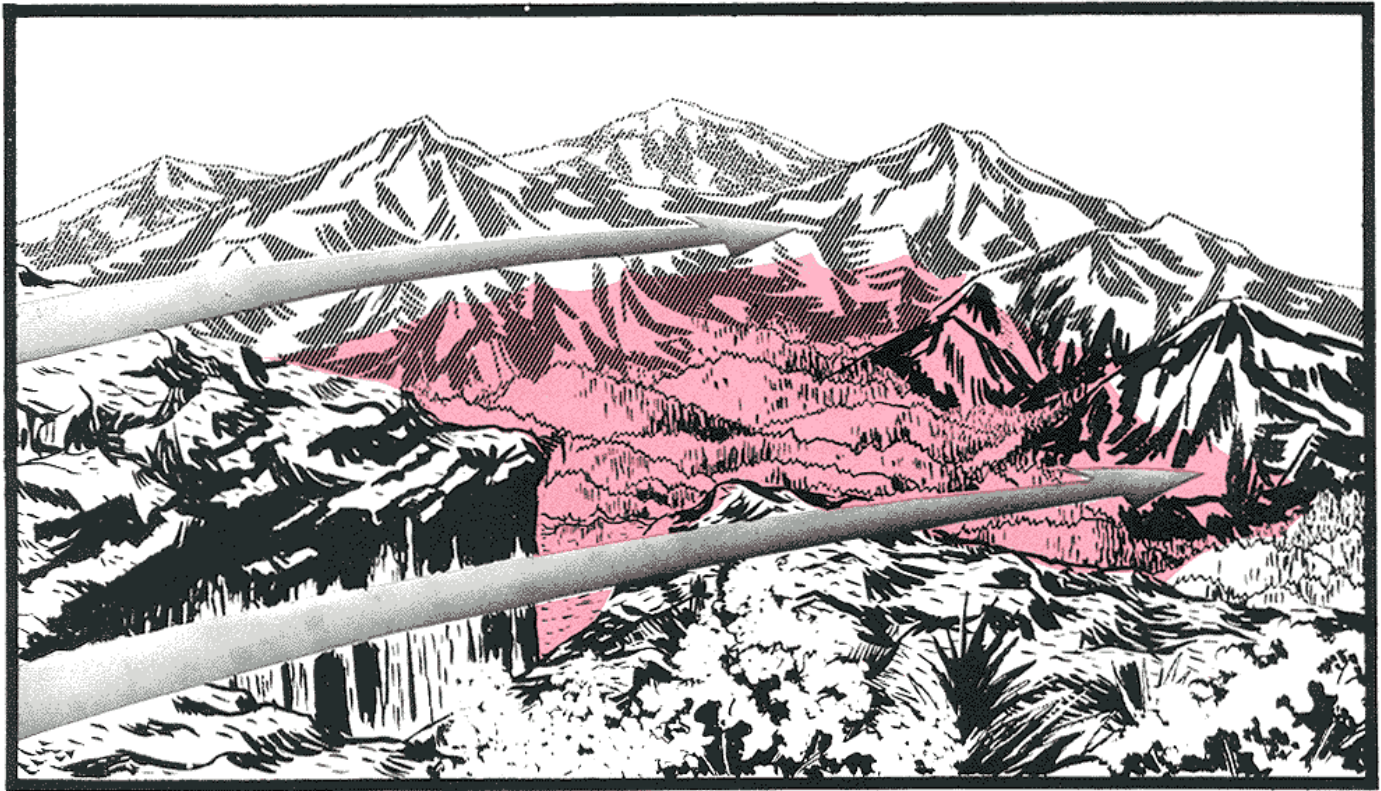
The ground cools rapidly after sundown and a shallow surface inversion is formed (1830). This inversion deepens from the surface upward during the night, reaching its maximum depth just before sunrise (0500).

and the resulting changes in air stability. Vegetated areas that are interspersed with openings, outcrops, or other good absorbers and radiators have very spotty daytime stability conditions above them.

Topography also affects diurnal changes in the stability of the lower atmosphere. Air in mountain valleys and basins heats up faster during the daytime and cools more rapidly at night than the air over adjacent plains. This is due in part to the larger area of surface contact, and in part to differences in circulation systems in flat and mountainous topography. The amount of air heating depends on orientation, inclination, and shape of topography, and on the type and distribution of ground cover. South-facing slopes reach higher temperatures and have greater instability above them during the day than do corresponding north slopes. Both cool about the same at night.

Instability resulting from superheating near the surface is the origin of many of the important convective winds which we will discuss in detail in chapter 7. On mountain slopes, the onset of daytime heating initiates upslope wind systems. The rising heated air flows up the slopes and is swept aloft above the ridge tops in a more-or-less steady stream.

Over level ground, heated surface air, in the absence of strong winds to disperse it, can remain in a layer next to the ground until it is disturbed. The rising air frequently spirals upward in the form of a whirlwind or dust



Strong heating may produce a pool of superheated air in poorly ventilated basins. If upper winds are unable to provide the triggering mechanism needed to overcome inertia and release the instability in this superadiabatic layer, a potentially explosive fire weather situation develops.

devil. In other cases, it moves upward as intermittent bubbles or in more-or-less continuous columns. Pools of superheated air may also build up and intensify in poorly ventilated valleys to produce a highly unstable situation. They persist until released by some triggering mechanism which overcomes inertia, and they may move out violently.

The amount of solar radiation received at the surface during the summer is considerably greater than in the winter. As explained in chapter 1, this is due to the difference in solar angle and the duration of sunshine. Temperature profiles and stability reflect **seasonal variation** accordingly. In the colder months, inversions become more pronounced and more persistent, and superadiabatic lapse rates occur only occasionally.

In the summer months, superadiabatic conditions are the rule on sunny days. Greater variation in stability from day to day may be expected in the colder months because of the greater variety of air masses and weather situations that occur during this stormy season.

In addition to the seasonal effects directly caused by changes in solar radiation, there is also an important effect that is caused by the lag in heating and cooling of the atmosphere as a whole. The result is a predominance of cool air over warming land in the spring, and warm air over cooling surfaces in the fall. Thus, the steepest lapse rates frequently occur during the spring, whereas the strongest inversions occur during fall and early winter.

SUBSIDENCE

Air that rises in the troposphere must be replaced by air that sinks and flows in beneath that which rises. Local heating often results in small-scale updrafts and downdrafts in the same vicinity.

On a larger scale, such as the up-flow in low-pressure systems, adjacent surface high-pressure systems with their divergent flow normally supply the replacement air. The

outflow at the surface from these high-pressure areas results in sinking of the atmosphere above them. **This sinking from aloft is the common form of subsidence.**

The sinking motion originates high in the troposphere when the high-pressure systems are deep. Sometimes these systems extend all the way from the surface up to the tropopause. Deep high-pressure systems are referred to as **warm Highs**, and subsidence through a deep layer is characteristic of warm Highs.

Subsidence occurs in these warm high-pressure systems as part of the return circulation compensating for the large upward transport of air in adjacent low-pressure areas. If the subsidence takes place without much horizontal mixing, **air from the upper troposphere may reach the surface quite warm and extremely dry.**

For example, the saturation absolute humidity of air in the upper troposphere with a temperature of -50° to -60°F. is less than 0.02 pounds per 1,000 cubic feet. In lowering to the surface, this air may reach a temperature of 70°F. or higher, where saturation would represent 1.15 pounds or more of water per 1,000 cubic feet. If no moisture were added to the air in its descent, the relative humidity would then be less than 2 percent.

Subsiding air may reach the surface at times with only very little external modification or addition of moisture. Even with considerable gain in moisture, the final relative humidity can be quite low. The warming and drying of air sinking adiabatically is so pronounced that saturated air, sinking from even the middle troposphere to near sea level, will produce relative humidities of less than 5 percent. Because of the warming and drying, subsiding air is characteristically very clear and cloudless.

Subsidence in a warm high-pressure system progresses downward from its origin in the upper troposphere. In order for the sinking motion to take place, the air beneath must flow outward, or diverge. Thus, **horizontal divergence** is an integral part of subsidence in the troposphere. The descent rate is observed by following the progress of the subsidence inversion on successive upper-air soundings.

The accompanying chart shows a simplified illustration of the subsidence inversion on 3

successive days. The temperature lapse rate in the descending layer is nearly dry-adiabatic, and its bottom surface is marked by a temperature inversion. Two features, a temperature inversion and a marked decrease in moisture, identify the base of a subsiding layer. Below the inversion, there is an abrupt rise in the moisture content of the air.

The rate of descent of subsiding air varies widely. It is typically fastest at higher levels and becomes progressively slower near the surface. It is commonly about 5,000 feet in 6 hours around the 30,000-foot level, and about 500 feet in 6 hours at the 6,000-foot level.

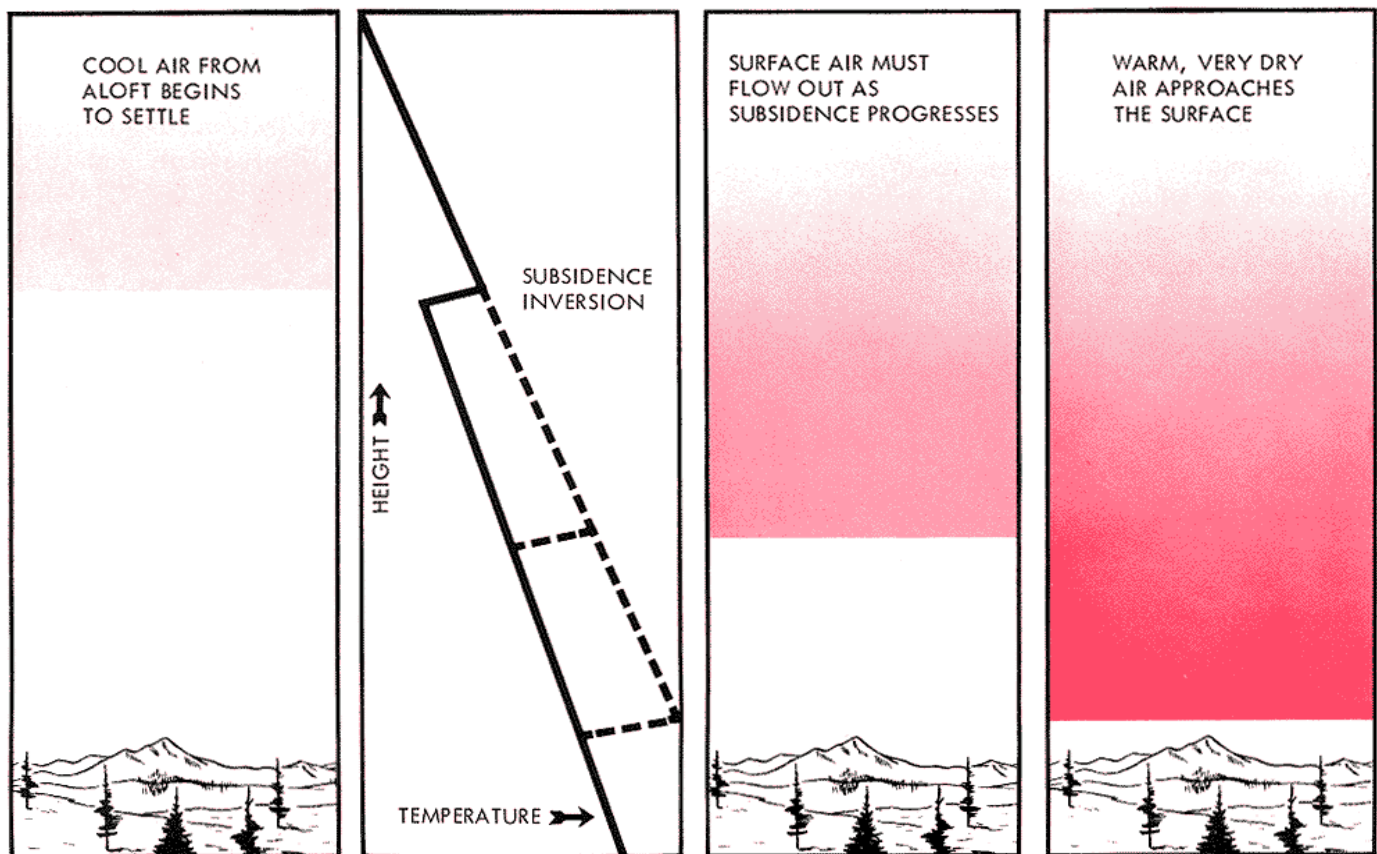
Frequently, the subsiding air seems to lower in successive stages. When this happens, a sounding will show two or more inversions with very dry air from the top down to the lowest inversion. This air may be drier than can be measured with standard sounding equipment.

Subsiding air seldom reaches the surface as a broad layer. Often, it sinks to the lower troposphere and then stops. We need, therefore, to consider ways in which the dry air no longer lowering steadily over a broad area can affect the surface.

Along the west coast in summer we generally find a cool, humid advected marine layer 1,000-2,000 feet thick with a warm, dry subsiding layer of air above it. This subsidence inversion is usually low enough so that coastal mountains extend up into the dry air. The higher topographic elevations will experience warm temperatures and very low humidities both day and night. Some mixing of moisture upward along the slopes usually occurs during the daytime with upslope winds.

As the marine layer moves inland from the coast during clear summer days, it is subjected to intensive heating and becomes warmer and warmer until finally the subsidence inversion is wiped out. The temperature lapse rate from the surface to the base of the dry air, or even higher, becomes dry-adiabatic. Then, convective currents can be effective in bringing dry air from aloft down to the surface and mixing the more moist air from near the surface to higher levels.

This process can well take place in other regions when the subsidence inversion reaches low-enough levels so it can be eliminated by



The descent of a subsidence inversion may be followed on successive soundings, as shown by dashed lines. As the more humid surface air flows outward, the drier air aloft is allowed to sink and warm adiabatically.

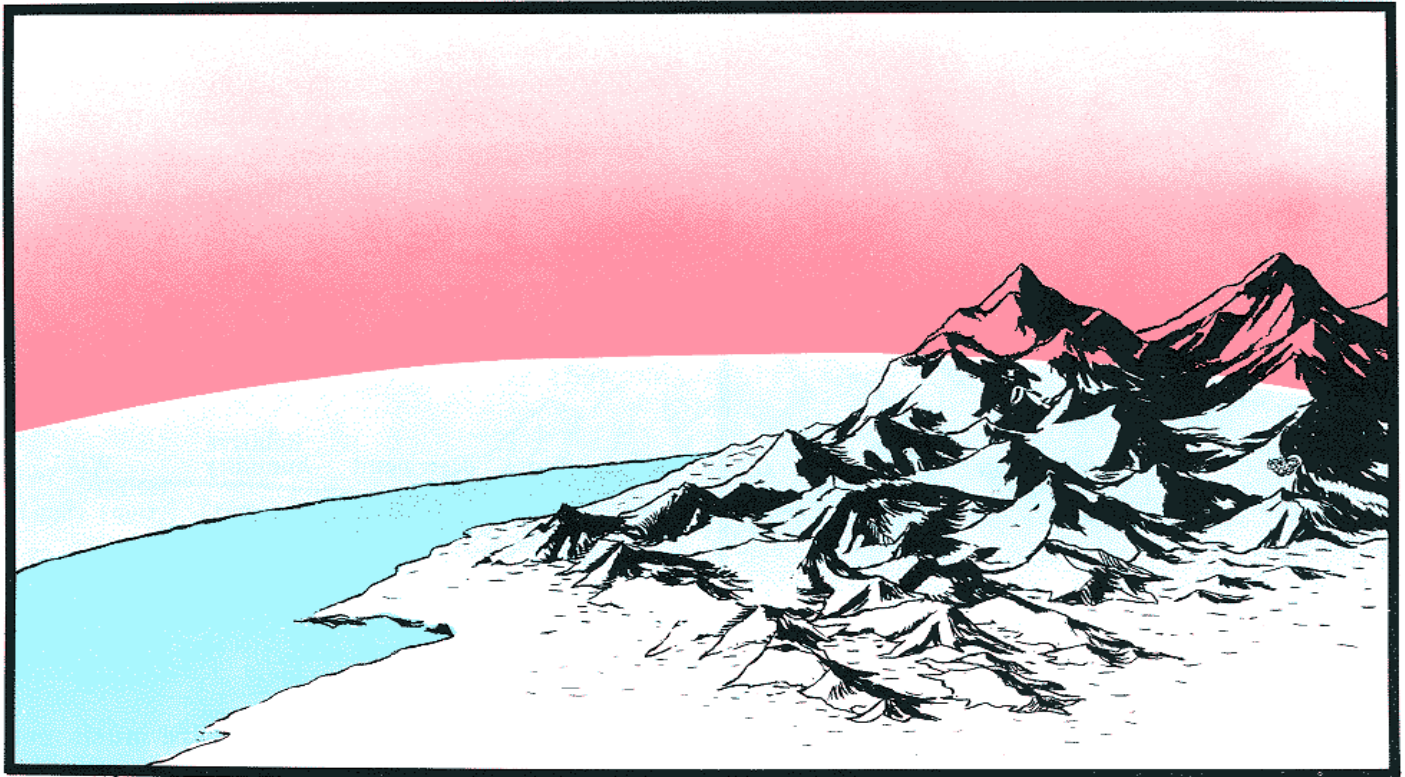
surface daytime heating, The inversion will be wiped out only in local areas where surface heating is intense enough to do the job. If the heating is not sufficient to eliminate the inversion, the warm, dry air cannot reach the surface by convection. Convective currents in the layer beneath the inversion may be effective in eating away the base of the inversion and mixing some of the dry air above with the more humid air below. This process will warm and dry the surface layer somewhat, but humidities cannot reach the extremely low values characteristic of a true subsidence situation.

Another method by which dry, subsiding air may reach the surface is by following a sloping downward path rather than a strictly vertical path. A vertical sounding may show that the subsiding air is much too warm to reach the surface by sinking vertically, because the layer beneath it is cooler and denser. However, if surface air temperatures are warmer downstream, the subsiding air can sink

dry-adiabatically to lower levels as it moves downstream and may eventually reach the surface. This process is most likely to occur around the eastern and southern sides of a high-pressure area where temperatures increase along the air trajectory. By the time the sinking air reaches the surface, it is likely to be on the south, southwest, or even west side of the High.

Subsiding air may reach the surface in a dynamic process through the formation of mountain waves when strong winds blow at right angles to mountain ranges. Waves of quite large amplitude can be established over and on the leeward side of ranges. Mountain waves can bring air from great heights down to the surface on the lee side with very little external modification. These waves may also be a part of the foehn-wind patterns, which we will touch off only briefly here since they will be treated in depth in chapter 6.

In the mountain areas of the West, **foehn winds**, whether they are the **chinook** of the

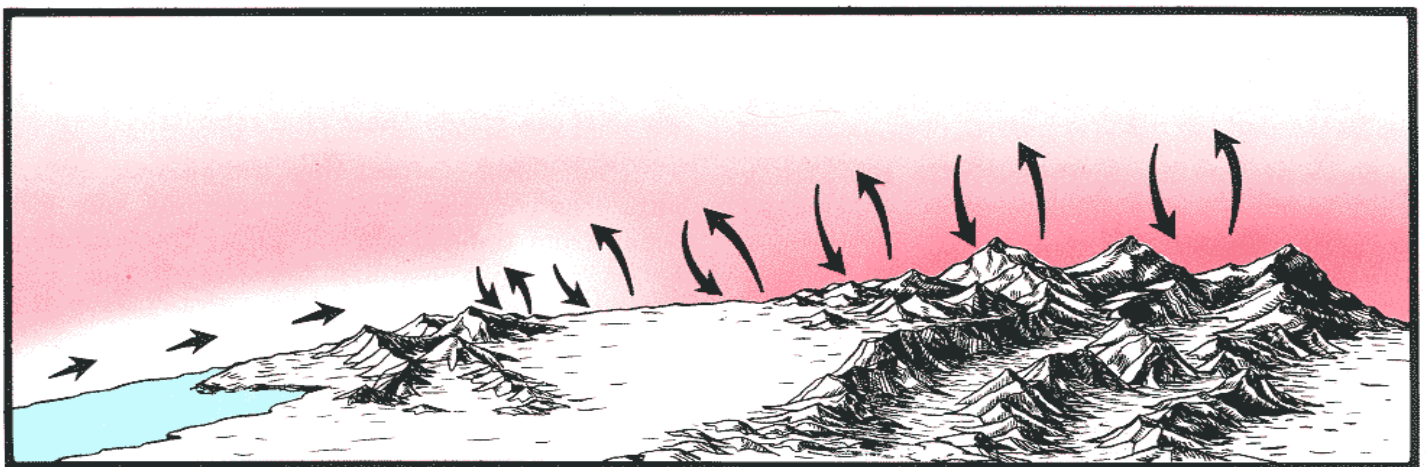


Along the west coast in summer, high elevations in the coastal mountains, extending into the dry, subsiding air have warm temperatures and very low humidities both day and night, while lower coastal slopes are influenced by the cool, humid marine layer.

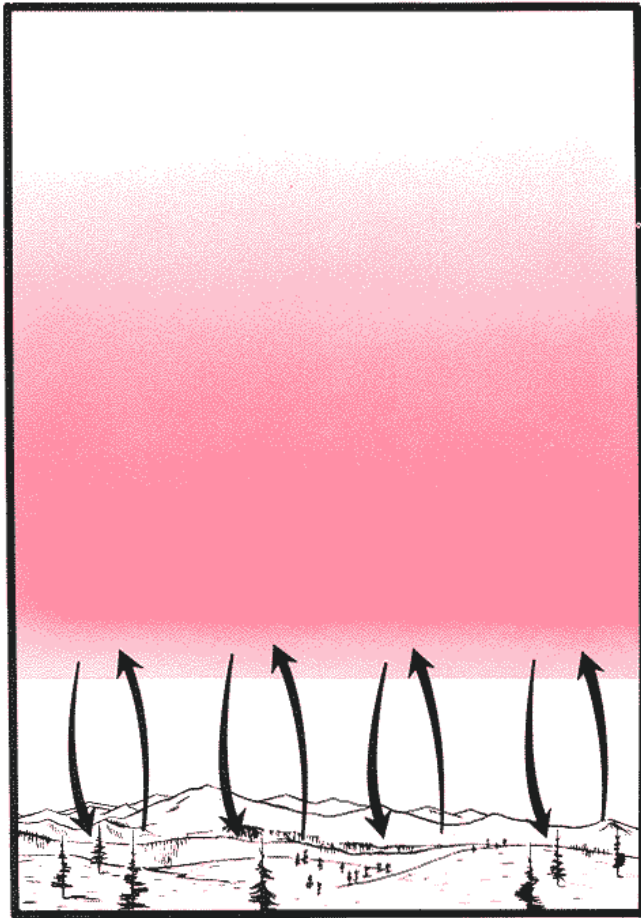
eastern slopes of the Rockies, the **Santa Ana** of southern California, or the **Mono** and northeast wind of central and northern California, are all associated with a high-pressure area in the Great Basin. A **foehn** is a wind flowing down the leeward side of mountain ranges where

air is forced across the ranges by the prevailing pressure gradient.

Subsidence occurs above the High where the air is warm and dry. The mountain ranges act as barriers to the flow of the lower layer of air so that the air crossing the ranges comes



Heating of the west coast marine layer as it moves inland on clear summer days may destroy the subsidence inversion. As a dry-adiabatic lapse rate is established, convective mixing can bring dry air from aloft down to the surface, and carry more moist air from the surface to higher levels.



Daytime convective currents may eat away the base of a subsidence inversion and mix some of the dry air above with the more humid air below. This process will warm and dry the surface layer slightly, but humidities cannot reach extremely low values unless the subsiding air reaches the surface.

from the dryer layer aloft. If the pressure gradient is favorable for removing the surface air on the leeward side of the mountain, the dry air from aloft is allowed to flow down the lee slopes to low elevations. **The dryness and warmth of this air combined with the strong wind flow produce the most critical fire-weather situations known anywhere.**

Mountain waves, most common and strongest in the West, are also characteristic of flow over eastern and other mountain ranges. When they occur with foehn winds, they create a very spotty pattern. The strongest winds and driest air are found where the mountain waves dip down to the surface on the leeward side of the mountains.

An example of a severe subsidence condition associated with chinook winds, and in which mountain waves probably played an important part, is the Denver, Colo., situation of December 1957. On December 9, chinook winds were reported all along the east slope of the Rocky Mountains in Wyoming and Colorado. Surface relative humidity at Denver remained at 3 percent or below from noon until midnight that day. The Denver observation at 1900 hours showed:

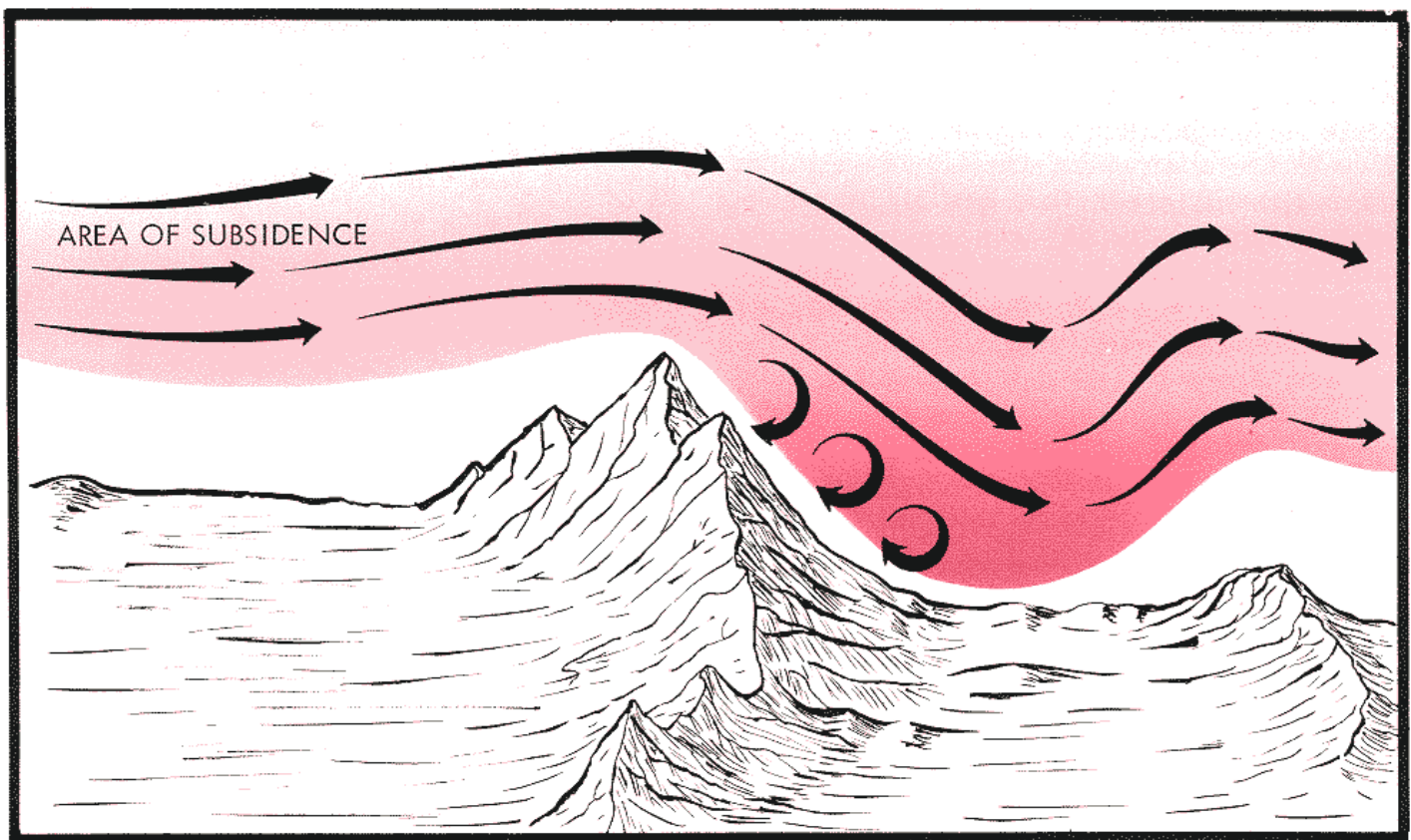
Temperature	Dew point	Relative humidity	Wind
(°F.)	(°F.)	(Percent)	(Direction) (m.p.h.)
60	-29	1	W 22

The extremely low dew point indicates that the air must have originated in the high troposphere.

Cases of severe subsidence are much more frequent in the western half of the country than in the eastern regions. Most of the Pacific coast area is affected in summer by the deep semi-permanent Pacific High. This provides a huge reservoir of dry, subsiding air which penetrates the continent in recurring surges to produce long periods of clear skies and dry weather. Fortunately, marine air persists much of the time in the lower layer along the immediate coast and partially modifies the subsiding air before it reaches the surface.

In the fall and winter months, the Great Basin High is a frequent source of subsiding air associated with the foehn winds, discussed above. It is the level of origin of this air that gives these winds their characteristic dryness.

Subsiding air reaching the surface is perhaps less common in eastern regions, but does occur from time to time. Usually the subsiding air is well modified by convection. But subsidence is often a factor in the severe fire weather found around the periphery of Highs moving into the region east of the Rockies from the Hudson Bay area or Northwest Canada mostly in spring and fall. It also occurs during summer and early fall periods of drought, when the Bermuda High extends well westward into the country.



Subsiding air above a High windward of a mountain range may be carried with the flow aloft and brought down to the leeward surface, with little modification, by mountain waves.

LOCAL INDICATORS OF STABILITY

The continent-wide network of weather stations that make regular upper-air soundings gives a broad general picture of the atmospheric structure over North America. These soundings show the major pressure, temperature, and moisture patterns that promote stability, instability, or subsidence, but they frequently do not provide an accurate description of the air over localities at appreciable distances from the upper-air stations. We need, therefore, to supplement these observations with local measurements or with helpful indicators.

At times, it may be possible to take upper-air observations with portable instruments in fixed-wing aircraft or helicopters. In mountainous country, temperature and humidity measurements taken at mountaintop and valley-bottom stations provide reasonable estimates of the lapse rate and moisture conditions in the air layer between the two levels. In areas where inversions form at night, similar measurements indicate the strength of the inversion. The heights of surface or low-level

inversions can be determined by traversing slopes that extend through them. The height at which rising smoke flattens out may indicate the base of a low-level inversion. The tops of clouds in the marine layer along the Pacific coast coincide with the base of the subsidence inversion. The height of the cloud tops provides a good estimate of the height of the inversion.

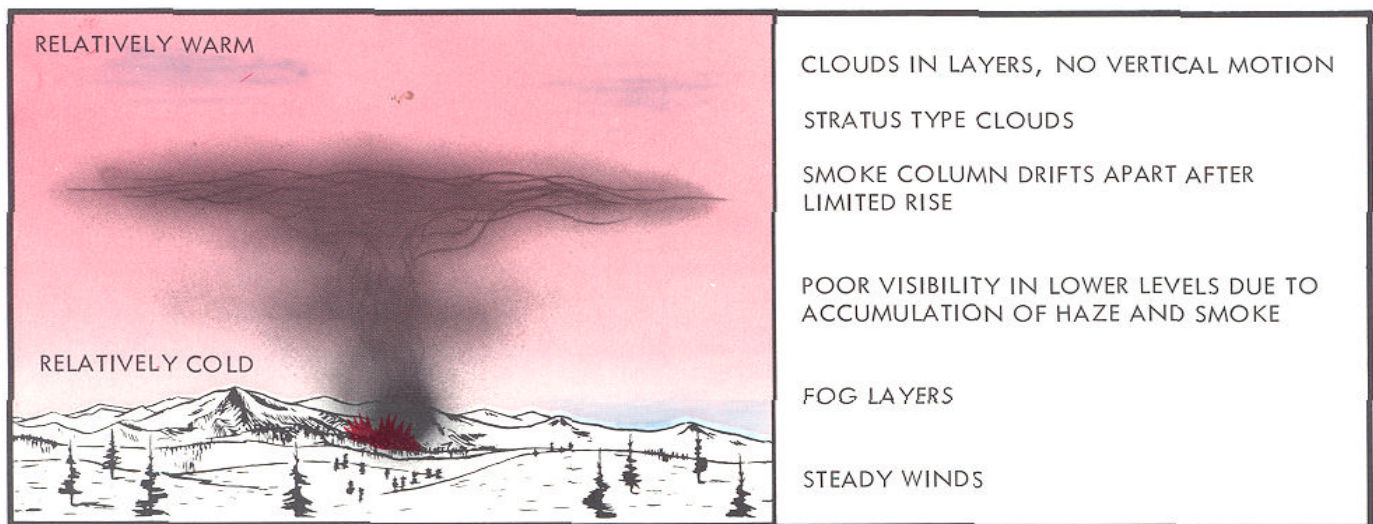
Other visual indicators are often quite revealing. Stability in the lower layers is indicated by the steadiness of the surface wind. A steady wind is indicative of stable air. Gusty wind, except where mechanical turbulence is the obvious cause, is typical of unstable air. Dust devils are always indicators of instability near the surface. Haze and smoke tend to hang near the ground in stable air and to disperse upward in unstable air.

Cloud types also indicate atmospheric sta-

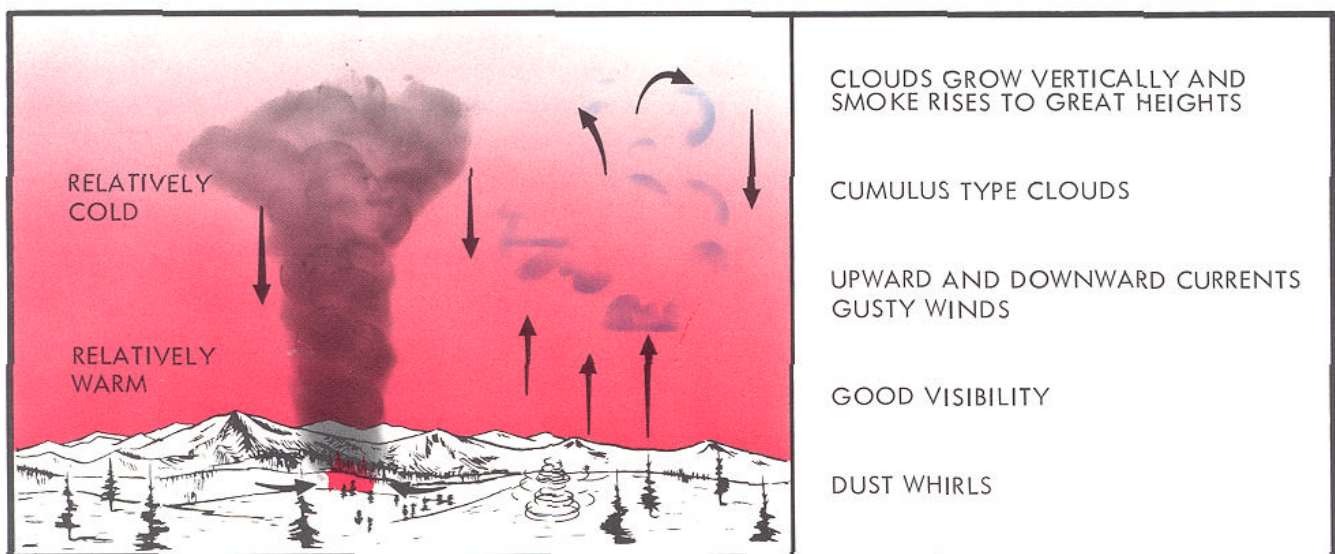
bility at their level. Cumulus-type clouds contain vertical currents and therefore indicate instability. The heights of cumulus clouds indicate the depth and intensity of the instability. The absence of cumulus clouds, however, does not necessarily mean that the air is stable. Intense summer heating can produce strong convective currents in the lower atmosphere, even if the air is too dry for condensation and cloud formation. Generally, though, the absence of clouds is a good indication that subsidence is occurring aloft. Even if scattered cumulus clouds are present during the day and are not developing vertically to any great extent, subsidence very likely is occurring above

the cumulus level. Stratus-type cloud sheets indicate stable layers in the atmosphere.

In mountainous country, where fire lookouts on high peaks take observations, a low dew-point temperature may provide the only advance warning of subsidence. Hygrothermograph records and wet- and dry-bulb temperature observations show a sharp drop in relative humidity with the arrival of subsiding air at the mountaintop. Early morning dew-point temperatures of 20°F. or lower in summer or early fall may signal the presence of subsiding air, and provide a warning of very low humidities at lower elevations in the afternoon.



Visible indicator of a stable atmosphere.



Visible indicator of a stable atmosphere.

SUMMARY

In this chapter we have seen how the distribution of temperature vertically in the troposphere influences vertical motion. A large decrease of temperature with height indicates an unstable condition which promotes up and down currents. A small decrease with height indicates a stable condition which inhibits vertical motion. Where the temperature increases with height, through an inversion, the atmosphere is extremely stable.

Between stable and unstable lapse rates we may have a conditionally unstable situation in which the atmosphere's stability depends upon whether or not the air is saturated. During condensation in saturated air, heat is released which warms the air and may produce instability; during evaporation, heat is absorbed and may increase stability.

Atmospheric stability varies with local heating, with wind speed, surface characteristics, warm- and cold air advection, and many

other factors. We can use type of cloud, wind-flow characteristics, occurrence of dust devils, and other phenomena as indicators of stability.

Subsidence is the gradual lowering of a layer of air over a broad area. When it begins at high levels in the troposphere, the air, which has little initial moisture, becomes increasingly warmer with resulting lower relative humidity as it approaches the surface. If some mechanism is present by which this warm, dry air can reach the surface, a very serious fire situation can result.

The first four chapters have been concerned with basic physical laws and with the statics of the atmosphere-its temperature and moisture and their distribution both horizontally and vertically, and to some extent its pressure. In the next chapter, we will consider pressure distributions more thoroughly and see how they are related to atmospheric circulation.