4

Fire Weather

Weather is an important leg of the fire environment triangle (see Figure 2.2). At times it can totally dominate the fire environment, overshadowing the influence of fuel and topography. Running crown fires can spread through mountainous terrain essentially without regard to topography. The influence of fuel moisture and fuel distribution can be neutralized when a fire is being driven by strong winds. On the other hand, the influence of weather can be more subtle. The diurnal variation in temperature and humidity, for example, has a significant influence on fuel moisture and therefore on fire behavior.

Weather is the state of the atmosphere and the changing nature of the atmosphere surrounding the earth. Fire weather includes those elements that influence wildland fire, both on the surface and up to 10 miles above the land.

Methods of fire behavior prediction were discussed in Chapter 2. The success of predicting fire behavior is in a large part dependent on the ability to forecast the weather, a difficult task even for a fire weather forecaster. Success in fire behavior prediction also depends on observation and interpretation, the ability to recognize patterns and interactions. Anyone involved with wildland fire should have a basic understanding of fire weather. Its interpretation is an art.

The question of scale in both time and space is important. Meteorology may be considered on a macro scale, the circulation of air around the globe; on a meso scale, the movement of particular air masses and large scale eddies and on a micro scale, where local heating and cooling differentials, site-specific winds, surface inversions, and microclimates are the objects of interest.

Weather changes and effects can be abrupt, as in the case of lightning strikes and microbursts. Fire weather is often viewed at an hourly or daily

scale. Changes throughout the year, in both weather and fuel, determine the *fire season*. A look at weather for many years determines the *fire climate*. And from the very long view, discussions of *global climate change* examine the role of wildland fire as both a cause and an effect of change that occurs on the order of centuries.

4.1 BASIC WEATHER PROCESSES

The Atmosphere

The atmosphere extends hundreds of miles above the earth, but thins so gradually that it is impossible to say exactly where it ends and space begins. The atmosphere is divided into several layers, based primarily on temperature characteristics (Figure 4.1).

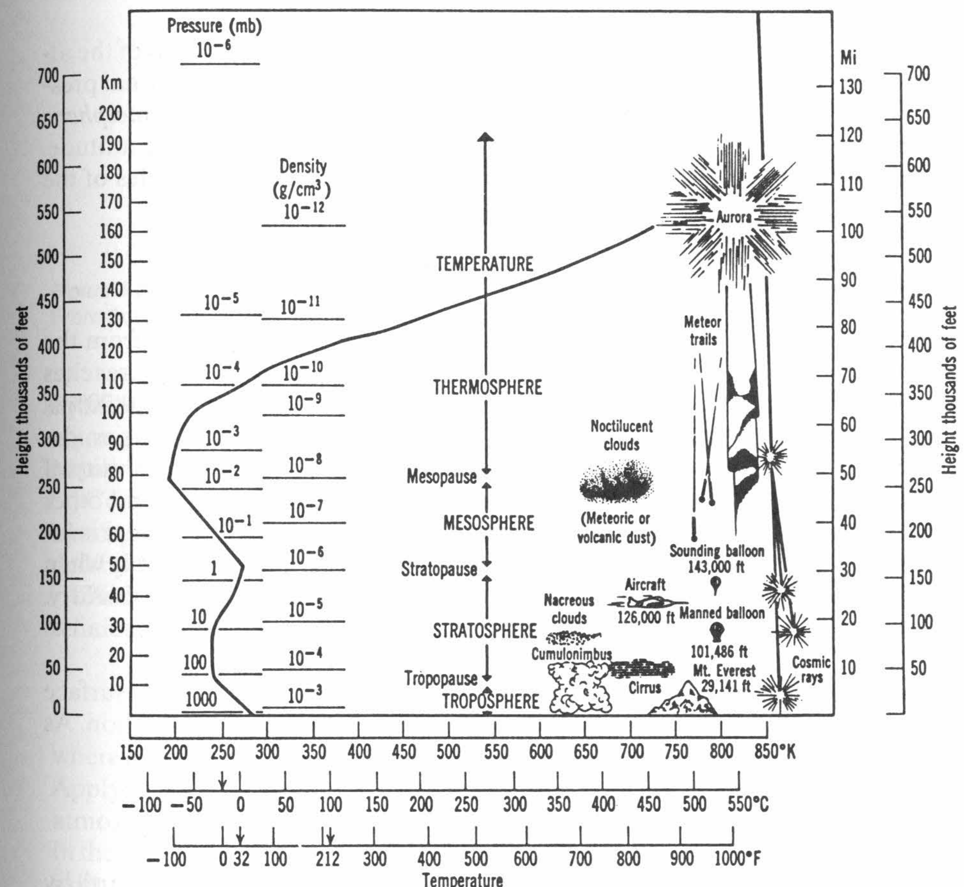


Figure 4.1. *Structure of the atmosphere.* From Cole (1975).

The troposphere is the lowest layer of the atmosphere. Temperature generally decreases with altitude, except for occasional shallow layers. Its depth varies from 5 miles over the North and South poles to about 10 miles over the equator. The transition between the troposphere and the stratosphere is the tropopause, the approximate top of thunderstorm activity.

The troposphere is the region of changeable weather and contains three-quarters of the earth's atmosphere by weight. We will concentrate our attention on this layer. Air in the troposphere is composed of primarily two gases. Dry air contains about 78% nitrogen by volume and 21% oxygen. Of the remainder, argon makes up about 0.93% and carbon dioxide between 0.03 and 0.04%. In addition to these gases, the troposphere contains a small, though significant, amount of water vapor, from near zero in desert and polar regions up to 4 or 5% in jungle regions. Water vapor has a very important effect on weather; without it, there would be no clouds or rain. The troposphere also contains salt and dust particles, smoke, and other industrial pollutants. These impurities affect the visibility through the atmosphere and also serve as nuclei for the condensation of water vapor and cloud formation.

The total weight of a 1-inch square of air from sea level to the top of the atmosphere averages 14.7 lb or 29.32 inches of mercury. This is the normal pressure exerted by the atmosphere at sea level and is called the *standard atmospheric pressure.* Atmospheric pressure rapidly decreases with increasing altitude. Nearly half of the weight of the atmosphere is within about 3.5 miles of the surface.

Earth's Heat Balance

Nearly all heating of the earth's surface and its atmosphere comes from the sun through solar radiation. On the average, 50% of the sun's energy reaches the earth's surface, 30% of the heat energy is reflected into space, and 20% is absorbed by the atmosphere. The distribution of solar radiation varies, depending on location and atmospheric conditions (Figure 4.2). The ability of the atmosphere to retain heat through absorption by water vapor and other gases is called the *greenhouse effect*. Thus, the drop in surface temperature is far less on cloudy nights than on clear nights. This effect is quite noticeable when comparing the rapid temperature fall at night in the desert, where the air is dry, with the slower decrease in temperature in coastal regions, where the air is moist.

The earth also loses heat by processes other than radiation. Some surface heat is transferred to the surrounding air by conduction and convection. As the ground cools, the air in contact with it also cools.

Atmospheric Pressure, Volume, and Temperature

The earth's weather circulation results from differential heating of the earth by the sun and by the process the atmosphere goes through in trying to come to

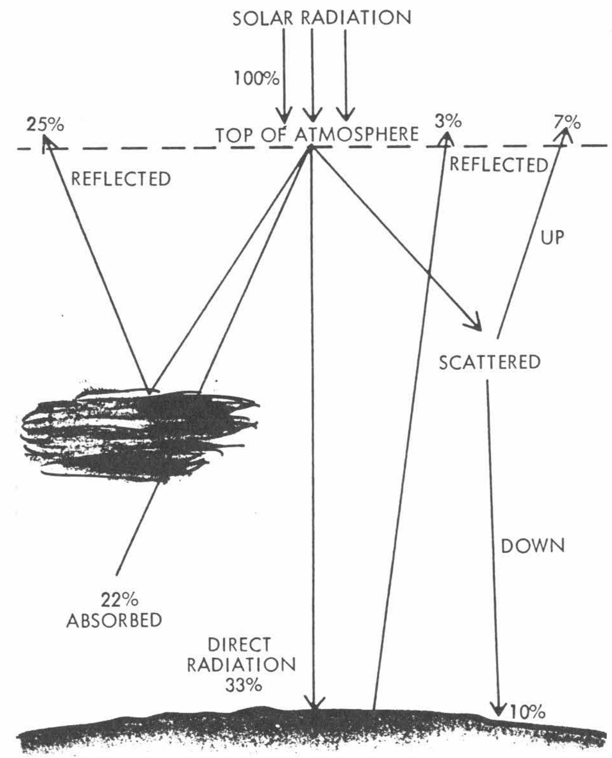


Figure 4.2. *Approximate distribution of incoming solar radiation during average cloudiness.* From Schroeder and Buck (1970).

equilibrium. Winds, storms, and clouds are the result of the relationships among atmospheric pressure, volume, temperature, and water vapor content. Many weather processes can be attributed to one or both of these two concepts: (1) Compression of the atmosphere results in warming and expansion of the atmosphere in cooling, and (2) colder air is denser and will sink when surrounded by warmer air.

The temperature of air changes as pressure and volume change. This relationship can be described by the Ideal Gas Equation:

$$pv = mRt$$

where p is pressure, v is volume, m is mass, t is temperature, and R is a constant. Applying this equation to air, we find that temperature and volume change as atmospheric pressure changes. For example, if a parcel of air is lifted vertically in the atmosphere, its volume will increase and its temperature will cool due to a decrease in atmospheric pressure. Conversely, a descending parcel of air will warm and decrease in volume because of increasing atmospheric pressure.

Thus, air expands and cools as it is lifted to higher elevations and warms and compresses as it descends to lower elevations.

The Ideal Gas Equation also shows another important relationship, this time in the horizontal. Since volume is related to density (mass per unit volume or m/v), we can say that warming is accompanied by a decrease in density, and cooling by an increase in density. Thus, colder air is denser and will sink when surrounded by warmer air.

Temperature-Humidity Relationships

When discussing atmospheric moisture, the amount present at any one time or place is commonly quantified as dew point temperature, relative humidity, and wet bulb temperature. *Dew point* is the temperature to which air must be cooled to reach its saturation point at constant pressure. It is a measure of the air's absolute humidity, or how much water vapor is in the air. If air cools to its dew point, condensation occurs and dew, fog, or clouds will form.

To determine exactly how dry or moist the air is at any given time or place, we use a unit of measure called relative humidity. *Relative humidity* is the ratio of the amount of moisture in the air to the amount that the air could hold at the same temperature if it were saturated. Relative humidity is always expressed as a percentage.

The amount of moisture in the air can also be determined by the *wet bulb temperature*, the lowest temperature to which the air can be cooled by evaporating water into it at constant pressure. The greater the difference between the wet and dry bulb temperatures, the drier the air.

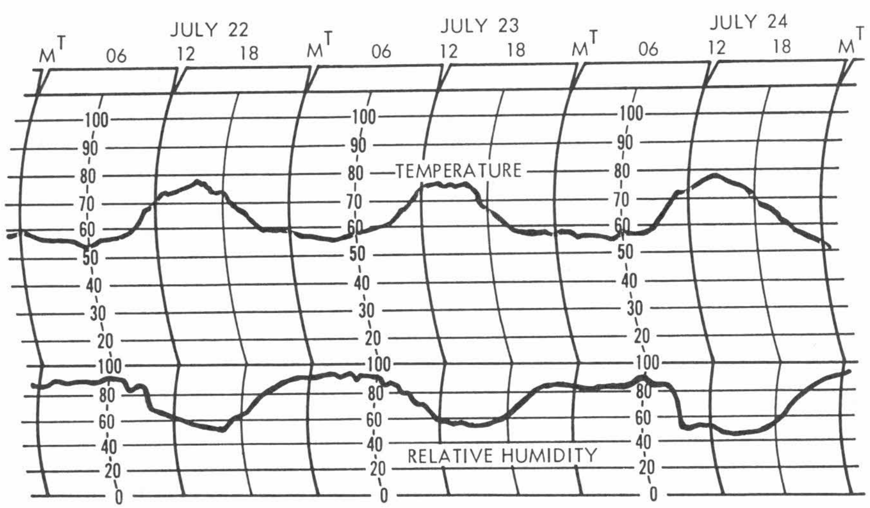


Figure 4.3. *Typical temperature and relative humidity traces for a low-level station are nearly mirror images of each other. They are less closely related at mountain stations.* From Schroeder and Buck (1970).

As temperature goes up at constant pressure, relative humidity goes down and vice versa. Maximum relative humidity generally occurs about sunrise, at the time of minimum temperature. After sunrise, humidity drops rapidly and reaches a minimum at about the time of maximum temperature. It rises more gradually from the late afternoon through the night. This is called the *diurnal variation* of temperature and humidity (Figure 4.3).

Changes in terrain, vegetation, clouds, and wind can create wide ranges in temperature and relative humidity over a small area. The controlling factor is how these parameters affect the amount of incoming and outgoing radiation that impacts the ground. Wind mixes the air near the ground, and vegetation intercepts incoming solar radiation during the day and outgoing radiation at night (Figure 4.4).

Daytime temperatures normally decrease with altitude with a corresponding rise in relative humidity. When nighttime cooling begins under clear skies, the temperature change with height is usually reversed. Cold air flows downslope and collects in the valley, resulting in colder air and higher relative humidity at lower elevations than at higher elevations.

4.2 ATMOSPHERIC STABILITY

Atmospheric stability is the resistance of the atmosphere to vertical motion. The earth's atmosphere is constantly moving and mixing. Air moves horizontally or vertically in response to the earth's rotation and to large and small changes in temperature and pressure. Wind is the horizontal movement of air. The vertical movement of air is related to stability. Depending on the temperature distribution in the atmosphere, air can rise, sink, or remain at the same level. Stable air resists vertical motion. Unstable air encourages vertical motion.

Lapse Rates

Lapse rate is the change in temperature with altitude. Normally, temperature decreases with altitude, but this can change as daytime solar heating and nighttime cooling change the temperature distribution of the lower atmosphere. The temperature lapse rate can range from a $+15°F/1000$ ft to a $-15°F/1000$ ft. The lapse rate continually changes as air circulates within the atmosphere and is warmed by compression or cooled by expansion. If this movement occurs without a transfer of heat or mass into the system, it is called an *adiabatic process.* When discussing atmospheric stability we are concerned with three lapse rates: dry lapse rate, moist lapse rate, and average lapse rate.

The *dry adiabatic lapse rate* is 5.5°F/1000 ft. A bubble or layer of dry (unsaturated) air will always have a temperature decrease of 5.5°F/1000 ft when lifted due to decreasing pressure and expansion. On the other hand, it will have a temperature increase of 5.5°F/1000 ft if it is forced downward and is compressed.

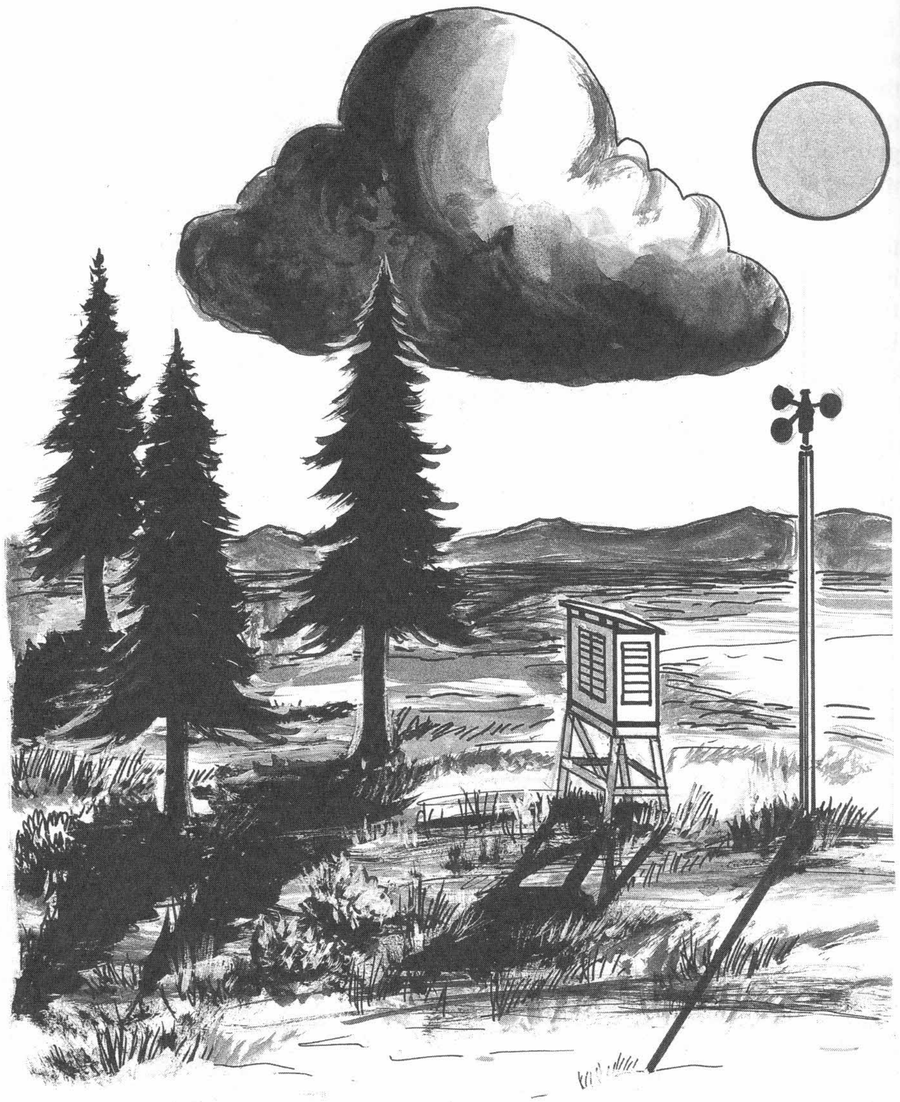


Figure 4.4. *Incoming solar radiation is affected by vegetation cover and clouds.* From Rothermel and others (1986).

The *moist adiabatic lapse rate* is 3°F/1000 ft. When air cools, its relative humidity increases. When a bubble or layer of air rises in the atmosphere, it cools with a corresponding rise in relative humidity. Air rises and cools at the dry adiabatic lapse rate until it reaches its saturation point. If it continues to rise, it will cool at the moist adiabatic lapse rate (Figure 4.5). If the relative humidity reaches 100%, the air becomes saturated and clouds form. This saturated air now cools at a lesser rate due to latent heat of condensation.

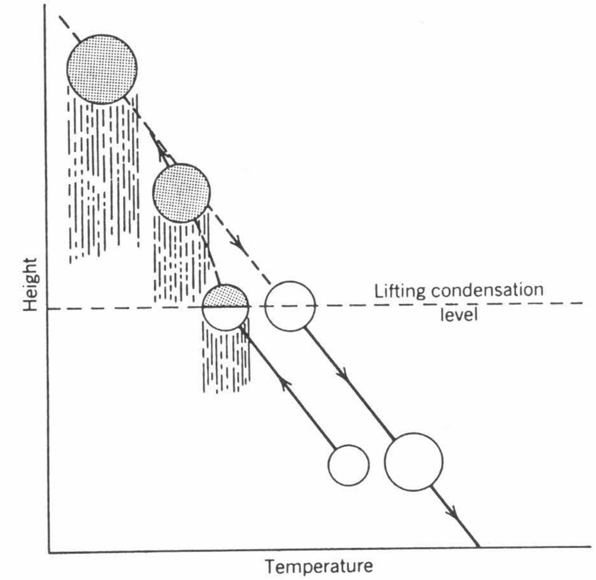


Figure 4.5. *A parcel of air is lifted according to the dry adiabatic rate until it reaches saturation, from which point it rises according to the wet adiabatic rate. Precipitation means that energy and mass have been lost from the system, so the descent process does not simply reverse the ascent. The parcel returns warmer and drier than it began.* From Cole (1975).

The atmosphere may or may not have a temperature distribution that fits the dry or moist lapse rates. Usually it does not. The average temperature change throughout the lower atmosphere over time and space is about 3.5°F/1000 ft.

Stability

Air may be stable or unstable, depending on its temperature lapse rate. If the temperature decrease in the air mass is greater than 5.5°F/1000 ft, the air is *unstable* (Figure 4.6). Unstable air encourages the vertical movement of air and tends to increase fire activity. If the temperature decrease in the air mass is less than 5.5°F/1000 ft, the air is *stable*. Stable air discourages the vertical movement of air and usually decreases or holds down fire activity.

Unstable air can contribute to increased fire behavior by increasing the chances of firewhirls, by increasing the potential for gusty surface winds, by increasing the heights and strengths of convection columns, and by increasing the chance of firebrands being lifted by the column. With unstable air, stronger winds aloft can be brought down to the lower atmosphere and produce stronger and gusty surface winds. There is better ventilation so air quality problems are usually minimized. Smoke convection columns rise much higher in unstable air. Chimneys, of a sort, develop, with indrafts feeding the

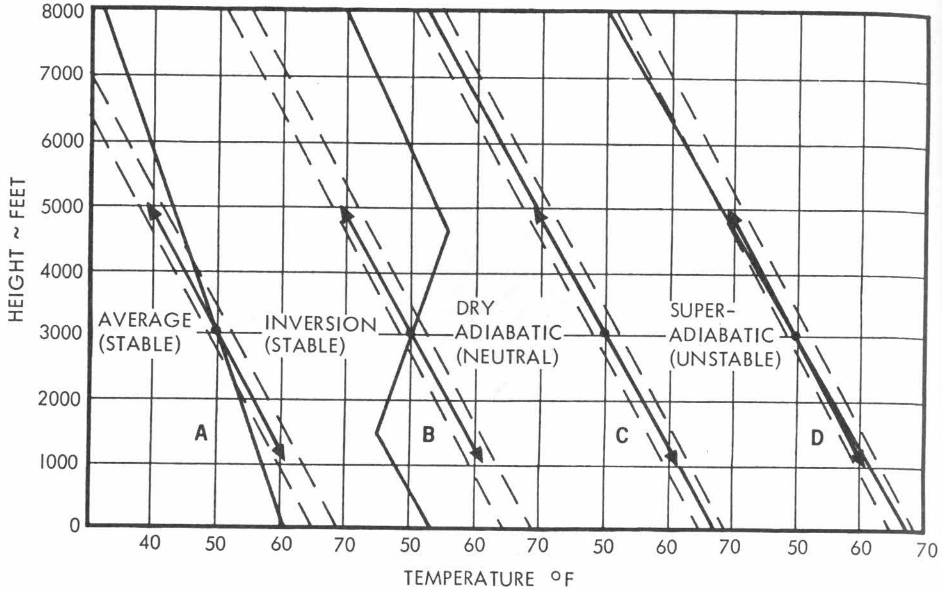


Figure 4.6. *In unsaturated air, the stability can be determined by comparing the measured lapse rate (solid lines) to the dry-adiabatic rate (dashed lines). The reaction of a parcel to lifting or lowering may be examined be comparing its temperature (arrows for parcel initially at 3000 feet and 50°F) to the temperature of its environment.* From Schroeder and Buck (1970).

fire at the base of the column and strong convective currents rising through the column. The greater the instability and fire intensity, the stronger the indrafts and convection column updrafts.

The *Haines Index* (1988) is a stability index designed for fire weather use. It is used to indicate the potential for wildfire growth by measuring the stability and dryness of the air over a fire. It is calculated by combining the stability and moisture content of the lower atmosphere into a number that correlates well with large fire growth. The stability term is determined by the temperature difference between two atmospheric layers; the moisture term is determined by the temperature and dew point difference. The Haines Index can range between 2 and 6. The drier and more unstable the lower atmosphere is, the higher is the Haines Index.

Dry air affects fire behavior by lowering fuel moisture. Instability can enhance the vertical size of the smoke column, increasing chances of a plume-driven crown fire (see Figure 2.21).

Inversions

An *inversion* is a layer of very stable air where the temperature increases with increase in altitude. Temperatures in an inversion may increase as much as

15°F/1000 ft in altitude. Inversions act as a lid and severely limit vertical motion in the atmosphere. Smoke will generally rise to the inversion, then flatten out and spread horizontally. There are three types of inversions, categorized by how they are formed and whether they are located at the earth's surface or aloft: radiation or nighttime inversion, marine inversion, and subsidence inversion.

Radiation or *nighttime inversions* are the most common type of inversion. They are formed when air is cooled at night, primarily by contact with the earth's surface, creating a condition of cool, heavier air below warmer air. The layer of cool air near the surface deepens as the night progresses. This produces a very stable condition, warm air above cool air. Conditions usually begin to reverse after sunrise. Inversions usually disappear sometime before noon as unstable conditions continue to develop. When the inversion dissipates and unstable conditions develop, fire activity can increase, sometimes rapidly. Winds may increase suddenly, temperatures increase, and relative humidity decrease.

The *marine inversion* is a common type of inversion found along the coast, particularly along the west coast of the United States. In this case, cool, moist air from the ocean spreads over low-lying land. The marine layer is topped by much warmer, drier, and relatively unstable air.

Subsidence inversions are associated with high-pressure systems in the upper atmosphere. Sinking air in a high-pressure system warms and dries as it descends to lower altitudes. This results in a layer of warm, dry air that becomes progressively warmer and drier as it drops closer to the surface. Subsidence is a slow process that occurs over a period of several days.

Normal daily changes in stability are related to temperature changes. For a typical summer day with clear skies and light winds, the lowest layers of the atmosphere are stable at night, with the greatest stability just before sunrise, and unstable during the late morning and afternoon, with the greatest instability during late afternoon, the hottest part of the day.

Normal seasonal variations in stability are related to the seasonal variations in temperature. Winter has more stable conditions than the other seasons due to colder temperatures and longer nights. Conversely, summer is the most unstable season due to warmer temperatures and longer hours of sunlight. Spring and fall are harder to define.

Stability affects the shape, growth, and size of smoke columns. If we hold factors such as fuel, topography, and other weather elements constant, unstable air will enhance the vertical development (shape), rapid expansion (growth), and increased proportions (size) of smoke columns as compared to stable conditions.

Lifting Processes and Thunderstorm Development

There are four lifting processes that can cause thunderstorm development: convection or thermal, orographic, frontal, and convergence (Figure 4.7).

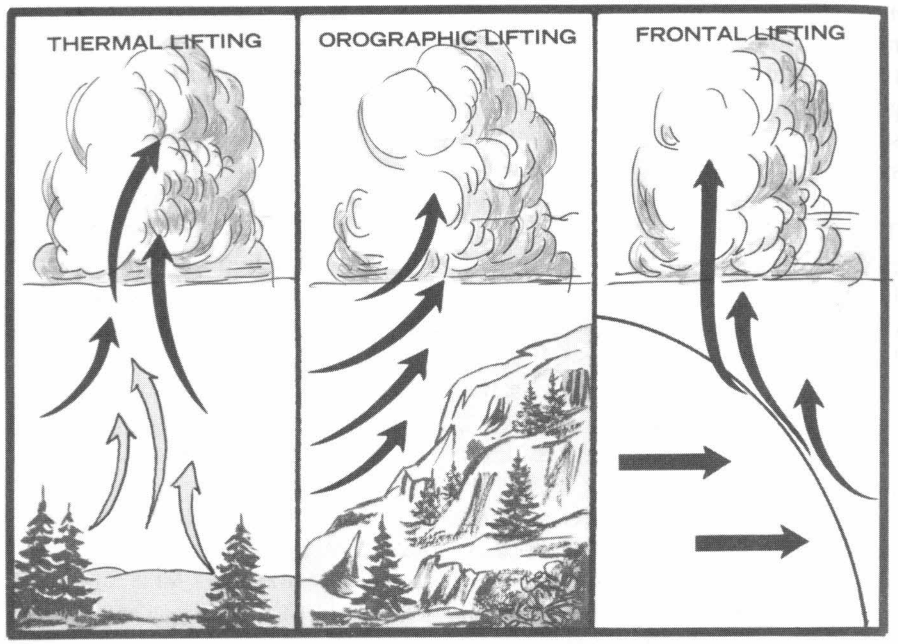


Figure 4.7. *Lifting processes that can cause thunderstorm development.* From Schroeder and Buck (1970).

Strong heating of air near the ground results in *thermal lifting*. As the air warms, it becomes buoyant and cools as it is forced aloft. If the heated air contains enough moisture and rises high enough in the atmosphere, condensation will occur and cumulus clouds will form.

Orographic lifting occurs in mountainous terrain when a mass of moving air is forced to rise because of the presence of a slope. Air that is forced upward cools at the dry lapse rate. If this air cools to its dew point temperature, clouds develop.

With the *frontal lifting* process a moving, cooler air mass pushes its way under and lifts a warmer air mass. Again, this lifting action can produce cumulus clouds if saturation occurs. Cumulus cloud development is usually associated with cold front passages, while stratus clouds generally accompany a warm front.

Convergence is present during all of the three preceding lifting processes, but it can also be an independent mechanism. Convergence occurs when more air moves horizontally into an area than moves out. The excess air is forced upward. Lifting by convergence always occurs around low-pressure systems or in areas of opposing wind directions. With more air flowing toward the center of the low, air piles up and is forced upward.

Thunderstorms have many variations in growth and behavior, but typically go through three stages of development and decay: cumulus, mature, and dis-

sipating stages (Figure 4.8). The *cumulus stage* starts with a rising column of moist air that develops a cumulus cloud that grows vertically. The towering cumulus cloud takes on a cauliflower-like appearance, with clear-cut tops. These clouds have strong indrafts into the base of the cloud that may increase surface winds. Updrafts in the cumulus or towering cumulus stage can cause wind directions to change as the air begins to move toward the developing cumulus cloud.

The *mature stage*, the most active portion of the thunderstorm cycle, begins when rain or virga starts falling out of the base of the cloud. The frictional drag exerted by the rain initiates a downdraft. There is now a downdraft in part of the cloud and an updraft in the remainder. The updraft is warmer and the downdraft is colder than the surrounding air in the cloud. An anvil-shaped layer, composed of ice crystals, forms at the top of the cloud and lightning occurs. It is now a thunderstorm and the cloud is called a cumulonimbus cloud. Downdrafts that reach the ground result in cool, gusty surface winds that can be experienced within about 10 miles or so of the thunderstorm.

During the *dissipating stage*, the entire thunderstorm becomes an area of downdrafts. As the updrafts end, the source of moisture and energy for continued growth and activity is cut off. Rain falls from the cloud, but becomes lighter and eventually ends. Gradually the downdraft weakens, the rain ends, and the cloud begins to dissipate.

Lightning

In fair weather, the atmosphere has a positive charge with respect to the earth. When a cumulus cloud grows into a cumulonimbus, the electrical fields in and near the cloud are altered and intensified. The charges are held on water drops and ice particles. The upper portion of the cloud becomes predominately positively charged and the lower portion becomes negatively charged. The negative charge near the cloud base induces a positive charge on the ground, a reversal of the fair-weather pattern (Figure 4.9).

Most cloud-to-ground discharges originate in the cloud and progress to the ground. They take place in two stages. First, a leader stroke begins with a negative discharge from the cloud and works its way downward to the ground in a series of probing steps. The negative core is surrounded by positive charges. When the core gets within 30 ft or so of the ground, a breakdown event occurs. The negative core and positive surrounding charges combine, making a very hot, rapid discharge that is seen by the eye as lightning.

Lightning sometimes occurs in the cumulus stage, but reaches its greatest frequency at the time the cell reaches maturity and its greatest height. The start of rain beneath the cloud base at the beginning of the mature stage marks the onset of the greatest lightning danger. Once lightning has started, it may continue well into the dissipating stage of the cell.

Most lightning discharges are within a cloud or cloud-to-cloud. Cloud-to-ground lightning is usually a discharge between the negative lower portion of

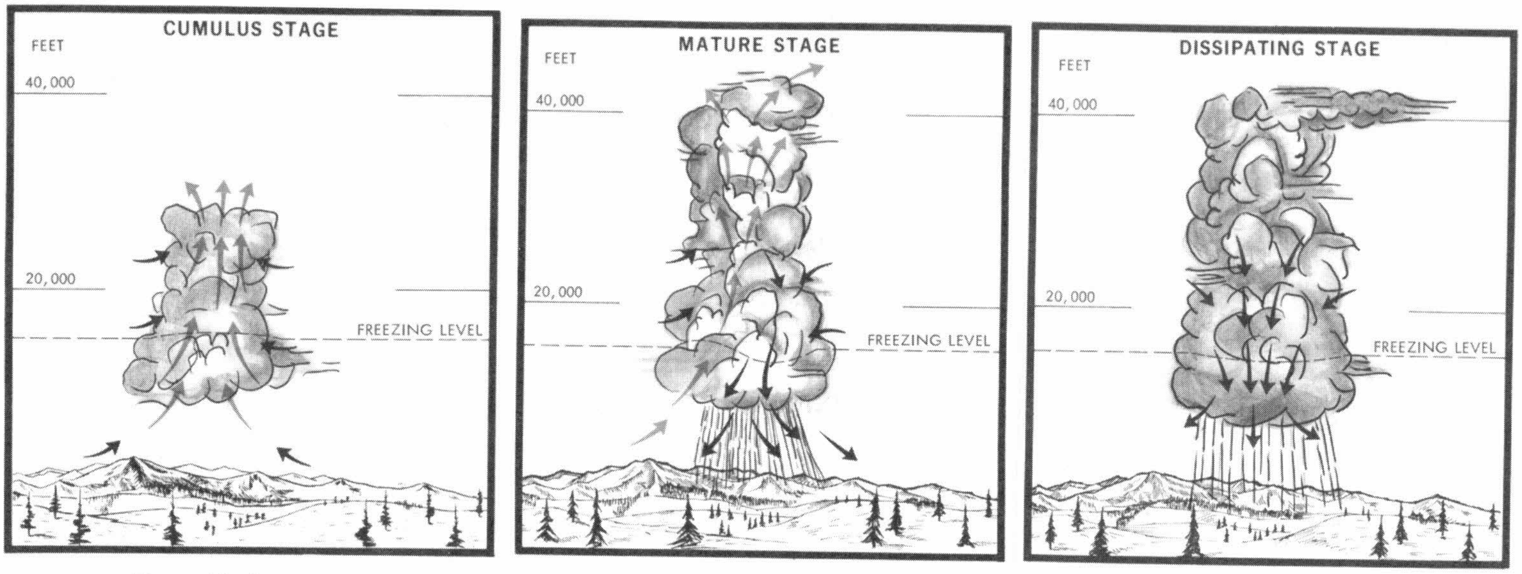
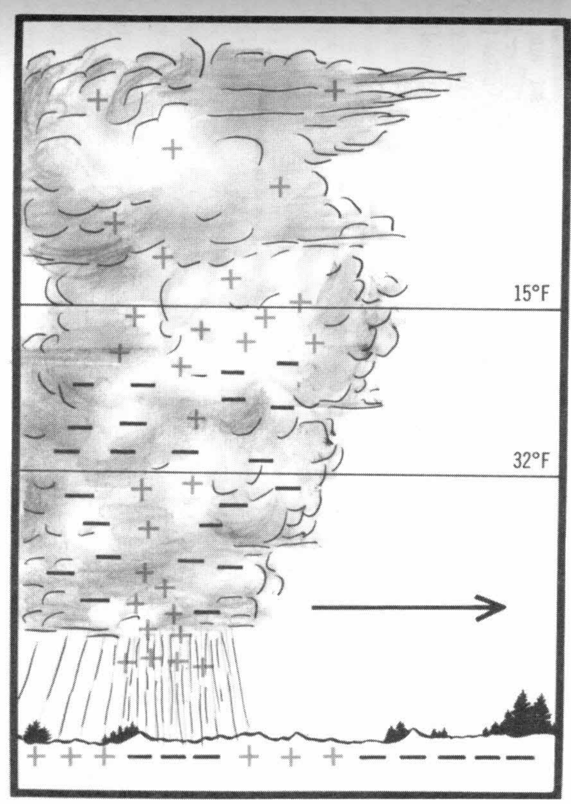


Figure 4.8. *Stages of thunderstorm development: cumulus, mature, and dissipating stages. From Schroeder and Buck (1970).*



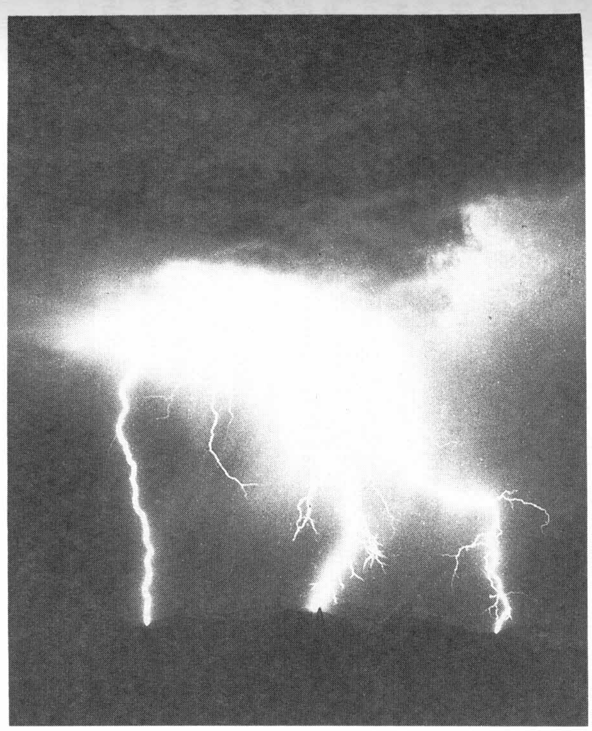


Figure 4.9. *Electrical charge distribution in a thunderhead.* From Schroeder and Buck (1970). Photo courtesy USDA Forest Service.

the cloud and the induced positive charge on the ground. There are, however, discharges that originate in the positive charge center near the top of the cloud or in anvils. In normal thunderstorms, about 5% of the discharges to ground are positive discharges. Positive discharges can occur at distances up to 30 miles from the cloud.

Discharges to ground have been separated into two types, positive and negative, according to the sign of the charge that is effectively moved from the cloud to the ground. These events are called positive and negative flashes. Each flash to the ground, whether positive or negative, is composed of several discrete events called strokes. Each stroke is a charge-moving event, starting with a leader and ending with a return discharge.

Few of the positive or negative lightning strokes to ground are suitable for fire ignitions. Fuquay and others (1972) showed that strokes that cause ignition must have a special component called a continuing current (or long continuing current). Flashes that have continuing currents are called hybrid flashes. The characteristic that is important for fire ignition is its duration. Continuing currents can last up to about 0.5 s; the median value is about 0.125 s. About 20% of negative flashes and 95% of positive discharges are hybrid flashes. Of all flashes to ground, then, about 25% have a continuing current.

4.3 WIND

By simple definition, *wind* is the horizontal movement of air relative to the earth's surface. In the study of wildland fire we are concerned with winds of two scales in the atmosphere: the larger scale general wind and the smaller scale local wind.

General Winds

All winds blow in response to pressure differences. In the very broad synoptic scale, the winds that are produced are called *general winds. Winds aloft* are the winds that blow in the upper atmosphere, unaffected by friction from terrain or other surface characteristics.

The surface of the earth is characteristically rough and will disrupt the winds due to frictional effects. This creates a turbulent zone next to the earth's surface that varies in thickness with the roughness of the surface and the speed of the wind. The *free air* or *gradient winds* are those winds that occur at the top of the frictional disturbance layer. The average depth or thickness of this layer may be quite shallow over uniform, flat terrain; or it can be as deep as 2000 to 3000 ft in complex, mountainous terrain. The frictional layer also varies in depth from day to day.

Temperature differences lead to pressure differences, called *pressure gradients*. Whenever horizontal pressure gradients exist, the air will be subject to a

force pushing it from higher toward lower pressure. This can be on a very large scale of hundreds of miles or on a very small scale of a few feet.

Several natural forces interact to produce continual movement of the air around the earth, thus producing widely varied weather patterns. Because the earth is not heated uniformly by the sun, resulting temperature differences lead to large-scale pressure differences. Warm air near the equator tends to rise and spread poleward in both hemispheres. Meanwhile, cold air at the poles sinks and flows toward the equator. As the air moves from the equatorial regions, it is affected by the earth's rotation. In the northern hemisphere, the air is deflected to the right as the earth rotates on its axis. This deflection is called the *Coriolis force*. The uneven heating and cooling of the earth is responsible for producing a series of *wind belts* that circle the earth at various latitudes. One of these wind belts is called the prevailing westerlies. Closer to the North Pole, high pressure is maintained and easterly winds occur.

At some latitudes, the air tends to pile up to cause belts of high surface pressure. These belts are never uniform, but instead consist of a series of rather large pressure cells. Some of the pressure cells are relatively fixed, such as the polar high, while others are migratory. Weather is closely dependent on the location and movement of these primary pressure cells and other smaller-scale pressure patterns. Surface pressure maps show cells of high pressure and cells of low pressure (Figure 4.10). Isobars are the lines of equal air pressure,

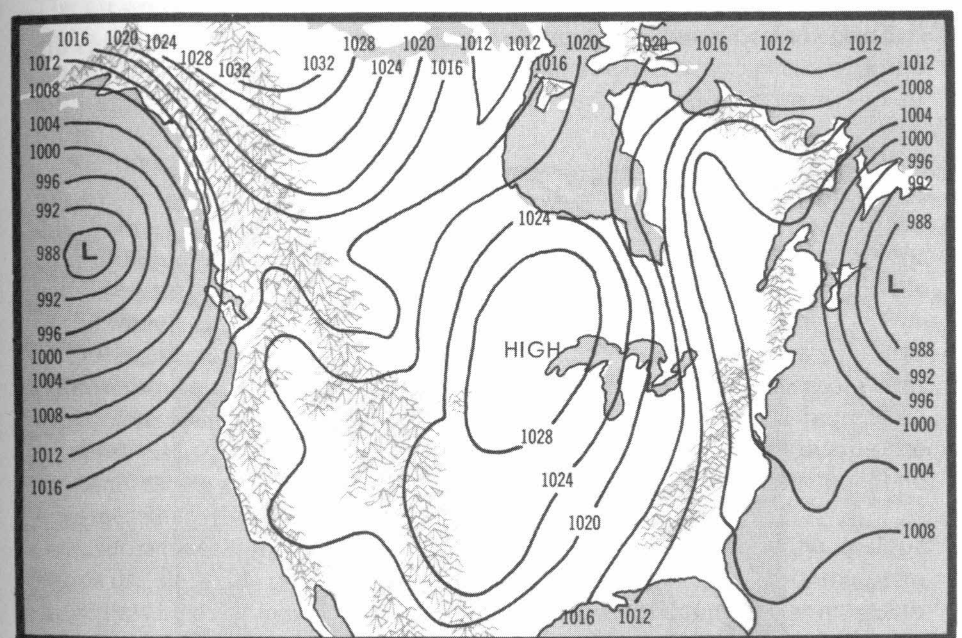


Figure 4.10. *A surface weather map is a graphical picture of the pressure distribution obtained by drawing lines, called isobars, through points of equal sea-level pressure. Isobars outline areas of high and low pressure.* From Schroeder and Buck (1970).

with their patterns outlining the areas of high and low pressure. The distance between isobars shows the pressure gradient. Windspeed is proportional to the gradient; where the gradient is tighter, the general winds are stronger, and where gradient is relaxed, the winds are lighter.

Typically, pressure cells move from west to east across the United States, being guided by the upper level westerlies. However, at times they can move in other directions, which further complicates weather patterns. Air in a high-pressure cell moves clockwise and outward from the cell; air in a low-pressure cell moves counterclockwise and toward the center of the cell. The boundary between two air masses of different temperatures and other characteristics is called a *weather front* (Figure 4.11). Typically, air masses are stable ahead of a front, unstable near the front, and stable behind the front.

Local Winds

Local or *convective winds* are smaller-scale winds caused by local temperature differences. Terrain has a very strong influence on local winds, and the more varied the terrain, the greater the influence. In many areas, the general winds are blocked by high terrain, and local winds are the predominant daily winds. Common local winds include sea and land breezes, slope winds, and valley winds.

As a result of local-scale temperature and pressure difference, a *sea breeze* begins to flow inland from over the water, forcing the warm air over the land to rise and cool adiabatically. In the absence of strong general winds, this air

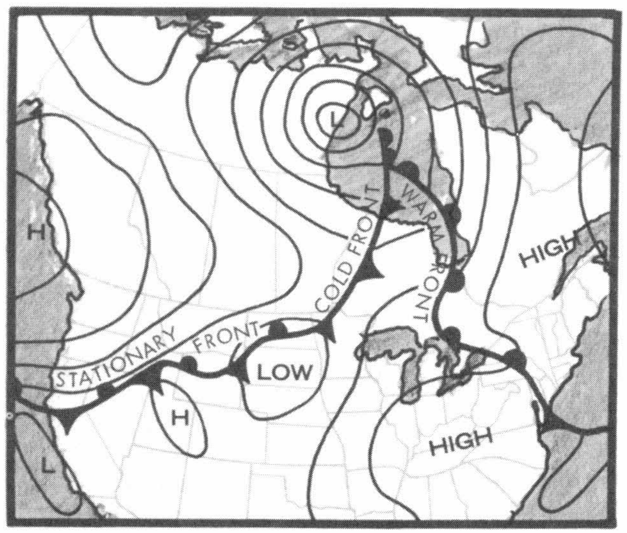


Figure 4.11. *Fronts are classified by the way they move relative to the air masses involved. At a cold front, cold air is replacing warm air. At a warm front, warm air is replacing cold air.*

flows seaward aloft to replace air that has settled and moved toward shore, and thus completes a circulation cell.

The sea breeze begins between midmorning and early afternoon, depending on time of year and location. It strengthens during the afternoon and then ends shortly after sunset. The sea breeze begins at the coast, then gradually pushes farther inland during the day, reaching its maximum penetration about the time of maximum heating. Cooler temperatures and higher relative humidities accompany the sea breeze as it moves inland. Typical speed of the sea breeze is 10 to 20 mi/h. However, it can locally attain 20 to 30 mi/h along the California, Oregon, and Washington coasts.

Along the Pacific coast, fog or low clouds, very cool temperatures, and high humidity accompany the sea breeze as it moves inland, usually resulting in diminished fire activity. In the southeastern United States, lines of thunderstorms frequently develop along the sea breeze as it moves inland from the coast. This results in strong shifting winds, cooler temperatures, higher relative humidities, and possible thundershowers—weather very similar to cold fronts. Strong shifting winds associated with these "sea breeze fronts" have caused control and safety problems for many fires in the Southeast.

The *land breeze* at night is the reverse of the daytime sea breeze circulation. At night, land surfaces cool more quickly than water surfaces. Air in contact with land then becomes cooler than air over adjacent water. Again, a difference in air pressure develops between air over the land and over the water. The air must be replaced, but return flow aloft is likely to be weak and diffuse and is diminished in the prevailing general winds. The land breeze begins 2 to 3 h after sunset and usually ends shortly after sunrise. Windspeeds with the land breeze are lighter than with the sea breeze, typically between 3 to 10 mi/h.

Another type of convective wind is the *slope wind*. Slope winds are local diurnal winds present on all sloping surfaces. They flow upslope during the day as the result of surface heating, and downslope at night due to surface cooling (Figure 4.12). Slope winds are produced by the local pressure gradient caused by the difference in temperature between the air near the slope and air at the same elevation away from the slope.

During the day, the warm air sheath next to the slope serves as a natural chimney and provides a path of least resistance for the upward flow of warm air, *upslope winds*. The layer of warm air is turbulent and buoyant, increasing in depth as it progresses up the slope. This process continues during the daytime as long as the slope is receiving solar radiation. When the slope becomes shaded or night comes, the process is reversed.

A short transition period occurs as a slope goes into shadow; the upslope winds die, there is a period of relative calm, and then a gentle, smooth downslope flow begins. *Downslope winds* are very shallow and may not be represented by an adjustment to the 20-ft windspeed (see Figure 2.6). The cooled dense air is stable, and the downslope flow tends to be quite smooth and slower than upslope winds. The principal force here is gravity. Downslope winds usually continue throughout the night until morning, when slopes are again warmed

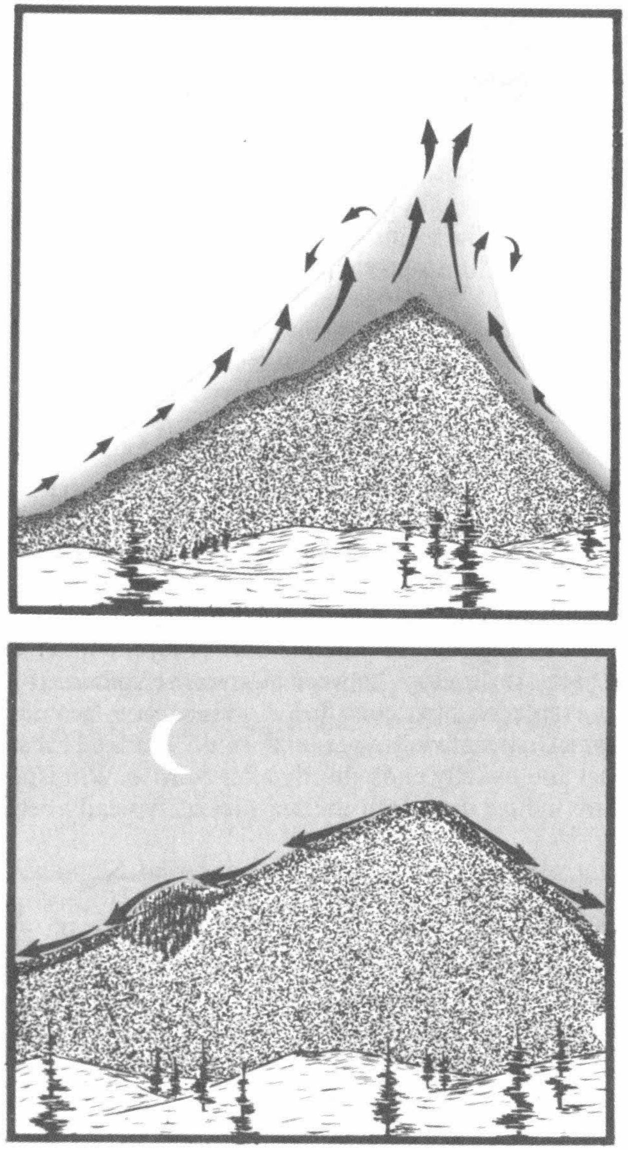


Figure 4.12. *Upslope winds are shallow near the base of slopes but increase in depth and speed as more heated air is funneled along the slope. Warm air bubbles forced upward cause turbulence which increases the depth of the warmed layer. Downslope winds are shallow, and the flow tends to be laminar. The cold air may be dammed by obstructions such as dense brush or timber.* From Schroeder and Buck (1970).

by solar radiation. The times during which winds change from downslope to upslope and vice versa depend on aspect, time of year, slope steepness, current weather conditions, and other lesser factors.

During the day, air in mountain valleys and canyons ends to become warmer than air at the same elevation over adjacent plains or larger valleys, thus creating a pressure gradient that results in *upvalley winds*. The main difference between upslope winds and upvalley winds is that the upvalley winds do not start until most of the air in the valley has warmed. Usually, this is late morning or early afternoon, depending largely on the size of the valley. These winds reach their maximum speeds by mid- or late afternoon and continue into the evening.

The transition from upvalley to *downvalley winds* takes place in the early night. The transition is gradual: first the downslope winds, then a pooling of cool, heavy air in the valley bottoms. The cool air in the higher valley bottoms will flow to lower elevations and increase in velocity as the pool of cool air deepens. This continues through the night and diminishes after sunrise.

The velocities of the slope and valley winds vary considerably with the terrain and current weather conditions. For example, slope and valley winds develop better under clear skies when the heating and cooling processes are more pronounced. Slope and valley winds are less pronounced and may not even develop under cloudy skies. Other factors to consider would be the length and steepness of the slope. Aspect is also important; north slopes typically have the lightest upslope winds due to their reduced insolation.

Effect of Topography on Windspeed and Direction

The several ways that topography can affect wind can be put into three categories: mechanical, turbulent, and frictional. There is some overlap.

Mechanical Effects The earth is solid, and the atmosphere is a fluid. When moving air collides with a topographic feature, such as a mountain range or peak, the air's motion is modified. Directional channeling, Venturi effect, and wave actions are all variations of mechanical and/or diverting effects of topography. *Directional channeling* is the case where a large drainage in a plateau region or any area allowing a general wind flow diverts some of the wind flow and sends it in a direction parallel to the drainage. Air pouring through a pass, saddle, or gorge will be speeded up due to the so-called *Venturi effect*. Air flows from higher toward lower pressure. At the surface, the higher pressure air is the cooler air mass, that is, it contains the heavier, denser air.

When strong winds move across a prominent mountain range in a direction perpendicular to the range, a wave form can be imparted to the general wind flow at and above the ridge crest. This is most likely when the air mass is stable aloft. These so-called *mountain waves* can extend many miles downstream as they gradually dampen out (Figure 4.13).

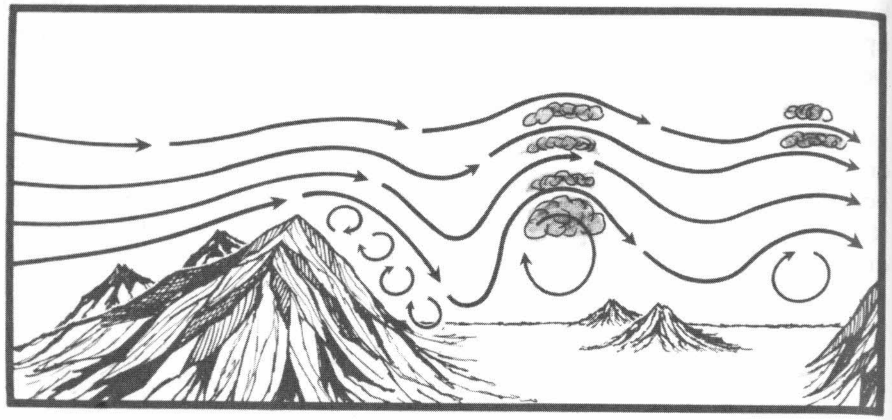


Figure 4.13. Mountain waves form when strong winds blow perpendicular to mountain ranges. Considerable turbulence and strong updrafts and downdrafts are found on the lee side. Crests of waves may be marked by lens-shaped wave clouds, but at times there may be insufficient moisture to form clouds. From Schroeder and Buck (1970).

Below the ridge line on the lee side, the air motions may be quite turbulent. The danger with mountain waves is that as daily heating in the lee-side basins progresses, the air there become increasingly unstable. Quite often, the mixed layer will become deep enough to link up to the level of the mountain waves, and then pull the strong winds down to the surface. Another atmospheric factor that contributes is the presence of strong high pressure and associated subsidence on the windward side of the range.

Turbulent Effects Whenever airflow is diverted over or around a prominent obstruction, it is unlikely to make a smooth transition back into a smooth, unified wind. Zones of turbulence known as *eddies* will usually form on the lee side of a significant obstruction to the wind. These eddies may be in the vertical plane or the horizontal plane. Eddies often form at the confluence of tributaries during strong canyon or valley winds. Another turbulent place is to the lee of spur ridges extending down into the main canyon. This effect would be most pronounced during late afternoon, when local upcanyon winds are at their peak.

Thermal turbulence is caused by differential surface heating, and it can have a great deal of effect on the low-level wind. Different land surfaces absorb, reflect, and radiate varying amounts of heat. Warm air rises and mixes with other air moving across the terrain. This mixing action has differing effects on surface winds, but often makes them gusty and erratic.

Frictional Drag All types of winds are slowed down by the drag caused by friction as they approach the earth's surface. Varying surface roughness causes varying amounts of frictional drag.

Winds of Most Importance to Wildland Fire

The winds discussed in this section can produce severe fire weather conditions. Cold front winds, foehn winds, thunderstorm downdrafts, microbursts, and low-level jets can totally dominate the fire environment.

Cold Front Winds A weather front is the boundary zone between two adjacent air masses. With a cold front, the colder denser air mass behind the front actively displaces the warmer, less dense air ahead of the front (Figure 4.14). A typical United States cold front moves at 20 to 30 mi/h, but speeds can be considerably faster or slower. Winds ahead of an approaching cold front usually shift gradually from southeast to south, and on to southwest. As the cold front passes, winds shift rapidly to west, then northwest (Figure 4.15). Windspeeds increase in strength as a front approaches, and usually become quite strong and gusty when the front passes an area. This is because pressure gradients are tight, and strong upper winds are more easily mixed down to the surface in very unstable air. Typical cold front windspeeds range between 15 and 25 mi/h, but can be much higher with strong cold fronts.

Cold fronts may bring thunderstorm activity, with possible precipitation. However, during the summer months in the western United States, cold fronts are often dry. They have moderate to strong winds and cooler air, but not enough moisture for precipitation. Warm fronts are relatively weak compared to cold fronts, and have little significance during fire season.

Foehn Winds Foehn winds are a special case of general winds, associated with mountain range systems. They occur as heavy, stable air pushes across a mountain range and then descends the slopes on the leeward side, becoming warmer and drier due to compression. Foehn winds tend to be stronger at night because they combine with local downcanyon winds.

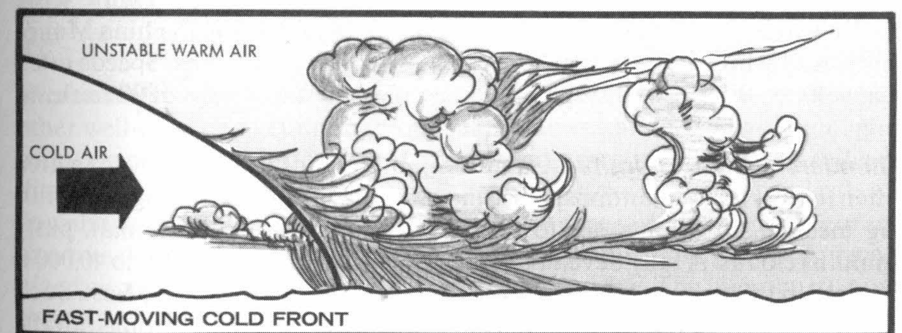


Figure 4.14. With rapidly moving cold fronts, the weather is more severe than with slow-moving cold fronts. If the warm air is moist and conditionally stable, as in this case, then scattered showers and thunderstorms form just ahead of the cold front. From Schroeder and Buck (1970).

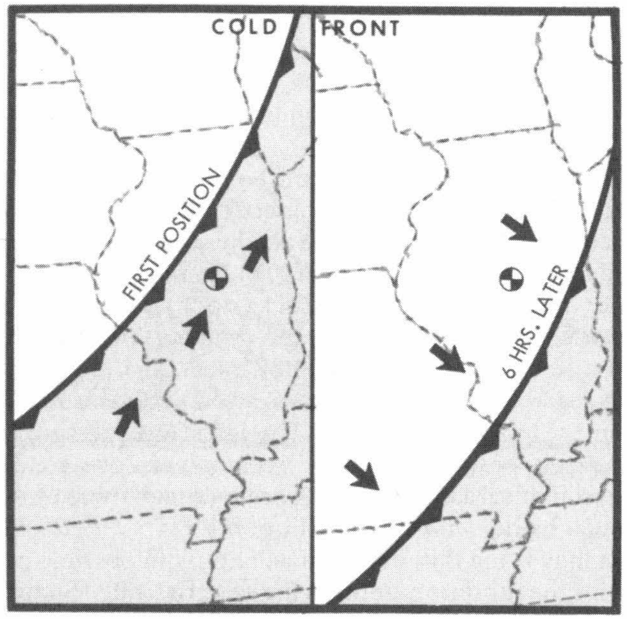


Figure 4.15. *Winds increase ahead of a cold front, become gusty and shift abruptly, usually from a southwesterly to a northwesterly direction, as the front passes.* From Schroeder and Buck (1970).

The Santa Ana wind creates critical fire weather situations in areas of southern California during fall and winter. The chinook wind occurs on the east slopes of several large mountain ranges in the western United States. Chinook winds are most prevalent on the east side of the Rocky Mountains during fall and winter. Among other well-known foehn winds in the western United States are East winds, North winds, and Mono winds (Figure 4.16). Foehn winds can also occur on the eastern slopes of the Appalachian Mountains. All foehn winds can cause serious fire control problems. Speeds often reach 40 to 60 mi/h, and some have been measured in excess of 90 mi/h.

Thunderstorm Downdrafts Cumulonimbus clouds can build over an area when there is adequate atmospheric moisture and instability, along with a lifting mechanism to force air to rise. Thunderstorms begin as small, puffy cumulus clouds. A fully developed thunderstorm can reach 30,000 to 40,000 ft or more in the western United States, and 60,000 to 70,000 ft in the East. Such clouds have vast amounts of stored energy. Not only are there strong indrafts into the base of the cloud, but strong downdrafts occur with the release of its energy.

The bases of mature thunderstorms are ragged from downdrafts and virga. Air moving down out of the thunderstorm base is cooled by evaporation, becoming much heavier than surrounding air. It also accelerates due to

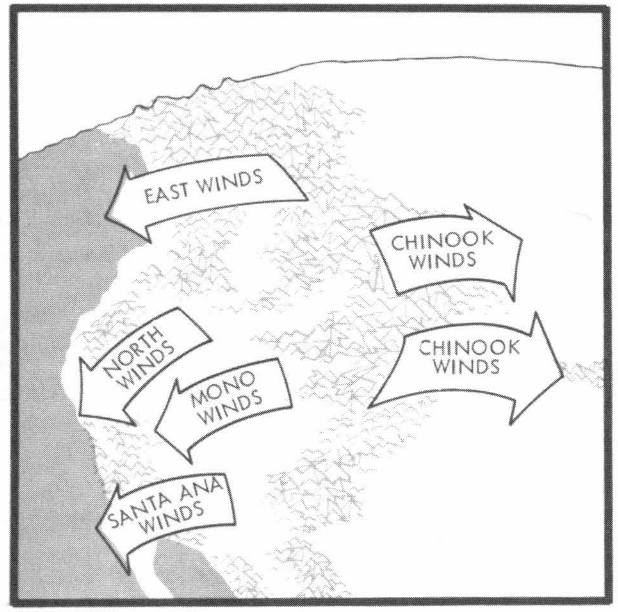


Figure 4.16. *Foehn winds are known by different names in different parts of the mountainous West. In each case, air is blowing from a high-pressure area on the windward side of the mountains to the low-pressusre or trough area on the leeward side.* From Schroeder and Buck (1970).

gravity. Downdrafts that reach the ground usually spread radially in all directions. This results in cool, gusty surface winds (often referred to as a *gust front*) that can be experienced within about 5 to 10 miles of the thunderstorm. In mountainous terrain, this distance can be considerably farther due to channeling by ridges and/or canyons. Surface wind velocities will often be 25 to 35 mi/h. Thunderstorm downdrafts will be cooler and somewhat more moist than surrounding air.

Microbursts A *downburst* is a downdraft associated with a thunderstorm or other well-developed cumulus clouds that induces an outburst of damaging winds on or near the ground. When the downburst is small (0.25–2.5 miles in diameter), it is a *microburst*, larger ones are *macrobursts*. Not all downdrafts are downbursts. Horizontal windspeeds generally exceed 40 mi/h on the ground in a true downburst. Downdrafts from thunderstorms have long been recognized as contributing to the dynamics of wildland fire. The microburst, however, was first identified in 1974.

A great variety of conditions can produce microbursts, but two extreme situations seem to create them in large numbers—wet and dry. It is believed that both types require raindrops as an initial condition because evaporation of these drops cools the air, which then falls as it gets heavier (Figure 4.17). Humid areas, like the southeastern United States, usually experience wet mic-

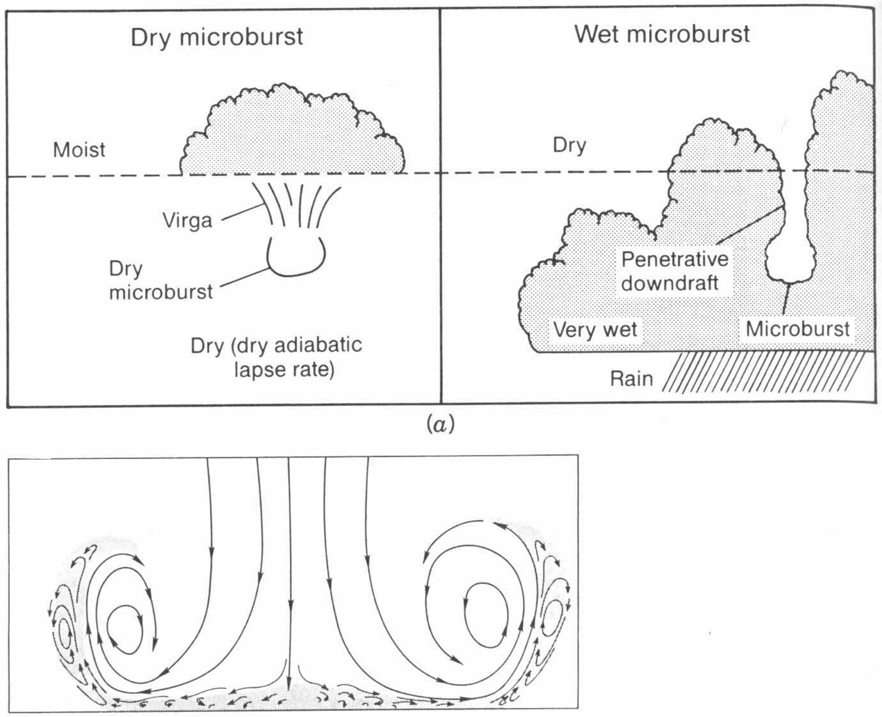


Figure 4.17. *(a) Conceptual models of wet and dry microbursts. (b) Cross section of a conceptual vortex ring model of a microburst.* From Caracena and others (1990).

roburst, associated with moderate to heavy rain. Haines (1988a) described a 1981 Florida wildland fire that killed two firefighters. Following the microburst, a heavy rainstorm, lasting for 15 to 20 min, nearly extinguished the wildfire.

Dry microbursts occur in a very dry environment, as is common over the semiarid western Great Plains and intermountain region, where cloud bases are often higher and precipitation may evaporate before it reaches the ground. The evaporative cooling of the air intensifies the downdraft, producing a microburst. Near the cloud base, winds and rain converge around the descending air, feeding into it.

Microbursts can be generated by a convection column over a fire as well as by a cumulus cloud. The resulting strong winds can instantly change the character of a fire from a low intensity surface fire to an uncontrollable crown fire. The occurrence of precipitation or virga may indicate the development of a microburst from a plume-dominated fire.

Low-Level Jet Streams These are currents of relatively fast moving air near the earth's surface. They are similar to the well-known jet stream in the upper

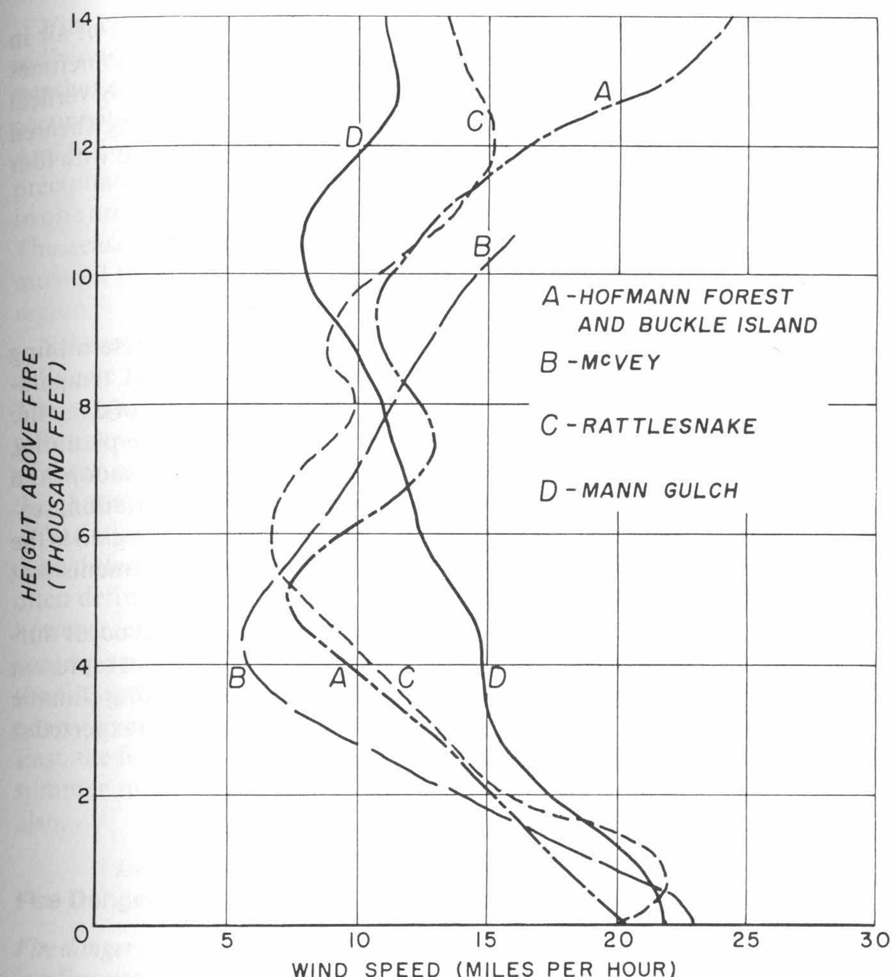


Figure 4.18. *Wind profiles associated with four fires that exhibited extreme fire behavior. The wind profile was taken from the station nearest the fire area: (A) Charleston, South Carolina, 17 April 1950; (B) Rapid City, South Dakota, 10 July 1939; (C) Red Bluff, California, 10 July 1953; (D) Great Falls, Montana, 5 August 1949.* From Byram (1954).

atmosphere. However, the maximum windspeed is considerably less, typically ranging from 25 to 35 mi/h. The maximum speed comes at an altitude from 100 ft to several thousand feet above ground level. They are most often observed in the central plains of the United States ahead of cold fronts. Low-level jet streams can cause rapid fire spread when they drop to the earth's surface. And they can increase surface wind speeds and gustiness through the downward transport of momentum. Of the 64 large fires examined by Haines (1988a), 47% of the eastern fires and 27% of the western fires were associated with low-level jets.

Because windspeed normally increases with altitude, a stratum of air in which the lower layers are moving faster than the upper layers is sometimes called a "reverse wind profile." A reverse wind profile allows a strong vertical convection column to develop directly over the fire without being sheared away by winds aloft. Figure 4.18 shows the wind profiles associated with four fires that exhibited extreme behavior.

4.4 FIRE CLIMATE AND FIRE SEASON

Fire weather occurring on a particular day is a dominant factor in determining the fire potential on that day. Fire weather elements include wind, temperature, humidity, precipitation, and so on. *Fire climate* is a synthesis of daily fire weather over a long period of time, an important concept for fire planning, preparedness, and prescribed fire. Climate largely determines the amount and kind of vegetation in an area; and climate sets the pattern of variation, seasonally and between one year and another. The fire climate of a region is the composite or integration over a period of time of the weather elements that affect fire behavior.

Weather refers to the shorter-term atmospheric conditions that occur during individual days or months in individual years. Climatic statistics represent a synthesis of this weather over a period of many years. The resulting climatic description thus serves to indicate the range of weather that may be expected at future time (Figure 4.19).

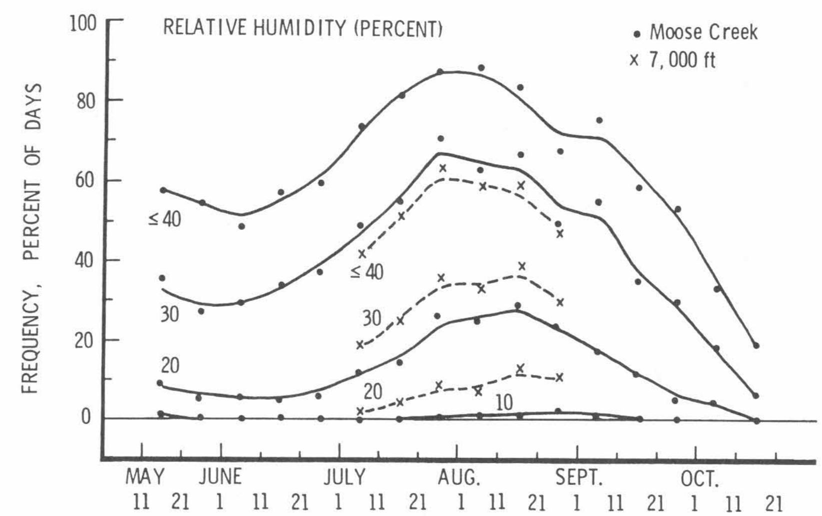


Figure 4.19. *Fire season frequencies of specified 10-day average relative humidity at 1500 PST, 1951–1970; Moose Creek Ranger Station (elevation 2460 ft) and 7000 ft ridgetop. Selway-Bitterroot Wilderness, Idaho.* From Finklin (1983).

Fire climate cannot be described by considering the weather elements individually. Fire potential responds to the combined effects of all of the fire-weather elements. For example, a region may have strong winds, but if they occur with precipitation, they are of much less importance to the fire climate. And fire climate is more than averages. Two areas may have the same annual precipitation; but the amount may be evenly distributed throughout the year in one area and concentrated during one portion of the year in another area. The seasonal distribution, the extremes, the frequency, and the duration must all be considered in describing precipitation in the fire climate of a region.

Schroeder and Buck (1970) considered geographic and climate factors to delineate 15 broad regions over the North American continent (Figure 4.20). These regions differ in one or more aspects, giving each a distinctive character affecting wildland fire. There is, of course, much local variation. A description of three of the regions is included in the selected examples at the end of the chapter.

Variations in climate, along with variations in vegetative conditions, produce differences in the *fire seasons* from one region to the next. Fire season is often defined in terms of wildfire occurrence. The use of prescribed fire can, however, effectively extend the fire season. In general, the fire season in the western and northern regions of the continent occurs in the summertime. But the fire season becomes longer as one goes from north to south, becoming nearly a year-round season in the Southwest and southern California. In the East, the fire season peaks in the spring and fall. Some fires occur during the summer months, and in the southeastern states they can occur in winter also.

Fire Danger Rating

Fire danger rating is an integration of weather elements and other factors affecting fire potential. In many fire danger rating systems, only the weather elements are considered. The other legs of the fire environment triangle (Figure 2.2), fuel and topography, are assumed constant. Fire danger rating systems produce numeric indexes of fire potential that are used as guides in a variety of fire management activities, including staffing for fire control, scheduling prescribed fire, and fire prevention.

Analysis of day-to-day fire weather, i.e., fire danger rating, offers a way to track the fire season and to compare one season to another. Seasonal plots of a fire danger index are useful in visualizing how the season is progressing and in comparing one year with another, or to seasonal average or extreme values (Figure 4.21).

Site-specific weather is needed to project the behavior of a specific fire. Fire danger rating, on the other hand, uses weather observations at a fixed site to give a broad area assessment of fire potential. The difference between fire behavior and fire danger is essentially a matter of scale.

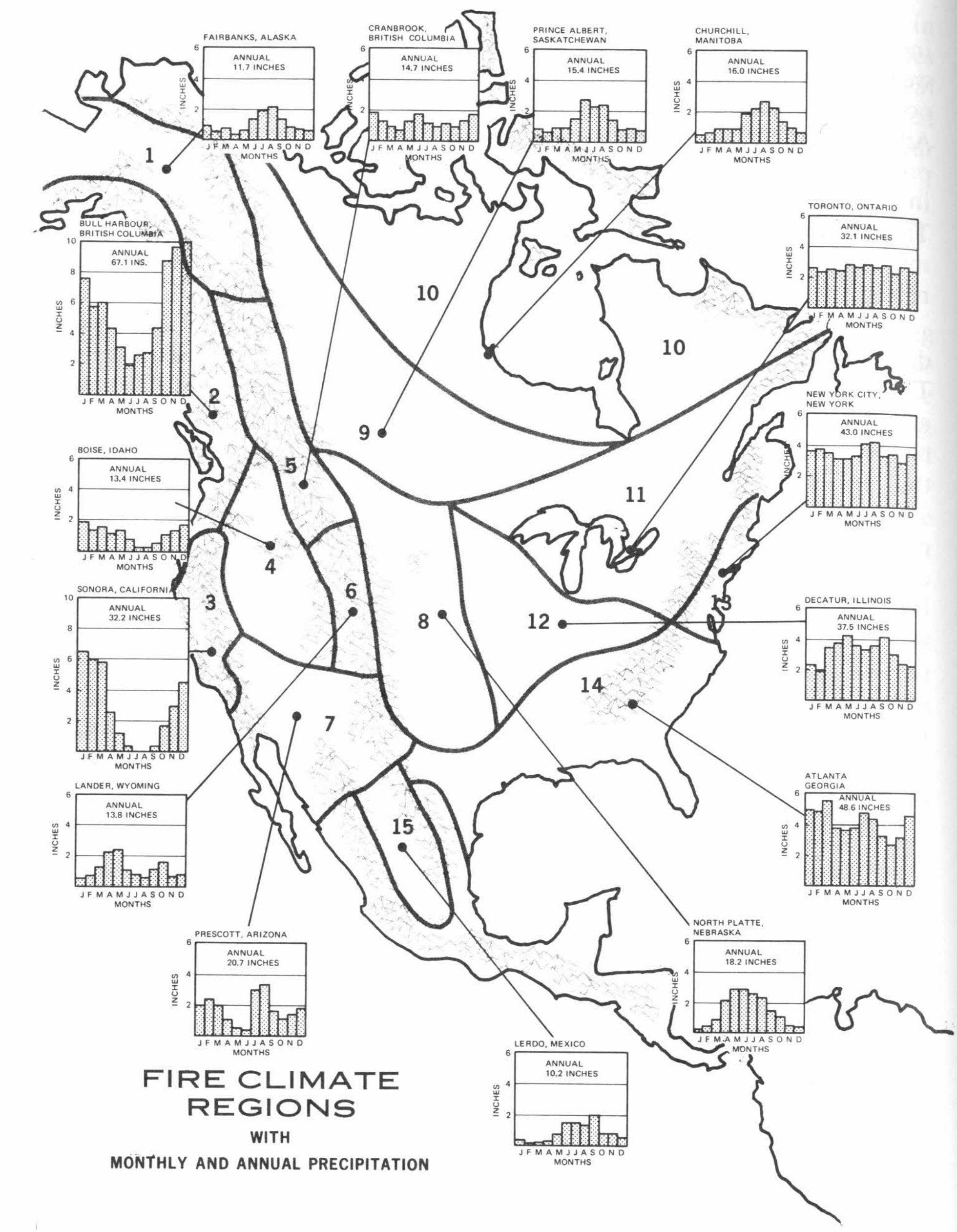


Figure 4.20. *Fire climate regions of North America. The associated bargraphs give the monthly and annual precipitation for a representative station.* From Schroeder and Buck (1970).

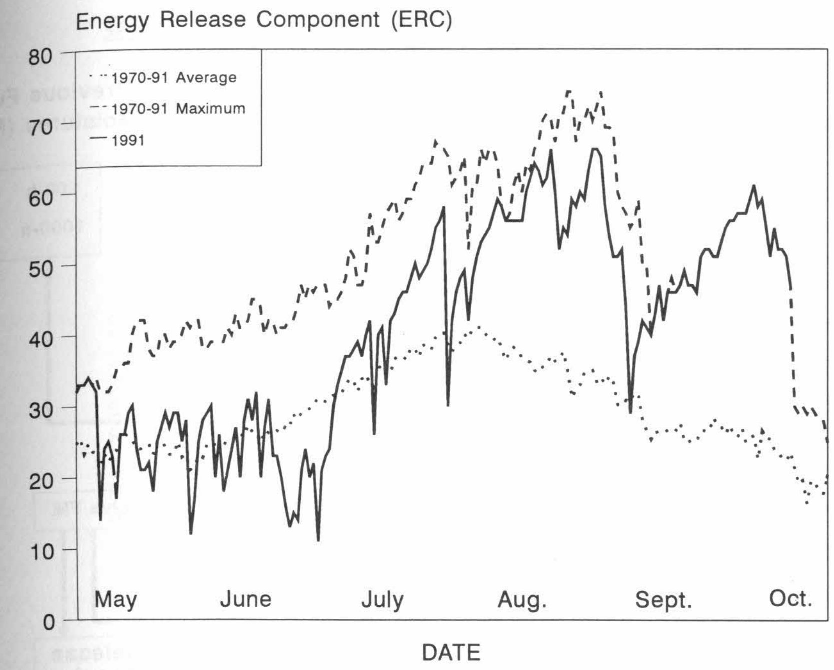


Figure 4.21. *Seasonal plot of Energy Release Component (ERC) for the Missoula, Montana area showing the 1991 fire season compared to the average and maximum for 1970–1991.* From Andrews and Bradshaw (1992).

Several successful fire danger rating systems are in use around the world. We discuss here only two North American systems, the United States National Fire Danger Rating System and the Canadian Fire Weather Index System.

U.S. National Fire Danger Rating System (NFDRS) The main indexes produced by NFDRS are Spread Component (SC), Energy Release Component (ERC), and Burning Index (BI) (Figure 4.22). Other indexes, such as lightning and human occurrence indexes and a fire load index, were part of the original formulation of NFDRS; they are rarely used and are therefore not discussed here. The Keetch–Byram Drought Index (KBDI) is now part of NFDRS and is discussed in the next section. The calculation of SC is related to spread rate, ERC to energy release, and BI to flame length or fireline intensity.

Because NFDRS provides an indication of fire potential over large administrative areas, a general description of the area is appropriate. Fuel and topography are assumed constant over the area. Site description includes slope class (choice of 5), fuel model (choice of 20), and a designation of the herbaceous fuel type as either annual or perennial.

Each afternoon at 1400 local standard time (LST), the following weather elements are measured: relative humidity, air temperature, state of weather

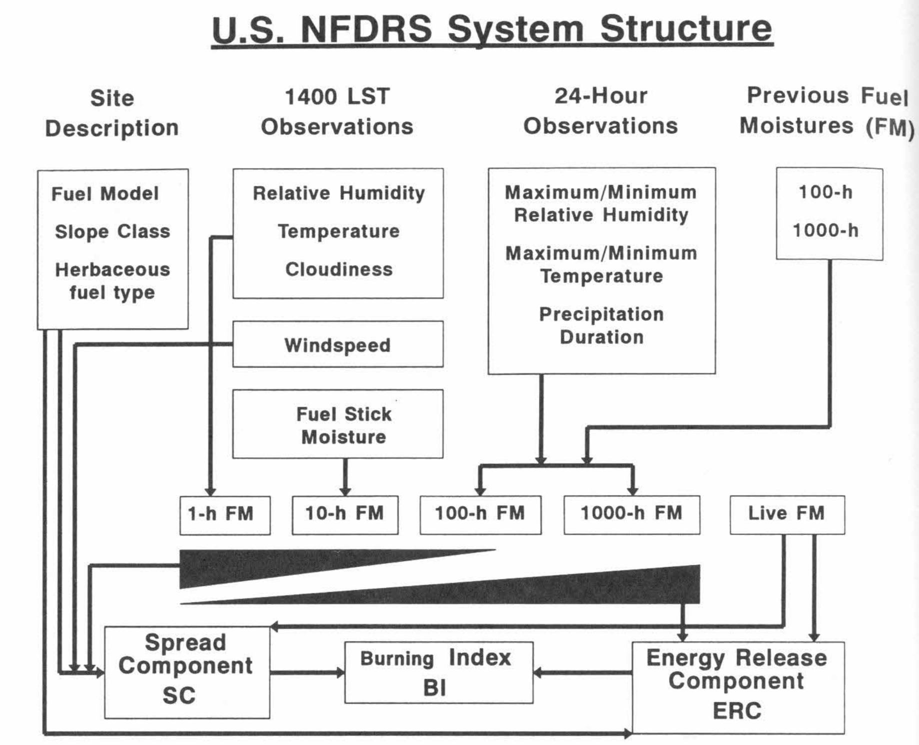


Figure 4.22. *Structure of the U.S. National Fire Danger Rating System (NFDRS) showing the relationship among site description, weather observations, intermediate fuel moisture calculations, and final indexes.* From Andrews and Bradshaw (1992).

(cloud cover and type of precipitation if any), and windspeed (10-min average). The weather observation also includes a number of elements for the 24-h period: maximum and minimum relative humidity and temperature, and precipitation duration and amount.

The moisture content of live fuel and four size classes of dead fuel are calculated from the weather observations and previous moisture values. Dead fuel is categorized by diameter or timelag as described in Chapter 3. The moisture content of the 10-h fuel is sometimes based on stick weight (see Figure 3.15). Calculation of live fuel moisture is based on the 1000-h moisture content. An option offered by the 1988 revision of NFDRS is entry of a greenness factor based on direct observation of state of live fuel.

All the indexes are affected by dead fuel moisture but to different extents. The wedge below the dead moisture boxes in Figure 4.22 indicates a mathematical weighting that emphasizes the large 100-h and 1000-h fuel for ERC, while smaller 1-h and 10-h fuels are emphasized for SC. (1000-h fuel has no influence on SC). Note that windspeed influences SC but not ERC. The relationship among SC, ERC, and BI is shown in Figure 4.23.

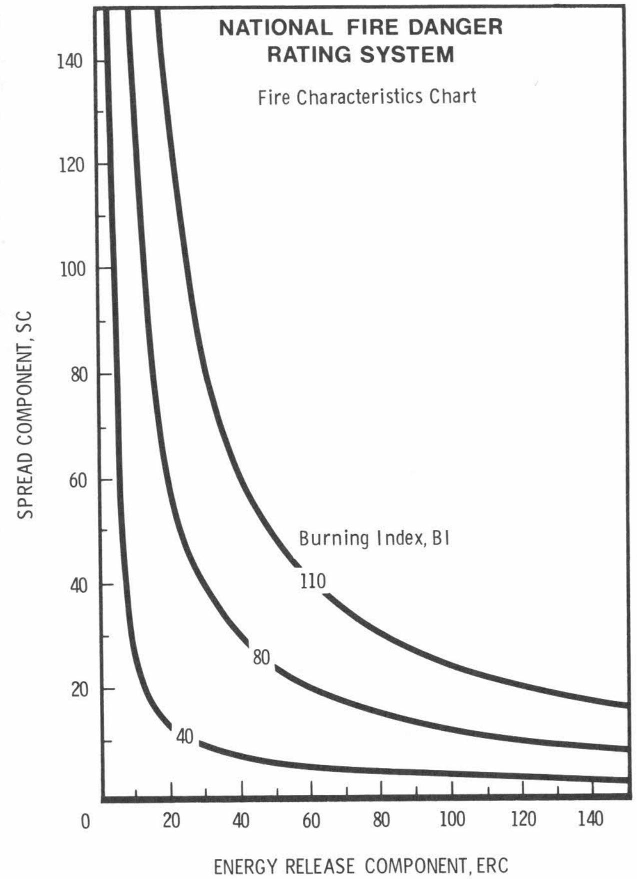


Figure 4.23. *Fire characteristics chart showing the relationship among Energy Release Component, Spread Component, and Burning Index.* From Andrews and Rothermel (1982).

Canadian Fire Weather Index (FWI) System The Fire Weather Index (FWI) System consists of six components (Figure 4.24). The three fuel moisture codes, the Fine Fuel Moisture Code (FFMC), the Duff Moisture Code (DMC), and the Drought Code (DC), are numerical ratings of the fuel moisture content of fuel moisture content of fine surface litter, loosely compacted duff of moderate depth, and deep compacted organic matter, respectively. The three fire danger indexes, the Initial Spread Index (ISI), the Buildup Index (BUI), and the Fire Weather Index (FWI) component itself, are intended to represent rate of fire spread, fuel available for combustion, and frontal fire intensity.

The FWI System components depend solely on daily measurements of dry-bulb temperature, relative humidity, a 10-m open wind speed, and 24-h cumulated precipitation recorded at noon local standard time. The three moisture

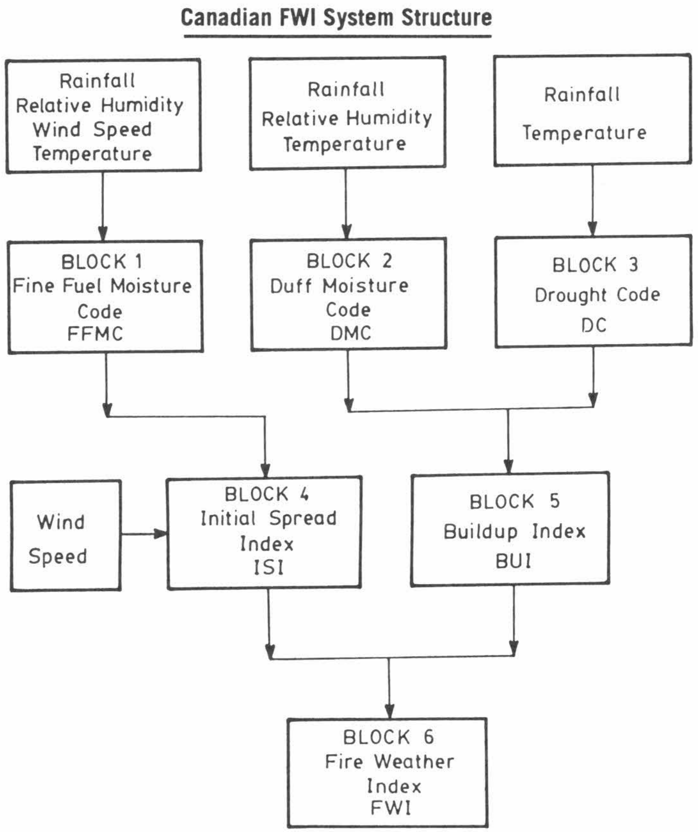


Figure 4.24. *Block diagram of the Canadian Fire Weather Index (FWI) System showing the relationship among weather observations, moisture calculations, and indexes.* From Van Wagner (1987).

codes are in fact bookkeeping systems that add moisture after rain and subtract some for each day's drying. The codes are expressed on scales related to actual fuel moisture.

The three fuel moisture codes plus wind are linked in pairs to form two intermediate indexes and one final index. The ISI, which combines the effects of wind and the fine fuel moisture content represented by the FFMC, represents a numerical rating of fire spread rate without the influence of variable fuel quantity. Because the ISI is dependent solely on weather, actual rate of spread (ROS) can be expected to vary from one fuel type to another over the range of the ISI because of differences in fuel complex characteristics and wind exposure. The BUI, which combines the DMC and DC, represents a numerical rating of the total fuel available for combustion. The BUI was constructed so that when the DMC is near zero the DC would not affect daily fire danger (except for smoldering potential) no matter what the level of DC. The

FWI, which combines the ISI and BUI, represents a relative measure of the potential intensity of a single spreading fire in a standard fuel complex (i.e., a mature pine stand) on level terrain. Jack pine and lodgepole pine forest types form a more or less continuous band across Canada, thus the concept of a standardized fuel type.

FWI System components and their values have different interpretations in different fuel types, because the System was developed to represent fire behavior in a generalized, standard fuel type.

Comparison of NFDRS and FWI From the above discussion of NFDRS and FWI, it is clear that there are many similarities. There are also significant differences. Both systems use daily weather observations or forecasts to calculate fuel moisture of several fuel elements and then combine them into indexes of fire danger, related to spread (SC and ISI), heat release or available fuel (ERC and BUI), and frontal fire intensity (BI and FWI). The relationship among SC, ERC, and BI is similar to that among ISI, BUI, and FWI (see Figure 4.23). Although there are differences in the weather elements required for each system, they are basically the same—temperature, humidity, wind, and precipitation. NFDRS requires more weather input than does FWI.

NFDRS attempts to evaluate the "worst" conditions on a rating area by measuring the weather for fire danger in the open on extreme (southwesterly or westerly) exposures. The FWI, in contrast, represents fire danger under a closed forest stand. NFDRS offers a choice of 20 fuel models that describe a range of grass, brush, forest, and logging debris. FWI represents conditions in a generalized pine forest, most nearly the jack pine and lodgepole pine type. NFDRS represents the moisture content of roundwood off the ground, without bark, in the open, whereas the FWI assesses the moisture of surface litter and duff under a canopy.

The type of model used to calculate moisture and fire danger indexes differ, to some extent because of differences in the training and approach taken by researchers in the two countries (engineers vs. foresters). The models in NFDRS are analytical, being based on the physics of moisture exchange, heat transfer, and other known aspects of the problem. Rothermel's fire spread model was described at the end of Chapter 1. It is the formulation of that model that allows a description of the fuel as an input. The Canadian models are primarily founded on a mass of field data of three kinds: weather, fuel moisture, and test-fire behavior, all collected over several decades. Canadian models are best applied to the fuel type in which the data were collected.

Drought Indexes

Palmer (1965) begins his discussion of drought by putting the term in perspective: "Drought means various things to various people, depending on their specific interest. To the farmer drought means a shortage of moisture in the root zone of his crops. To the hydrologist it suggests below average water levels in streams, lakes, reservoirs, and the like. To the economist it means a water shortage which adversely affects the established economy. Each has a concern which depends on the *effects* of a fairly prolonged weather anomaly." He suggested that drought be considered as a strictly meteorological phenomenon and evaluated as a meteorological anomaly characterized by a prolonged and abnormal moisture deficiency. But this approach wasn't entirely satisfactory. Because of misinterpretation of his drought index strictly as a measure of the current status of agricultural drought, he constructed a crop moisture index that takes into account only those moisture aspects that affect vegetation and field operations.

A drought period may be defined as an interval of time, generally of the order of months or years in duration, during which the actual moisture supply at a given place rather consistently falls short of the climatically expected or climatically appropriate moisture supply. Further, the severity of drought may be considered as being a function of both the duration and magnitude of the moisture deficiency. The term drought is reserved for dry periods that are relatively extensive in both time and space. A moisture shortage that is termed drought in one region may not be considered so elsewhere, and a shortage may be less serious in one season than it would be in another.

To a land manager concerned with wildland fire, drought may mean a situation in which fuels are drier and fires are more intense than normally expected. Under drought conditions, more fuel is available. Ground and crown fuels may burn, whereas under normal conditions, only the surface fuels would burn. A drought index for wildland fire should be chosen based on its reflection of the effects of the drought on wildland fuel. Different indexes represent moisture deficiency in different fuel elements—large, surface fuels (logs), litter and duff layers, soil profiles, and living foliage.

Drought can set the stage for severe wildfires. The supply of moisture to tree and plant roots decreases, and the loss of water by transpiration from leaves increases. Dried plant matter on the ground adds to the supply of combustibles. And the low moisture content of logs and duff makes those fuel components available to burn.

Long-term moisture deficiency in itself, however, cannot be used to forecast critical fire situations. If the smaller fuels are wet or green and winds are calm, serious fires usually will not occur at any time of year. Most critical fires are caused from a combination of factors that occur in conjunction with drought conditions.

Several indexes are available to track drought for wildland fire applications, including Palmer's drought index (1965) and crop moisture index (1968). Keetch and Byram (1968) developed a drought index (KBDI) specifically for wildland fire applications. The Energy Release Component and 1000-h moisture content from the U.S. NFDRS and the Drought Code and Duff Moisture Code from the Canadian FWI System are also used to indicate drought conditions.

Keetch–Byram Drought Index Keetch and Byram (1968) designed a drought index specifically for fire potential assessment. It is a number representing the net effect of evapotranspiration and precipitation in producing cumulative moisture deficiency in deep duff and upper soil layers. It is a continuous index, relating to the flammability of organic material in the ground.

The KBDI attempts to measure the amount of precipitation necessary to return the soil to full field capacity. It is a closed system ranging from 0 to 800 units and represents a moisture regime from 0 to 8 inches of water through the soil layer. At 8 inches of water, the KBDI assumes saturation. Zero is the point of no moisture deficiency and 800 is the maximum drought that is possible. At any point along the scale, the index number indicates the amount of net rainfall (in hundredths) that is required to reduce the index to zero, or saturation.

The KBDI is easy to compute and provides a continuous record because it is updated daily. It is now included as an NFDRS index. The inputs are weather station latitude, mean annual precipitation, maximum dry bulb temperature, and the last 24 h of rainfall. Reduction in drought occurs only when rainfall exceeds 0.20 inch (called net rainfall). The computational steps involve reducing the drought index by the net rain amount and increasing the drought index by a drought factor.

Figure 4.25 shows the ERC and KBDI for 1987 and 1988 for the Mammoth weather station in Yellowstone National Park. Those two fire seasons were quite different. In 1987 there were 34 fires, together covering 973 acres; only 1 fire was over 10 acres. The year 1988 set new standards for "worst case" with 60 fires covering 861,531 acres. The difference between the seasons is reflected by both the ERC and KBDI plots. The ERC dropped off in the fall of 1988 due to high nighttime humidity and precipitation less than 0.20 inch. This decrease in the ERC corresponded to a decrease in fire activity.

Canadian Drought Code The Drought Code (DC) component of the Canadian Fire Weather Index System is an indicator of the moisture content of deep organic layers, large downed wood, and the availability of water in small streams and swamps. As shown in Figure 4.24, rainfall and temperature are used to calculate the DC.

The DC shows definite seasonal trends influenced by the climate of a region. The seasonal trend in DC values precludes setting fire danger classes based on the DC. A single value early in the spring, while being very high for that time of the year, may only fall into a midrange of the whole season. Graphs produced by McAlpine (1990) can be used to determine how observed DC values compare with historical DC values for the station (Figure 4.26). Unusually dry (or wet) years show up as being above (or below) the one standard deviation line.

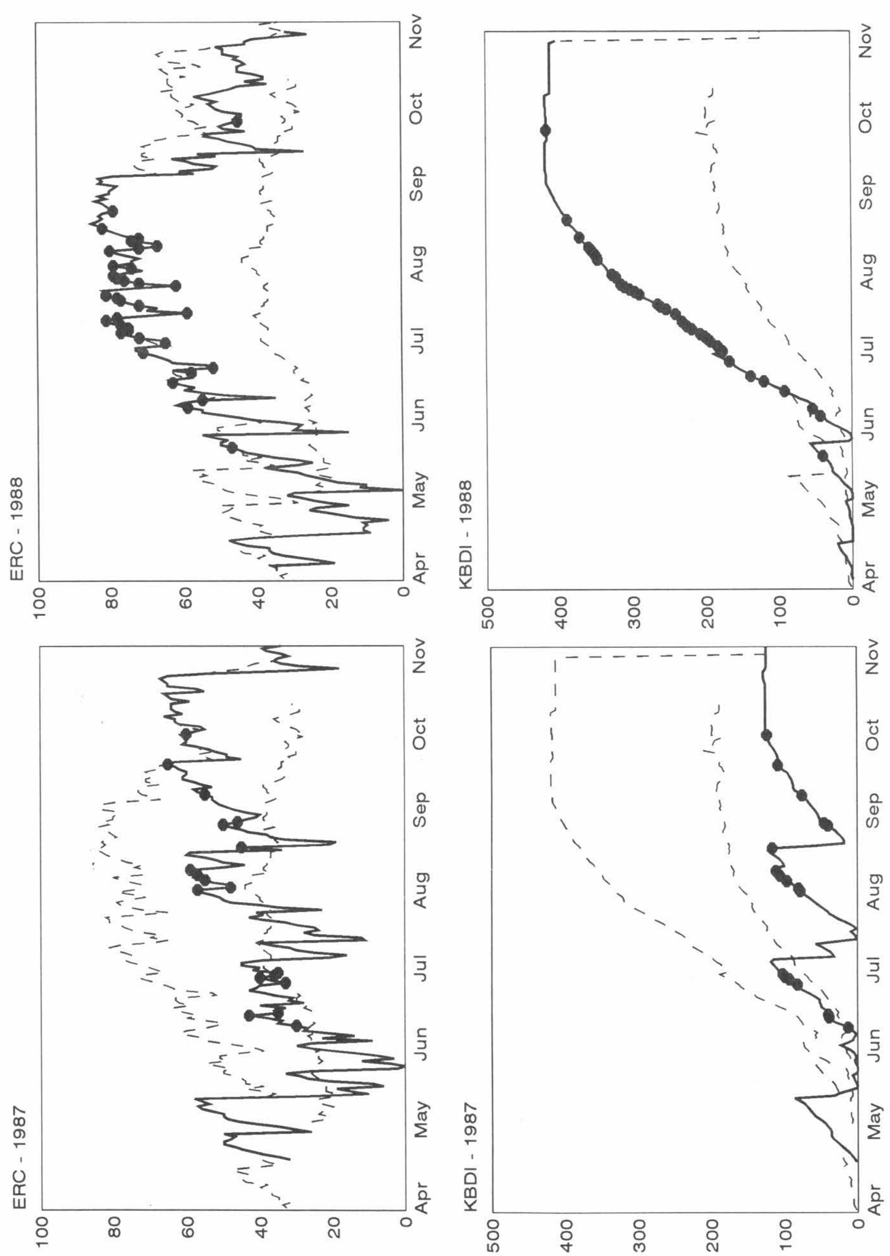


Figure 4.25. *Plots of Energy Release Component (ERC) and Keetch-Byram Drought Index (KBDI) for 1987 and 1988 in Yellowstone National Park (Mammoth weather station). Maximum and average for 1965–1992 are shown as dashed lines. Days on which fires were discovered are indicated with dots.* From Andrews and Bradshaw (1995).

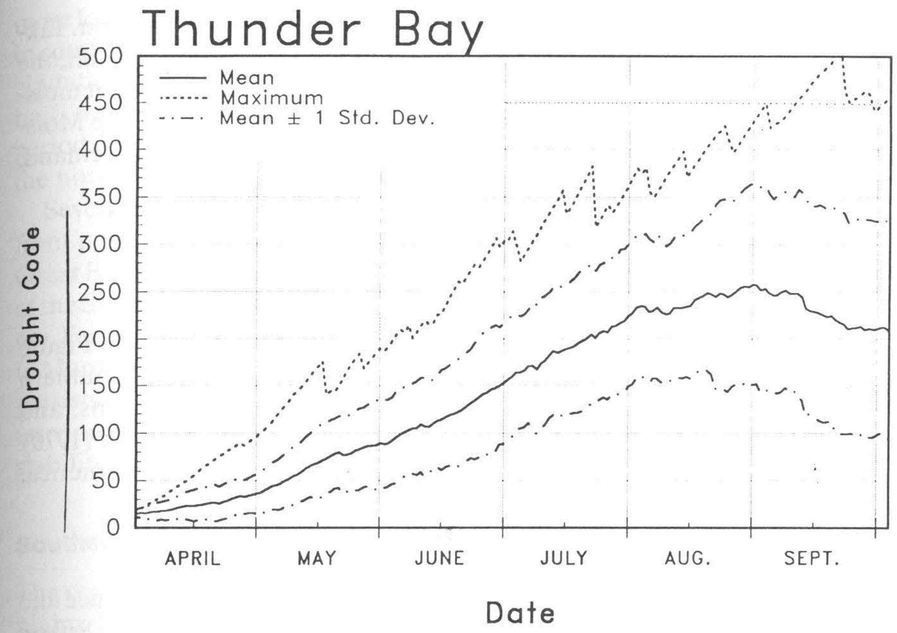


Figure 4.26. *Canadian Drought Code (DC) maximum, average, and ±1 standard deviation for Thunder Bay based on 35 years of weather data.* From McAlpine (1990).

Palmer Drought Index A sophisticated index of meteorological drought was developed in the 1950s by Wayne Palmer (1965) It is based on the concept of supply and demand: Supply is represented by precipitation and stored soil moisture; demand is figured by a formula that combines moisture loss through evapotranspiration (evaporation from land and water surfaces and transpiration from vegetation) with both the amount of moisture needed to recharge the soil moisture and the amount of runoff required to keep rivers, lakes, and reservoirs at normal levels.

The results of this water-balance accounting produce either a positive or a negative figure, which is weighted by a climatic factor. The final figure is an index that expresses the degree of abnormality over a period of several months in a particular place. An index of +2 to −2 indicates normal conditions; −2 to −3 indicates a moderate drought, −3 to −4 a severe drought, and below −4 an extreme drought.

The Palmer index is normalized. Similar values should mean similar drought level at any location. It is a slow response system. Given a wet spring followed by a dry summer, the index can still indicate a moderate drought by late summer.

Crop Moisture Index Palmer's Crop Moisture Index (1968) is a more short-term index that measures the degree to which growing crops have received adequate moisture during the preceding week. The Crop Moisture Index is computed from average weekly values of temperature and precipitation. Taking into account the previous moisture conditions and current rainfall, the index determines the actual moisture loss and hence the crop's current moisture demand. If moisture demand exceeds available supplies, the Crop Moisture Index has a negative value; if current moisture meets or exceeds demand, the value is positive.

4.5 SELECTED EXAMPLES

Following is a description of three of the 15 climate regions shown in Figure 4.20: Interior Alaska and the Yukon (1), South Pacific Coast (3), and Southern States (14). Vegetation, meteorological conditions, weather patterns, and typical fire season are described according to Schroeder and Buck (1970). These examples illustrate the wide range of fire climates in North America.

Interior Alaska and the Yukon

The vegetation in Interior Alaska and the Yukon is predominantly spruce and aspen, with some tundra and other lesser vegetation to the north. The Yukon Basin has a warm, short summer. Continental heating has produced summertime temperatures of 100°F; however, temperatures as low as 29°F have occurred in July. Winters are extremely cold. The high coastal mountains generally prevent the invasion of maritime polar air masses at low levels. The Brooks and other mountain ranges block the inflow of even colder continental polar air from the north.

Annual precipitation is only about 10 to 15 inches, the maximum occurring during the summer in convective showers and with weak fronts. Precipitation is highest in the southern portion, which includes the northern extension of the Cordilleran Highlands and their parallel chains of lesser mountains. Although precipitation is maximum in summer, it is so scant that wildland fuels dry out considerably during the long, clear, dry summer days. Dry thunderstorms are not infrequent. The usual fire season starts in May after melting of the winter snows and lasts until September.

South Pacific Coast

The vegetation in the south Pacific coast region consists of grass in the lowlands, brush at intermediate levels, and extensive coniferous stands in the higher mountains. Temperatures along the immediate coast are moderated both winter and summer by the ocean influence. But only short distances inland, winter temperatures are somewhat lower and summer temperatures average considerably higher.

The annual precipitation is generally light, around 10 to 20 inches at lower elevations. Precipitation in the mountain areas reaches up to 60 inches or

more locally. Summers are usually rainless, with persistent droughts common in southernmost sections. Widespread summer thunderstorms, with little precipitation reaching the ground, particularly in the mountains of the northern half, occasionally result in several hundred local fires within a 2- or 3-day period. The fire season usually starts in June and lasts through September in the north, but in the south critical fire weather can occur year round.

Several synoptic weather types produce high fire danger. One is the cold-front passage followed by winds from the northeast quadrant. Another is a Great Basin High that produces foehn-type Mono winds along the west slopes of the Sierras and Coast ranges, and the Santa Ana winds of southern California. Peak Santa Ana occurrence is in November, and there is a secondary peak in March. A third high fire danger type occurs when a ridge or closed high aloft persists over the western portion of the United States. At the surface, this pattern produces very high temperatures, low humidities, and air-mass instability.

Southern States

The vegetation in the southern states consists mainly of pines along the coastal plains, hardwoods in bays and bottomlands along stream courses, and mixed conifers and hardwoods in the uplands. Flash fuels, flammable even very shortly after rain, predominate in this region. The topography along the Gulf and Atlantic is low and flat. Inland from the Atlantic Coast it merges with an intermediate Piedmont area. The southern Appalachians are included in this region, and the central portion includes the lower Mississippi Valley.

Summers are warm and generally humid, because the region is almost continuously under the influence of a maritime tropical air mass. Winters have fluctuating temperatures. When maritime tropical air moves over the region, high temperatures prevail. Following the passage of a cold front, continental polar air may bring very cold temperatures—well below freezing—throughout the southern states.

Annual precipitation varies from 40 to 60 inches over most of the region, except for about 70 inches in the southern Appalachians and over 60 inches in the Mississippi Delta area, and falls mostly as rain. The influence of the moist maritime tropical air from the Gulf of Mexico causes abundant rainfall in all seasons, with slightly higher amounts in August and September due to the presence of hurricanes in some years. Spring and fall have less precipitation than summer or winter, with spring being wetter than fall. Winter precipitation is usually associated with frontal lifting or with lows that develop over the southern states or the Gulf of Mexico and move through the region. Summertime precipitation is mostly in the form of showers and thunderstorms. During the colder months, much fog and low stratus are formed by the cooling of the maritime tropical air as it moves northward.

The fire season in the Southern States is mainly spring and fall, although fires may occur during any month. Lightning accounts for only a minor number of fires. Very often the most critical fire danger weather occurs with the passage of a dry cold front. The strong, gusty, shifting winds with the cold front and dry unstable air to the rear set the stage for extreme fire behavior.

FURTHER READING

Schroeder and Buck (1970) prepared a thorough, well illustrated presentation of *Fire Weather*. That publication and weather lessons in the U.S. National Wildfire Coordinating Group (NWCG) fire behavior courses S-290 and S-490 (1994, 1993), prepared in large part by the fire weather meteorologists at the National Interagency Fire Center, were used extensively in writing this chapter. Baughman (1981a) compiled "An Annotated Bibliography of Wind Velocity Literature Relating to Fire Behavior Studies."

The U.S. National Fire Danger Rating System is described by Deeming, Burgan, and Cohen (1978) in "The National Fire Danger-Rating System—1978," by Bradshaw and others (1983) in "The 1978 National Fire-Danger Rating System: Technical Documentation," and by Burgan (1988) in "1988 Revisions to the 1978 National Fire-Danger Rating System." The Canadian system is described by Van Wagner (1987) in "Development and Structure of the Canadian Forest Fire Weather Index System" and by Stocks and others (1989) in "The Canadian Forest Fire Danger Rating System: An Overview."

Part Two

Fire Regime