5

CLOUDS AND PRECIPITATION

Chapter Highlights

- ✓ Learn the difference between fair-weather and storm clouds.
- ✓ Discover how rain and snow form in the atmosphere.
- ✓ Overview of all the major cloud types that are found in the mountains.

Clouds: An Overview

Water in the earth's atmosphere can of course exist in three different states: vapor, liquid, and solid; with water vapor being the most common. Water vapor is H_2O in a gaseous state, which is by the way-invisible to the human eye. Water gets into the atmosphere primarily via evaporation which occurs over the world's oceans and lakes. Secondary sources include evaporation from wet soils, and ponded surface water, as well as from the transpiration of plants. Once in the atmosphere, water vapor can be transported thousands of kilometers before it changes to a liquid state (condensation) and forms a cloud. The process of evaporation–water vapor–cloud--precipitation–runoff: is referred to as the hydrologic cycle. Even though the term atmospheric water is frequently used, recall from Chapter 2 that virtually all of the water in the atmosphere is located in the lowest 10 km (6 miles) of the atmosphere.

There is considerably less water in the atmosphere at any given time than you might imagine. For example, if all of the water vapor in the atmosphere was suddenly transformed into a liquid, on average it would comprise a layer of water roughly 2.5 cm (1 in) thick. The thickest layer would be in the tropics where evaporation is large, the thinnest layer would be in the polar regions, where evaporation is minimal. Evaporation occurs in all of the earth's climatic zones, however the rate at which it occurs is primarily a function of air temperature.

Meteorologist use a number of terms to specify how much water vapor is in the air, the most frequently used ones are: specific humidity, dew point, and relative humidity (RH). When the RH of a parcel of air reaches 100%, the air is full of water vapor, a condition which you already know as *saturation*. Using RH can be a bit tricky since it is temperature dependent. For example, warm air can hold more moisture than cooler air, therefore two parcels of air with a different temperature but equivalent RH's, actually contain different amounts of moisture!

A cloud forms when water vapor condenses into a liquid or a solid, the actual process will be discussed in the following sections, however at this point in our discussion it is sufficient to say that when cloud droplets or ice crystals do form, they are very small. Cloud droplets for example typically have diameters on the order of 0.01 mm (0.0004 in), so small in fact that they are held in suspension in the cloud. Most clouds never reach a mature enough state to form precipitation, their droplets simply re-evaporate back into water vapor. In the simplest of terms, a cloud can be defined as a collection of billions of liquid droplets (or ice crystals) that undergo a cycle of condensation – evaporation – condensation. Not all clouds of course consist of water droplets, cirrus clouds which form in the upper troposphere, are made-up of very small ice crystals. Some clouds like towering cumulus, consists of water droplets in the lower half and ice crystals in the upper-half.

There are a number of ways in which clouds can be classified. The most widely used classification divides them into two groups based on their structure: stratiform (layered) and cumuliform (billow), as illustrated in Figure 5.1 and listed in Table 5.1. Stratiform clouds typically have lifetimes on the order of hours to days, while cumulus clouds have lifetimes on the order of minutes to hours. Fog is not a separate cloud type, it is actually a type of stratiform cloud which happens to be in contact with the ground. In Table 5.1 notice how 'stratus' is also used in compound with other cloud types, for example, cirrostratus (high altitude ice clouds) or stratocumulus (layer of cumulus).

Many clouds are carried along by the geostrophic wind, the exceptions are mountain wave and lee-wave clouds, which are generated at site-specific locations above high mountains. Most stratiform clouds travel hundreds or even thousands of kilometers from the point at which they first formed. At the other end of the scale, fair weather cumulus clouds form and dissipate locally, frequently in less than one hour. There are times however, when large aggregates of cumulus clouds merge, forming cloud complexes that may last for several days, and cover an area on the order of several hundred kilometers.

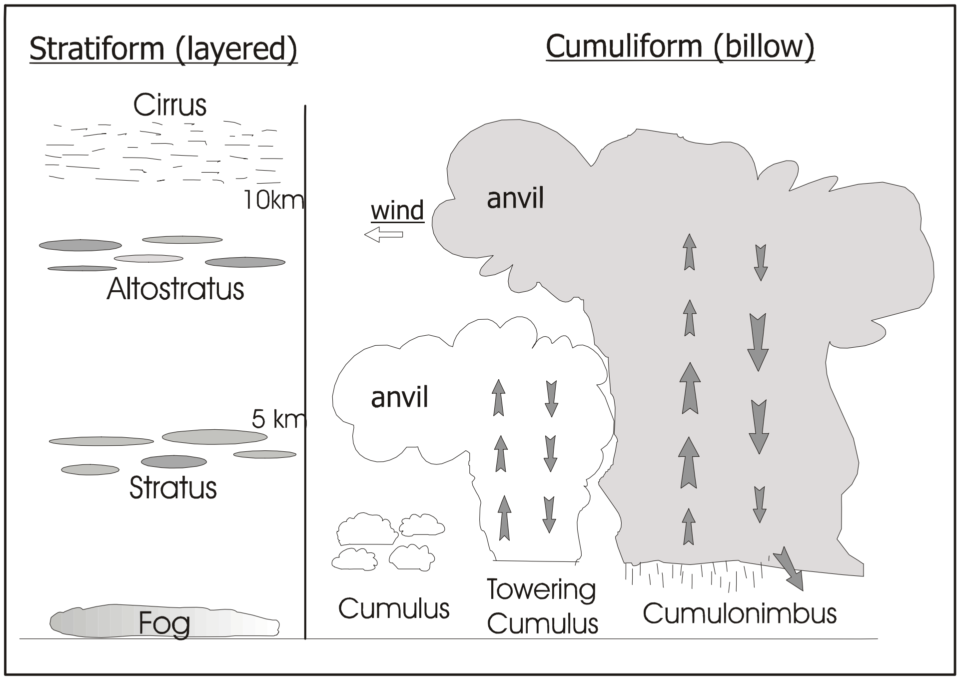


Figure 5.1- Basic cloud types.

Cloud Formation

If you really want to understand the life-cycle of a mountain storm, then you need to know something about how and why clouds form. Memorizing all of the Latin cloud names is not the goal, beneficial knowledge comes from knowing how the cloud formed and what it means in regards to future weather. Frankly, this is difficult to do much of the time. While a snap shot view of clouds is of some value, what is really important in weather analysis, is the evolution or <u>sequence of cloud formation over a period of time.</u>

Table 5.1 Cloud Classification

Name	**Latin Meaning**	**Abbreviation**	**Altitude**	**Structure**
Stratiform				
Cirrus	*curly*	*Ci*	high	wispy/filament
Cirrostratus	*little cirrus*	*Cs*	high	layered
Altostratus	*high stratus*	*As*	med-high	layered
Nimbostratus	*rain stratus*	*Ns*	low-high	layered but tall
	layered	*St*		

Stratus	------	*Fg*	low-med	layered
Fog			low	layered
Cumuliform	*humble cumulus*	*Cu*		
Cumulus Humilis	*grouped cumulus*	*Cg*	low-med	Separate clouds
Cumulus	------	*Tcu*	low-med	Aggregated *Cu*
Congestus	*rain cumulus*	*Cb*	low-high	Tall *Cg*
Towering			low-high	Aggregated *Tcu*
Cumulus				
Cumulonimbus				

Before we start discussing specific cloud types, it is instructive to consider what happens when a parcel of air is lifted, lets say over a mountain so that it becomes saturated. Figure 5.2 shows a temperature profile of a parcel of air as it is lifted from the surface starting at Pt **.A**. For this example we will assume for simplicity that the environmental lapse rate is -9.8° C km^{-1}. Between Pts. **A** and **B**, the parcel cools at the dry adiabatic lapse rate (-9.8° C km^{-1}), which in this example is the same as the environmental lapse rate. At Pt. **B** the parcel becomes saturated and condensation commences. This level is know as the Lifting Condensation Level (LCL), and varies in height according to the initial temperature and moisture content of the parcel. Once water droplets or ice crystals form, *latent heat* (energy released/required in phase changes) is released into the parcel, increasing its temperature. Further cooling due to lifting occurs at the moist adiabatic lapse rate which varies between 4° C km^{-1} in the lower troposphere to 7° C km^{-1} in the middle and upper troposphere. Typically a value of 6.5° C km^{-1} is used for most calculations. If the lifting continues past the LCL, it is possible that the cloud reaches a elevation where it becomes positively buoyant (Pt. **C**), this height is known as the Level of Free Convection (LFC). Once the top of a cloud passes the LFC, it continues to rise until it either runs out of moisture or when it encounters a stable layer of warm air. The height at which the parcel is no longer positively buoyant is called the equilibrium height (Pt. **D**). The height of the LFC is a function of the environmental lapse rate. Cool or cold air in the mid-troposphere usually has a high environmental lapse rate (>7° C km^{-1}), which is very conducive to cumulus development. The closer the LFC is to the LCL, the easier it is for cumulus clouds to develop. In the simplest of terms, clouds that are contained between the LCL and LFC are stratiform while those that continue to grow above the LFC are cumuliform.

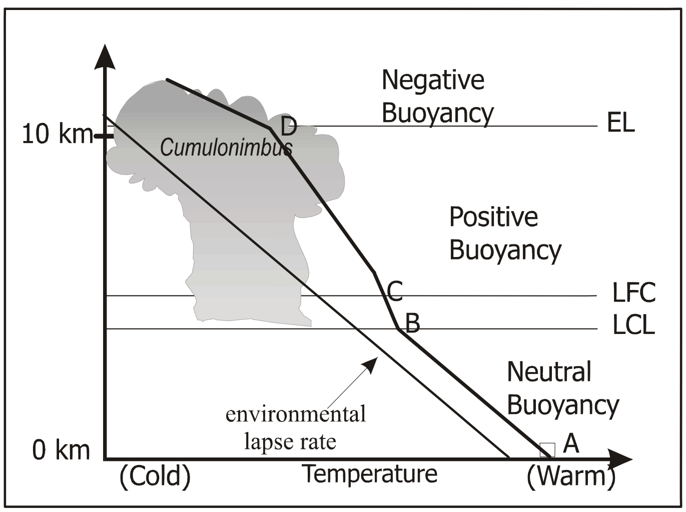


Figure 5.2- Temperature profile for idealized cloud.

Stratiform Clouds

Stratiform clouds often extend over thousands of square kilometers of the earth's surface, but are typically only two to four kilometers thick, which of course gives these clouds their layered appearance. These are the type of clouds that are most associated with low pressure systems and frontal lifting. There are specific sectors around a low as illustrated in Figure 5.3, that contain predominately stratiform or cumliform clouds. You should note that stratus layers frequently contain considerable amounts of embedded cumulus clouds as well.

Stratiform clouds form when a layer of air becomes saturated, either by the addition of moisture or by adiabatic cooling due to lifting. Layers of air are lifted when moving up a frontal boundary or when forced to flow over a mountain (Figure 5.4). Note that the slope of the frontal boundary is greatly exaggerated in Figure 5.4a. Most frontal boundaries are nearly horizontal, yet they do provide enough lift to produce massive areas of stratiform clouds.

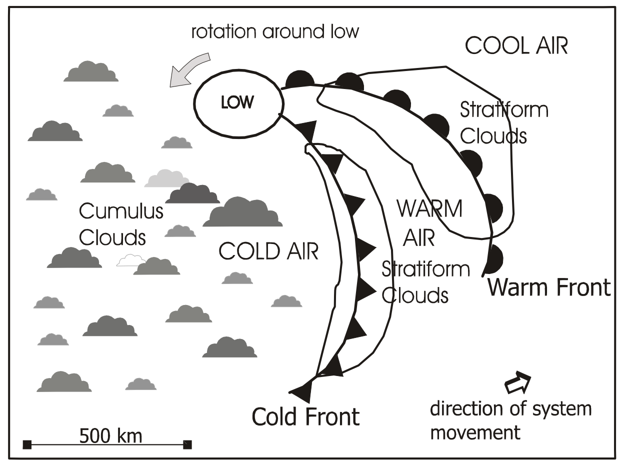


Figure 5.3- Plain-view of low pressure system with its fronts and associated clouds.

Cirrus clouds are made-up entirely of small ice crystals. Cirrus take on the form of long slender streaks, or thin semi-transparent sheets (cirrostratus). Since these clouds are located in the upper troposphere, their structure is often distorted by jet stream winds. Cirrus clouds that have a sharp linear boundary, indicate that the jet stream is roughly parallel to the edge of the cloud boundary. It is common for certain cirrus streaks to take on a hooked shape. When this occurs, relatively large ice crystals are falling from one cirrus layer into a layer below where the wind direction is different from the layer above. This results in the cirrus streak having a hooked shape when view from below. Cirrostratus are responsible for the halo or arcs that form around the sun at angles of 22^o (halos also form in ice fogs). Cirrus are the lead clouds in approaching fronts, however, the presence of cirrus in of themselves does not mean that a front is approaching, remember that it is the sequence of clouds that is important.

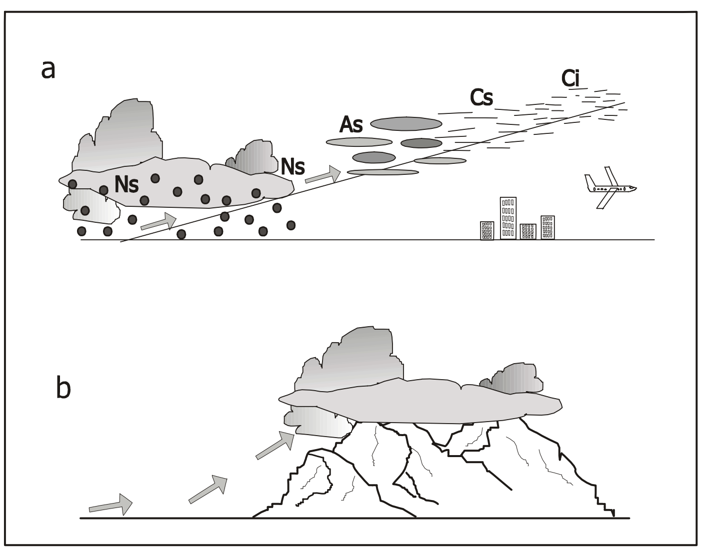


Figure 5.4- Two ways in which air is lifted to create clouds: (a) frontal (b) orographic.

Stratiform clouds which form in the middle troposphere are called altostratus while those that form in the lower troposphere are known as stratus. These clouds consist of either water droplets or ice crystals. These are the clouds that are most frequently responsible for overcast skies and light precipitation. It is common to find stratus clouds developing in distinct layers, with dry air in-between each layer. In addition, stratus play an important role in the radiation and temperature regime of the underlying surface. During the day, stratus obscure the sun, limiting the amount of incoming shortwave radiation that reaches the ground. The result is a cooler daily maximum temperature than would normally be recorded. At night, stratus clouds trap outgoing longwave radiation emanating from the surface, increasing the nighttime minimum temperature. The net result is a diurnal temperature range which is well below what would occur under clear sky conditions.

Fog is essentially a stratus cloud that is in contact with the ground. There are several different kinds of fog; advection and radiation being the most common. Advection fog forms in one location but then moves to another area. For example, fog may form over a lake at night, and later in the morning move over land as surface winds blow it away from the lake. Radiation fog forms *in situ* due to the cooling of air until it becomes saturated. In general, radiation fogs and even advection fogs form only when wind speeds are light (<5 m s^{-1} or 10 mph). Valley bottoms are preferred locations for the formation of radiation fog in mountainous environments, especially after a rainstorm when the ground is wet and the longwave radiation cooling at night is extensive. On occasion a thin fog will not form until after sunrise because heat from the sun is needed to produce evaporation, which in turn adds moisture to the air, which then becomes saturated.

Cumulus Clouds

Cumulus clouds range from small fair-weather cumulus (*Cu*) , whose tops just reach the level of free convection (LFC), to giant cumulonimbus clouds (*Cb*) that extend well into the lower stratosphere (thunderstorms). In order for a large cumulus cloud to develop, an ascending parcel has to reach the LFC. It does this in two different ways: via thermal heating or by forced lifting.

Thermal heating of the earth's surface and the adjacent layer of air above the surface, is the mechanism most often associated with convection and cumulus development. Dry bare ground tends to heat up much faster than ground that is either moist or vegetated. Heat from the ground warms the near surface air which becomes positively buoyant, these ascending parcels are called thermals or updrafts. If an updraft is weak, the parcel has little chance of reaching the LFC, the result is a short lived (on order of 5-15 minutes) cumulus cloud. On days with vigorous convection, individual cumulus will often aggregate into groups forming cumulus congestus (*Cg*). Another commonly observed feature of developing *Cu* and *Cg* is a cloud base that has a uniform height above the ground, indicating a consistent LCL over the region.

Towering cumulus (*Tcu*) are vertically well developed clouds that have strong updrafts and downdrafts. Towering cumulus typically have lifetimes on the order of 30-45 minutes, and are capable of producing short periods of heavy rain and at times hail. If conditions allow, *Tcu* often form clusters that are capable of taking on a new form and develop into a large cumulonimbus *(Cb)*. A typical *Cb* has a lifetime on the order of an hour, and can produce large amounts of rain or hail. They are also capable of generating thunder and lightning, hence these clouds are known as thunderheads to the general public. From time to time a *Cb* will remain stationary over hilly or mountainous terrain, dumping large amounts of rain (>5 cm) in a 15 to 30 minute period. The combination of heavy rain over mountainous terrain is what produces flash floods. Anvil clouds are the result of moderate to strong upper level winds blowing cloud material downstream from the tops of mature *Tcu* and *Cb*'s, and are therefore good indicators of upper level wind direction.

Some readers may have noticed that the lifetimes given for the various types of convective clouds, may appear to be considerably shorter than with what they have observed. This apparent discrepancy can be explained in the following way; cumulus clouds are in a continual state of growth and dissipation, new clouds grow adjacent to older clouds. To the casual observer the cloud appears to have a lifetime on the order of hours, when in reality a sequence of clouds with much shorter lifetimes have formed and dissipated. Another important point is that developing cumulus tend to be relatively fixed with respect to the ground below, indicating their dependancy on surface forcing. Large *Tcu* and *Cb* are less dependent on the surface forcing and as a result they usually move along in the direction of the mid-level wind.

Cumulus development is not limited to thermals that are generated at the earth's surface. From time-to-time layers within the lower and middle tropopause can be very conducive to cumulus cloud formation; this occurs when a layer of cold dry air moves over a layer of warm moist air. Even modest lifting of the warm air can trigger convection within the colder air. Since cumulus clouds form in all climatic zones, over all types of terrain, and in all air-mass types, what role do mountains play in their development? Mountains are important source regions for cumulus development for a number of reasons: 1) They are elevated sources of heat and moisture; 2) They allow cold air to ride over warm air, creating unstable conditions above the mountains, and; 3) They generate forced lifting as air moves over the mountain. It is important to note that even though mountains are source regions of cumulus clouds, the mountains themselves generally do not receive the brunt of severe weather (heavy rain, hail, strong winds, and tornadoes). Most severe weather occurs in the lowlands, downstream of the mountains.

Mountain Stratiform Clouds

Stratiform clouds that are found in mountainous environments come in a variety of forms, three of which will be discussed in this section. Banner clouds and mountain wave clouds which are discussed below, are not your typical stratus-type clouds, nevertheless they can be quite common in certain mountain ranges. They are considered a type of stratus cloud not because they look like the typical stratus, but rather because they occur when the troposphere has a stable lapse rate.

Mountain Stratus

When discussing mountain stratus we should differentiate between stratus which form on the synoptic-scale (frontal lifting) and subsequently move into the mountains, versus stratus that are generated locally in the mountains. In the first scenario, mountains act as a barrier to the moving cloud mass. As a result windward slopes are frequently cloudy while the lee slopes are predominately cloud free. This type of stratus can last for days, depending on the synoptic weather pattern and the moisture supply. They can be composed of a single cloud mass or several layers, with cloud free layers in-between (Figure 5.5a). It should come as no surprise that coastal mountain ranges have the highest frequency of stratus development. In this case marine stratus form over the cool oceanic waters that are found off the west coast of North America. These clouds in turn are transported towards the coastal mountains by westerly winds.

During the warmer months of the year when low-level moisture is plentiful, it is common for a single layer of stratus to move into a mountain range. The top of the stratus layer often corresponds with the base of an elevated inversion (Fig. 5.5b). This might occur for example in conjunction with a warm front or occlusion. In this scenario it is very common for stratus to be confined to the lowest one or two kilometers of the troposphere, so that the higher summits are cloud free.

Locally produced mountain stratus form when moist air is forced to move upslope or up valley (Fig. 5.5c). Stratus often linger in alpine valleys more than any other location because the valley atmosphere does not mix with drier ambient air as proficiently as air over the adjacent valley slopes. After a precipitation event when there are copious amounts of moisture in a valley, stratus often form at night in response to the radiative cooling of the valley atmosphere (Figure 5.6). During the day upslope and valley winds often help dissipate stratus via the mixing of drier air.

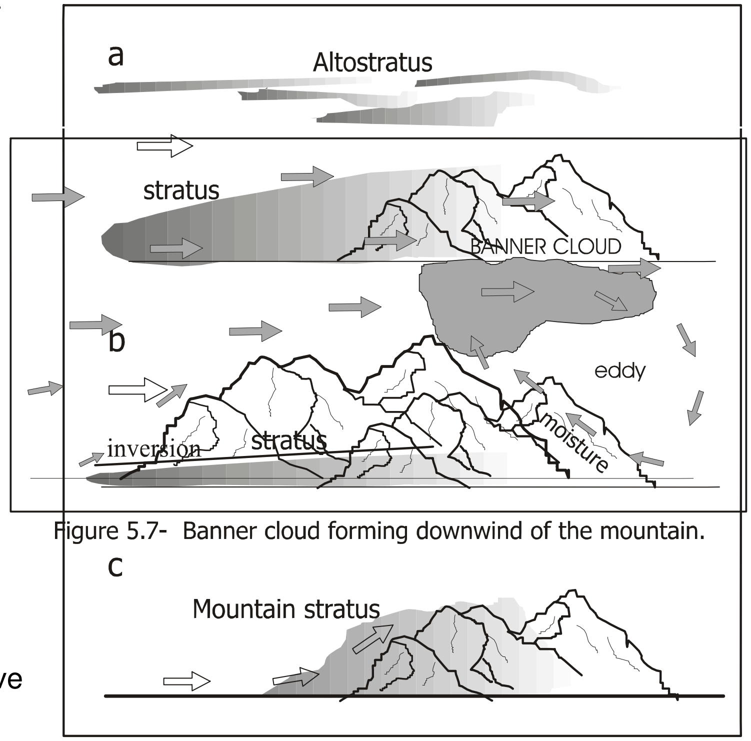


Figure 5.7- Banner cloud forming downwind of the mountain.

Figure 5.5- Three stratus cloud configurations.

In our discussion to this point we have portrayed stratiform and cumliform cloud formation as being virtually independent. In reality there is considerable interaction between lifting of layers and convection. For example, as air approaches a large mountain barrier, mid-tropospheric levels start to rise before lower levels, which can lead to the development of cumuliform clouds within a otherwise stratiform weather system. Smith (1982) pointed out the importance of mountains retarding the movement of an advancing cold front. As cold air approaches the mountain barrier, low-level air becomes blocked in front of and over the windward slopes. Above the mountain, cold air continues to move downstream. The result is a situation you should recognize as being convectively unstable, because if any of the warm air parcels are lifted into the cold air, they will become positively buoyant.

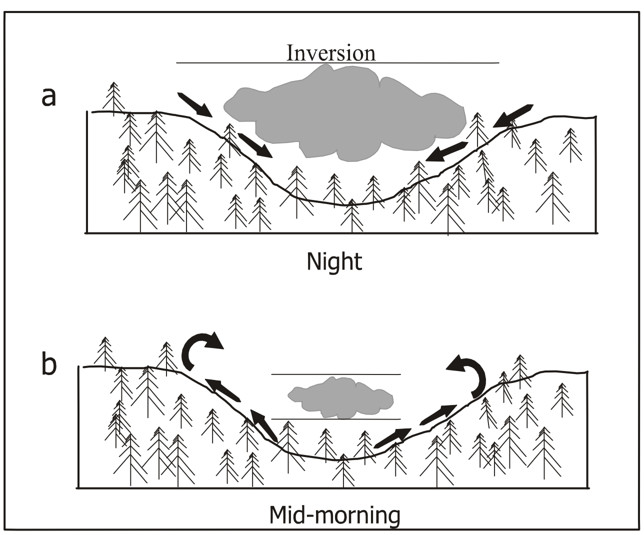


Figure 5.6- (a) Stratus cloud forms below inversion during the night. (b) Cloud dissipates as dry air from above inversion mixes moist valley air.

<u>Banner Clouds</u>

Banner clouds form directly in the lee of a prominent summit or ridge. It is easy to mistake a snow plume for a banner cloud, although at times they may form in conjunction with each other. Figure 5.7. illustrates the fundamentals of banner cloud formation as presented by Douglas (1928) in his study of clouds around the Matterhorn. The most important requirement is strong summit level winds, which as they blow over the top of a mountain creates a vertical eddy over the lee-ward slope. Upslope flow is produced by this eddy and if sufficient moisture is available, a cloud is produced in the ascending air. Due to the presence of strong summit level winds, the lee-side cloud becomes elongated in the shape of a banner or plume as it is blown downstream of the mountain. Banner clouds form quite frequently over some of the higher Himalaya peaks as a result of strong sub-tropical jet stream winds.

<u>Wave Clouds</u>

Virtually every mountain traveler has at some point in their travels seen a variety of wave clouds**,** which form in the crest of stationary mountain waves. If the cloud is near or in contact with the summit it is referred to as a *cloud cap*, when it is above the summit it is a *mountain wave cloud*. When wave clouds form some distance downstream of a mountain they are called *lee-wave clouds*. Mountain wave clouds form when the air at summit level and above is stable, and the wind speeds are moderate to strong (Figure 5.8). They may consist of a single or multi-layered clouds, and contain either water droplets or ice crystals. As seen from a distance, the edges and surface of wave clouds appear to be smooth. If you are climbing towards a cloud cap, expect poor visibility and a significant increase in winds once inside of the cloud. Note however that the surface winds lower down the mountain, well below the cloud cap are generally not very strong. Politovich & Vali (1983) summarized a decades worth of cloud cap observations made from the University of Wyoming's, Elk Mountain Observatory, located at 3307 m (10,800 ft) in the Medicine Bow Range. The typical Elk Mountain cloud cap ranged in thickness from 200 -2000 m (660-6600 ft), while the leading edge was located some 2-7 km (1-4 mi) upstream of the summit. The residence time for air parcels (the time it takes air to move from the upstream edge through the cloud to the downstream edge), ranged from 500-1500 seconds. Typical summit level winds during cloud cap events was from 10-20 m/s (22-44 mph), with 10 m/s (22 mph) considered the lower threshold for their development.

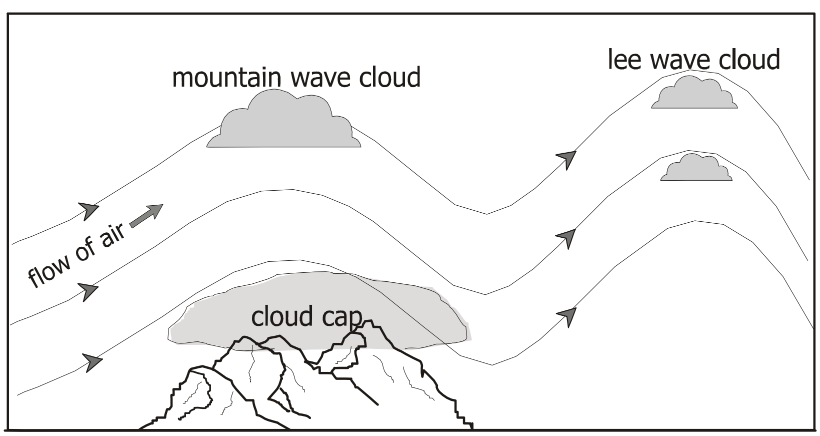


Figure 5.8- Three types of mountain wave clouds that fall under the generic category of lenticular clouds.

Multi-level mountain wave clouds form because the moisture in the air above the mountain is located in distinct layers. Theoretically it is possible to have one very thick mountain wave cloud if the air above the mountain contains significant moisture, however it is fairly rare to have a massive mountain wave cloud and no cloud formation at lower levels.

Cloud caps are best viewed over isolated summits where flow around the sides of the mountain help create a very well defined cloud outline. The higher peaks of the Cascade Range frequently have cloud caps or mountain wave clouds due to abundant moisture and stable lapse rates in the middle troposphere. It has been our observation that in the Cascades during the summer, there are times when cloud cap formation occurs on a diurnal cycle over the larger peaks. Around Mt. Rainier in particular, low-level cumulus convection occurs in the morning and early afternoon. At the same time, because of a stable lapse rate, mountain waves form over the summit. Cloud cap formation in the mountain wave is delayed until afternoon however, due to a lack of moisture during the morning hours. By afternoon, as cumulus clouds grow in size and moisture is 'pumped' into the mid-troposphere, a cloud cap finally forms. The cloud retains its structure until sometime during the night when the supply of moisture is shut-off. We have observed this phenomena on a number of successive days when the Pacific Northwest was under the influence of high pressure, and the sky was otherwise cloud free.

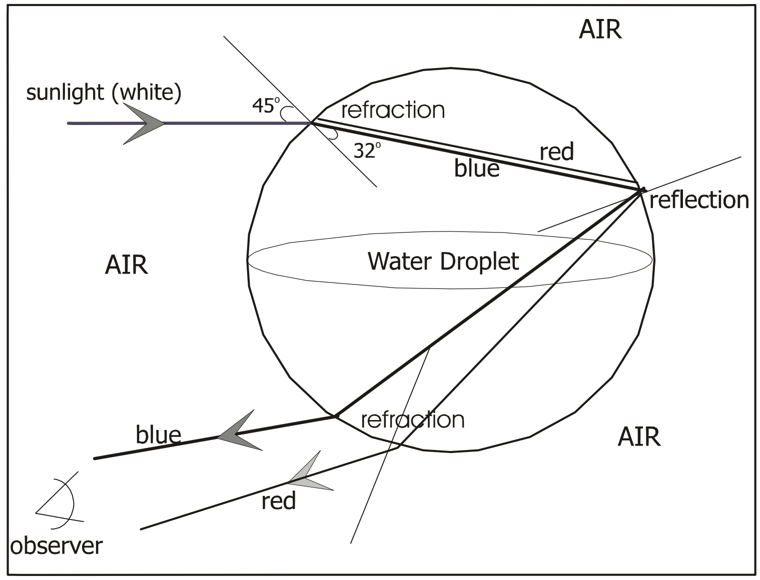
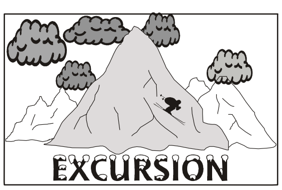


Figure 5.9-- Separation of colors as sunlight travels through a water droplet.

CLOUD COLORS AND RAINBOWS

Cloud droplets and ice crystals <u>scatter</u> sunlight-which means they absorb the original sunlight then re-emit it, at select wavelengths. Cloud droplets and ice crystals are quite large, relative to gas molecules, so they have the tendency of re-emitting the same colors as they absorb. Clouds appear white the majority of the time because cloud droplets and ice crystals scatter sunlight of all wavelengths equally well. When clouds do change colors it is generally because of ambient lighting effects rather than due to changes within the cloud. Thick clouds (*Ns, Cb*) of course restrict the amount of sunlight that is permitted to reach cloud base, giving them a charcoal or dark grey appearance when viewed from below.

Clouds often appear red and orange at dawn and dusk because of the low sun angle. Under these conditions, the path that light rays travel through the atmosphere is much longer than when the sun is high in the sky. The longer the path length the more the sunlight is depleted of its blue and green wavelengths. This means that the light reaching the clouds is rich in red and orange wavelengths, which when scattered by the clouds gives them their red/orange coloration. At dawn (dusk), high clouds are illuminated before (after) lower level clouds, due to the curvature of the earth.



Rainbows <u>do not form in clouds</u>, rather they form in rain droplets that are between cloud base and the ground. The illustration below shows the reflection and refraction pattern for light as it passes through a rain droplet (Figure 5.9). Refraction is the 'bending' of light rays as they pass from one material into another, such as air-to-water or water-to-ice. 'Bending' occurs because the speed of light is considerably slower in water and ice than it is in air (the speed of light in water and ice is about 75% of it's value in air).

White light enters a rain droplet and under goes two refractions and one internal reflection, before emerging from the droplet in distinct colors. Since violet and blue wavelengths are refracted (bent) more than yellows and reds, the inside colors of rainbows are violet and blue while the outer edges are reddish. Double, and the rare triple rainbow form in the same fashion as illustrated here, except that they undergo two and three internal reflections, respectfully. Also note that rainbows form when the sun is behind the observer. Rainbows do appear to move as the observer moves, this results from the fact that as long as the geometric reflection and refraction relationships are satisfied within the droplets, a rainbow will form.

When layers of cirrus or cirrostratus obscure the sun, halos often form around the solar disc because sunlight is refracted from ice crystals contained within the clouds (this is also true for ice fog). Halos form at an angle of 22^{o} from the sun. Unlike rainbows which form when light rays are reflected off the backside of rain droplets, light rays passing through ice crystals generally do not reflect off the back edge. The light simply passes through the crystal undergoing two refractions, the first as light rays transition from the air into the crystal, and the second as the light exits the crystal back into the air. Halos have less distinct color separation than rainbows because ice crystals do not refract light as well as water drops.

Mountain Cumilform Clouds

In this section a number of mountain cumulus cloud development scenarios will be presented. As a starting point, consider a mountain valley in which there are no synoptic winds (Figure 5.10a). The area where cumulus clouds (hereafter referred to as cumulus) first develop is primarily a function of the thermally generated winds. Over an idealized east-west valley, sunlight heats up the eastern slopes in the morning hours producing upslope flow. This favors cumulus development over the ridge of the eastern slopes. This scenario is reversed in the afternoon, when the western slopes receive the largest amounts of incoming shortwave radiation.

Whether or not cumulus develop on any given day depends on: the amount of available moisture, the volume of ascending air, surface heating, as well as the stability of the middle troposphere. Drier mountain environments such as those found in New Mexico, Arizona, and Nevada, tend to develop strong thermal winds during the day, however, cumulus development may be delayed or absent due to a lack of moisture When the surface is wet however, more shortwave radiation is used in evaporation, which means less energy is used to heat the ground, which translates into weaker flows. As you can see there has to be a balance between thermally generated winds and available moisture, one without the other limits cumulus development. Cumulus clouds do form when thermally generated winds are weak or non-existent. In this case, thermals will form over areas of the surface that heat up more rapidly than adjacent areas. (ridge lines).

You will rarely find cumulus developing over the center of a valley in the morning, because the air over the valley is usually sinking. Cumulus may drift over a valley, but they will usually dissipate in a few minutes when they do. One of the important consequences of mid-morning cumulus formation and dissipation is that despite the short life cycles of these clouds, they transport moisture from the surface into the mid-troposphere. This essentially pre-conditions the troposphere for more vigorous afternoon convection. As cumulus clouds grow larger they become self generating, in other words, they are able to move away from their place of origin and still remain vigorous. This is evident by mid-afternoon in many mountain locations when *Cg* and *Tcu* are located over valleys and other non-source regions.

One of the exceptions to the aforementioned rule of thumb concerning cumulus development over valleys, occurs when a glacier wind converges with a valley wind, as illustrated in Figure 5.10b. When the synoptic pattern favors light ambient winds, strong glacier winds develop at night in response to radiative cooling directly above the glaciers surface. At lower elevations where all of the snow has melted, valley winds usually develop by mid-morning. The two opposing winds typically collide near the terminus of the glacier, where they form a convergence zone. Since the air has no place to go except up, a well defined updraft is created. If there is sufficient available moisture either from the glacier or the valley, small cumulus will usually form in the updraft. One of the better locations to observe phenomena is around the base of the symmetrical volcanoes of the Cascades.

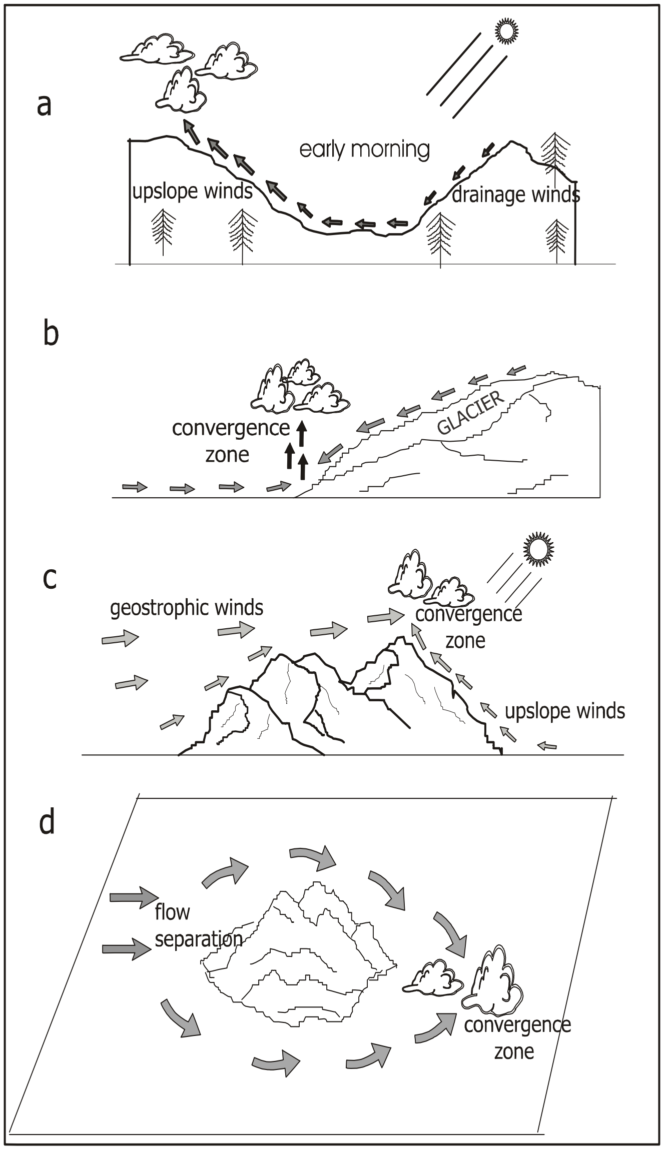


Figure 5.10- Development of cumulus clouds over various types of terrain.

When light to moderate geostrophic winds are blowing across the summit of a mountain**,** it is common for cumulus to form where the geostrophic wind and upslope or plain-to-mountain flows converge (Banta 1990). This is usually occurs on the downwind side of the ridge as seen in Figure 5.10c. Another example of cumulus formation is illustrated in Figure 5.10d. In this case flow is blocked on the upstream side of a large mountain. As air moves around the sides of the mountain it typically recurves on the backside, and at times it can form a lee-side convergence zone.

In large mountain ranges wind patterns are very complex, for example, it is possible to have several scales of thermally generated winds present at the same time. Upslope, valley, and plain-to-mountain circulations are usually operating by mid-day. Superimpose ambient winds that flow over and through the mountains with thermally generated winds, and it is easy to understand why such a broad spectrum of cloud scenarios are possible on any given day.

Towering cumulus (Tcu) form when either the low-level forcing increases or when the mid-troposphere becomes convectively unstable. *Tcu* originate as small *Cu*, which means that they form in the same preferred locations. Thirty to sixty minutes after the initial development of a *Tcu* or *Cb*, large precipitation sized droplets and/or ice crystals start to fall through the bottom of the cloud (Figure 5.11). Water droplets and ice crystals fall (both are called hydrometeors) out of the cloud because their fall velocities exceed the speed of the updrafts within the cloud. During the first few minutes of precipitation, many of the hydrometeors evaporate before they reach the ground (virga), this has an important effect on the air beneath the base of the cloud. Evaporation (or in the case of ice crystals: sublimation-) cools the air below the cloud, leading to an increase in its density, which in turn produces a downdraft. These downdrafts (gust fronts) often reach the ground and are capable of lifting surface air back into the cloud. In addition, evaporation below the cloud moistens the air, allowing subsequent precipitation to reach the ground because of reduced evaporation in this layer. The mountain traveler will recognize a gust front by a sudden blast of cold wind. Precipitation, thunder and lightning will follow in a few minutes.

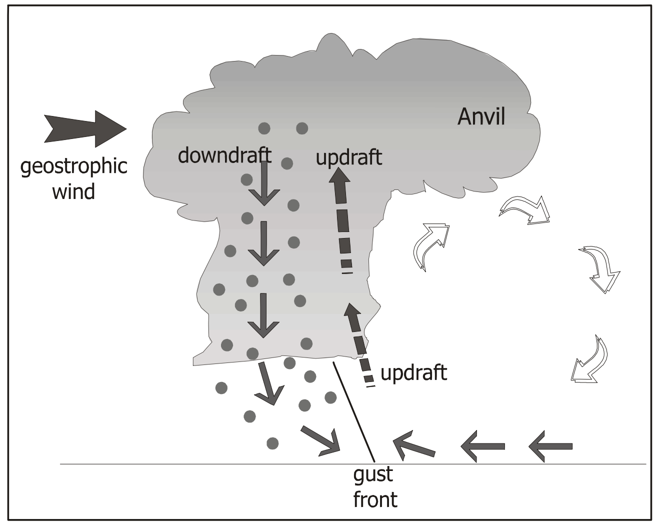
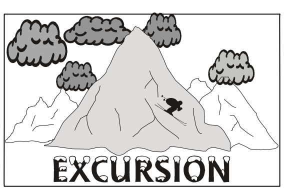


Figure 5.11- Cumulonimbus.

Evaporative Cooling

One aspect of convective storm activity that most mountain travelers are unaware of, is the potential for a rapid decrease in air temperature, due to evaporative cooling near a precipitating cumulus cloud. Consider the following scenario: imagine that you and your dog Spot, are enjoying a nice summer dayhike somewhere in the central Rockies. Over the course of the day you observe the development of *Cu* and *Tcu*. By mid-afternoon the air temperature is 25^{o} C (77^{o} F) and you notice it is just starting to rain. Over the next 30 minutes there is a very rapid and pronounced decrease in air temperature (> 12^{o} C or 20^{o} F). An analysis of this event would suggest that prior to the start of rain, the air beneath cloud base was warm and *very dry*. As rain fell through this dry air most of it was evaporated, which we noted in the previous section, cools the air. Evaporative cooling continues until the air reaches its wet bulb temperature, which is by definition: the temperature that a parcel of air can be cooled, by evaporative cooling (if you put a wet cloth on a thermometer, you can watch the temperature decrease to the wet bulb temperature). The drier the air, the lower its wet bulb temperature. Hence, drier mountain ranges such as the Rockies and Eastern Sierras experience greater evaporative cooling than the Coast Ranges or the Cascades. It is important to realize that these rapid temperature changes do occur quite frequently during summertime convective storms. In



addition, since most *Tcu* or *Cb*, develop significant localized winds in the form of downdrafts and gust fronts, not only does the ambient temperature decrease but the wind chill temperature drops substantially as well.

Hail:

Because the updrafts in *Tcu* and *Cb* can obtain speeds of 10-15 m s^{-1} (22-33 mph), large water droplets and large ice crystals are frequently transported to the top of the cloud, where temperatures range from -30° to -40° C (-22° to -40° F). Hail forms when *supercooled liquid water* (water that has a temperature below freezing but is still in liquid form) is collected and frozen on a embryonic hail nucleus, as it is transported through the upper regions of the cloud. As a hailstone grows, it starts to fall down through the cloud, accumulating more water droplets in the process. Researchers speculate that giant hail forms when "normal" sized hail is caught in a series of updraft/downdraft couplets, which means that it makes several vertical excursions through the cloud before it finally has enough mass to fall free of the strongest updrafts. Of course this scenario suggests that some very large updraft velocities must exist in these clouds. Most hail-stones are pea size. Giant hail can reach 10 cm (4 in) in diameter, hail of this size is fairly rare and typically only occurs away from mountainous terrain. Most hailstorms last for several minutes, and are often followed by a short period (< 5 minutes) of heavy rain.

Lightning

Lightning is an electrical discharge that occurs either within a large cumulonimbus cloud (intra-cloud), or between the cloud and the ground (cloud-to-ground). It is not known with certainty how clouds become electrified, although a number of theories have been established. What we do know is that in a typical *Cb,* a large portion of the upper half of the cloud contains positively charged ice crystals, while the middle half of the cloud contains negatively charged ice crystals intermixed with supercooled water droplets (Figure 5.12). Most of the electrified region of the cloud occurs in parts of the cloud that are colder than 0° C (32° F). The mostly widely accepted theories for cloud electrification suggest that as hail, small ice crystals, supercooled liquid water, and graupel are carried in updrafts and downdrafts, they collide, creating the separation of electrical charge on the surface of the particles. Electrification can occur in developing cumulus, however it reaches its greatest intensity in *Cb*.

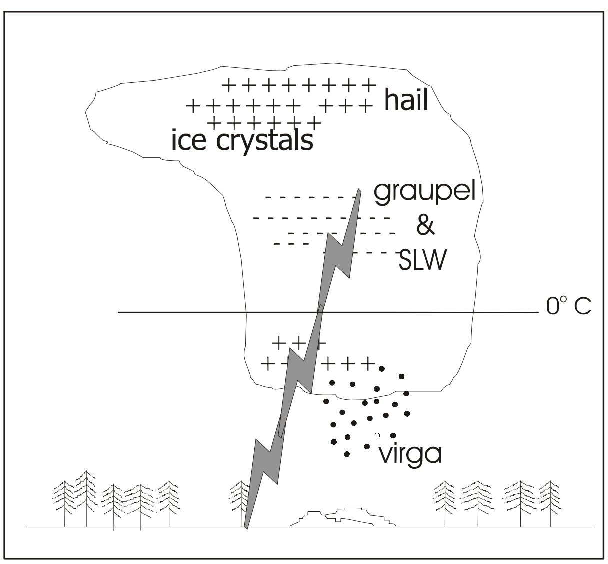


Figure 5.12- Cloud electrification. + sign indicates areas of the cloud which are positively charged, while - indicates negatively charged regions.

The most prolific researcher and writer on the subject of lightning is Martin Uman, much of the material outlined below is taken from his book: All About Lightning (1986). Most cloud-to-ground lightning involves negative current, that is: electrons flow from the cloud to the ground. The sequence of a lightning strike is as follows:

1. Electrons move from the negatively charged region of the cloud towards the ground in a zig-zagging path called a stepped leader. This path

or conduit of high current is invisible,
and ranges from 1-10 m (3-33 ft) in diameter.

2. As the stepped leader nears the ground (or tree, building, etc.), it induces positive charge segregation within the ground. As a result, the ground initiates a upward moving discharge conduit of its own, which in a few ten-thousandths of a second connects with the stepped leader.

3. As large amounts of negative charge move from the cloud to the ground, a visible channel of lightning known as the return stroke is initiated. This bright discharge occurs so fast that an observer typically sees one simultaneous flash. Temperatures within the conduit can reach 33,000^{o} C.

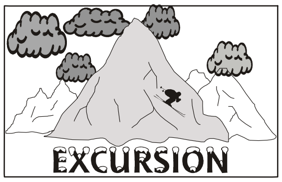
Thunder is produced when sections of the electrical conduit are heated so rapidly that areas of very high pressure are created (10 to 100 times atmospheric pressure). These areas of high pressure generate sound waves which travel at speeds near 330 m s^{-1} (730 mph) . If you have the misfortune of being within several hundred meters of a lightning strike, you will hear a single loud bang, followed in the next few seconds by a series of hissing and crackling sounds. If the strike is more distant, you hear the more familiar series of rumblings, as sound waves from the closest part of the strike arrive before the more distant parts. According to Uman (1986), the average person cannot hear thunder generated by lightning that is farther than about 20 km (12 mi). The old rule of thumb that thunder travels one kilometer in three seconds (or one mile in five seconds) is valid. As a point of safety, keep in mind that the closer a *Cb* is to the ground, the higher the statistical probability of a cloud-to-ground strike.

Lightning is much more common in drier continental mountain ranges than in coastal ranges. This may seem to be a conundrum since the air over coastal ranges has a higher water content. However, cloud moisture is only one essential requirement in thunderstorm development, high vertical cloud development being the second requirement. During certain times of the year, drier mountain ranges have the right combination of the two, hence the southern Rockies of Colorado and New Mexico have some of the highest incidence of lightning in the western US. The southern Rockies are lightning prone because it is a region where warm moist air from the Gulfs of Mexico and California converge with cool dry air moving down from the north. In a study of thunderstorm development in Colorado, Banta and Schaff (1987) found that certain mountain ranges were preferred thunderstorm source regions. Many of the taller peaks in the Front Range were high on that list.

On the micro-scale, lightning strikes are fairly spatially random except near protruding isolated points. So if you're caught in the open during a lightning event, do not seek shelter near a isolated tree or rock. Lightning data collected from the top of the Empire State Building shows that the top of the building has a much higher frequency of lightning strikes than the sides of the building a short distance from the top. However, do not be fooled into thinking that you can stand on a ridge with immunity during a lightning event simply because you are not on the highest point of the ridgeline. Lightning does not necessarily strike the highest terrain in a given local area.

Survival Tip: Lightning data shows that the most prolific number of strikes in the mountains occur between
3-5 PM local time, although there is mounting evidence that there are some regional differences. For example, thunderstorm development and associated lightning often peaks in the afternoon and early evening, over the tallest mountains. By late evening however, thunderstorms and lightning frequently

becomes more common over the foothills and plains.



Thunder and lightning is often followed by periods of moderate to heavy rain and/or hail. Some additional safety considerations during lightning events are: stay out of water, including shallow surface ponds, discard all potential lightning rods from your backpack: metal edge skis, ski poles, ice tools, fishing poles, metal backpack frames, etc. If caught in the open, lie down on the ground, hopefully on a sleeping pad. The best advice is also the easiest to follow, be weather smart, monitor lightning before it becomes a problem. This may entail a earlier morning start so that you can be off exposed ridges and summits by mid-afternoon. You may even consider delaying a trip by a few days in order to take advantage of a period when convective activity is reduced, this may occur in response to changes in air mass types or a reduction in available moisture.

Water Droplets and Ice Crystals

Up to this point in this chapter we have concentrated on the fundamentals of cloud formation. We now shift our discussion to the generation of cloud droplets and ice crystals from water vapor, a topic referred to as microphysics. This is a fascinating topic but it can also be quite complex, so only a brief overview of the subject will be given at present. You are probably wondering how meteorologist study and gather cloud data. There are three basic ways this is done: 1) Mountain top observatories, such as Elk Mountain (3307 m) in Wyoming and Sonnblick (3106 m) in Austria, collect air samples as clouds move over the stations; 2) Specially equipped airplanes fly through clouds collecting cloud droplets and ice crystals as well as observing cloud humidity, temperature, droplet and crystal concentrations, and; 3) Clouds can be created and studied in a laboratory using a device called a cloud chamber. The advantage of a cloud chamber is that the temperature, pressure and RH can be precisely regulated allowing a wide range of cloud types and conditions to be studied.

Fundamental to the understanding of microphysics is the transition of water vapor into liquid droplets and solid ice crystals. Cloud droplets or ice crystals form when water vapor condenses on some type of microscopic foreign material which happens to be floating around in the air. Foreign material consist of grains of dirt, small salt particles, bits of vegetation, and aerosols, which are carried into the troposphere via strong winds, volcanic eruptions, and forest fires to name a few. Once a droplet or crystal forms however, it may re-evaporate in a few minutes due to small fluctuations in the relative humidity of the cloud. With so many droplets forming in a cloud, there is fierce competition between the droplets for the remaining water vapor. It turns out that most cloud droplets or ice crystals never grow large enough to be able to survive the fall to the ground. Therefore, most clouds never produce rain or snow. Only when a select few droplets (or crystals) within the cloud grow very large, are they capable of forming raindrops or snowflakes.

ITS RIME TIME

Riming occurs when a cloud containing supercooled liquid water (SLW) moves across the summit of a mountain, at which time rocks, buildings, trees, etc. get coated with a layer of ice. One of the easiest places to observe riming is on ski lift towers; notice how the ice builds up in the direction of the prevailing wind. Riming occurs much more frequently in coastal mountains than in continental mountains because of the higher water content in the former. It is most often found at mid-tropospheric levels (700 mb) in the presence of inversions or relatively warm layers (0^{o} to -15^{o} C), which may occur in conjunction with occlusions or warm fronts.

One of the authors own riming experiences follows. During a climb of Cotopaxi (5898 m) in Ecuador, my partner and I made the summit climb in a cloud full of SLW. After about 5 hours of climbing we were both encased in about 1-2 cm (0.3-0.7 in) of ice. Fortunately, the standard route on Cotopaxi is not technical, I mention this because belaying under those conditions would have been ugly. There is no way to keep a layer of ice from building up on the rope. A number of years earlier, while on a March climb up Mt Hood's Leuthold couloir (3426 m), we were caught in a rapidly developing storm. Strong winds were blowing clouds of SLW over the summit ridge. I was wearing a pair of sun glasses at the time, and I distinctly remember my upper and lower eyelashes freezing together. Ski goggles probably would have given better protection under those conditions. The point in relating these two incidences is to stress that climbers should use extra caution if they are caught in a riming situation.

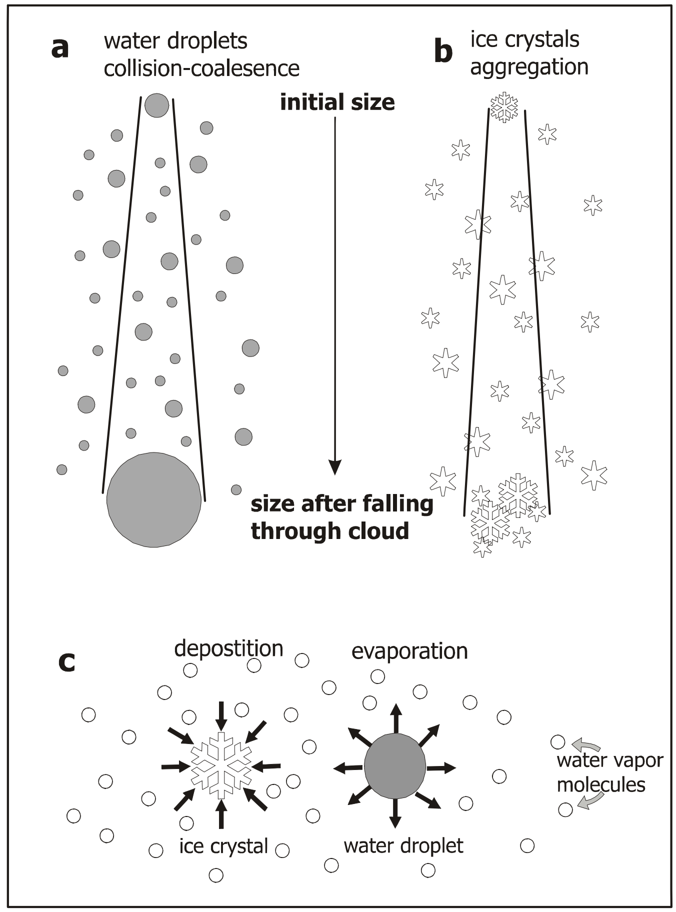
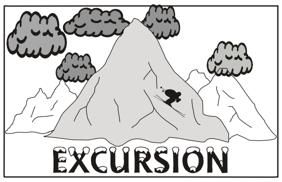


Figure 5.13- Growth of precipitation.

Most cloud droplets and ice crystals have diameters around 0.001 to 0.5 mm (0.00004 to 0.02 in). In order for them to ever have a chance of falling to the ground, they must grow to a size of 1 mm (0.04 in) or larger. Growth occurs in several ways. In clouds that are composed entirely of water droplets, many of the small droplets evaporate, producing extra water vapor which is in turn made available to the larger droplets. Once this process is initiated, a select few droplets grow quite rapidly. When droplets reach a size of about 0.1 mm (0.004 in) in diameter, they start to fall through the cloud relative to the smaller droplets. As the larger droplets fall through the cloud they collide and coalesce (merge) with the smaller droplets in a process called collision-coalescence (Figure 5.13a). Large droplets exit the base of the cloud as soon as their fall velocities exceed the speed of any updrafts. In clouds that contain a mixture of cloud droplets and ice crystals, due to different bonding properties, ice crystals can grow at the expense of water droplets. When this occurs some of the ice crystals grow large enough to fall through the cloud, colliding and merging with other ice crystals, in a process called aggregation (Figure 5.13b). Large snowflakes are composed of individual crystals that have aggregated on descent through the cloud.

Once rain sized droplets fall through the base of the cloud they are susceptible to evaporation, more so if the ambient air is very dry, such as commonly occurs beneath towering cumulus clouds. Rain drops typically range in size from 1-5 mm (0.04-0.2 in) and have fall velocities from 1-10 m/s (3-33 ft s^{-1}).



Rain droplets do not experience continued acceleration as they fall. Basic physics tells us that the rate of acceleration for two free falling bodies of unequal mass, is the same for both bodies (Galileo's famous rock experiment from the leaning tower of Pisa). However, this holds true as long as the speed of each object is below its terminal velocity. Due to aerodynamic drag, the terminal velocity of a free falling object is a function of its shape (Table 5.2). The amount of drag on a falling body is also a function of the density of air, at 5000 m (16,400 ft) for example, a droplet will travel about 20% faster than at sea-level. Most rain droplets have terminal velocities around 10 m/s (33 ft s^{-1}), and the most common droplet shape is spherical. Larger droplets become oblong as they fall, which makes them susceptible to breaking into smaller droplets. The whole rain process is all about the evolution of small cloud droplets into larger rain drops. In this case the evolutionary path is full of obstacles, the result being that few clouds have the 'right stuff' to produce precipitation.

Pristine ice is the name given to the most fundamental of ice crystals, those having a single crystalline structure. The generic term snowflake is most often used for aggregated ice crystals which reach the ground. Of course ice crystals can melt on the decent from the cloud, forming small water droplets (drizzle). If liquid droplets freeze on descent and have a solid structure before reaching the ground, they are called ice pellets. If a droplet does not freeze until it makes contact with the ground it is considered *freezing rain.*

Table 5.2 Properties of rain and snow

Type	**Phase**	**Size (mm)***	**Fall Speed (m/s)***	**Notes**
Drizzle	Liquid	0.05-0.5	3-5	Very light rain from stratus
Rain	Liquid	0.5-3	5-10	Light to moderate intensity
Rain (heavy)	Liquid	3-6	10+	From *Tcu, Cb*
Pristine Ice	Solid	0.5-2	0.5	Disc or needle shape
Aggregates	Solid	3-15	0.5-1	Classic snowflake
Graupel	Solid	2-5	1-3	Spherical structure, rimed
Hail	Solid-Liquid	10-100	>10	Can have a sponge structure

* data taken from Rogers & Yau (1989)

IS IT EVER TOO COLD TO SNOW ?

There is a common misunderstanding that it can get too cold to snow. Even in regions that are very cold, it is possible to receive small amounts of snow or settling ice fog during the coldest part of the year. Granted it's not very much snow, but even at these extreme temperatures the atmosphere contains small amounts of moisture. It is true that as parcel of air gets colder, its capability of holding moisture is reduced. However, as discussed in a earlier chapter, in the winter at high latitudes and in mountainous regions, the temperature in mid-level clouds is frequently considerably warmer than the temperature at the surface. You should recognize this as a temperature inversion. As a result, the temperatures at which snow crystals are forming and growing, is not as nearly low as you might think it is. Therefore, the correct answer to the question: "Is it ever too cold to snow?" is NO. Keep in mind however, that in general, heavy snowfalls (accumulated depth) are associated with warmer surface temperatures (0^o to -5^o C or 32^o to 24^o F) and light amounts of snowfall with colder temperatures. Along the same lines, it can be very difficult trying to estimate the density of freshly fallen snow based on surface air temperature. For example, we have seen some very dry snow fall while the air temperature was a balmy -3^o C ($+27^o$ F), and conversely we have seen some wet snow fall while the thermometer read - 6^o C ($+22^o$ F)

Mountain Precipitation

We will now turn our attention to the topic of orographic precipitation, which is the study of the formation and distribution of rain and snow in mountain environments. The material in this section is applicable to mountains located in middle and high latitude. We will address the unique situation in the Tropics at the end of this section. Comprehensive knowledge of the timing (start and stop times), and the quantity of precipitation is vital to the forecasting of flash floods, mud slides, river flooding, avalanche mitigation, snow removal logistics, and road closures to name a few. Since precipitation is the result of very specific dynamic and thermodynamic processes, it is the most difficult weather element to forecast. The primary controls of orographic precipitation are: storm type (stratiform versus convective), prevailing wind direction, terrain configuration (small hills versus large mountain range), and climatic zone (maritime versus continental).

<u>Stratiform precipitation over hills</u>

Numerous field studies have shown that precipitation increases in and around hilly terrain, when compared to the observed precipitation in the surrounding low-lands (Browning 1980). This precipitation enhancement occurs for individual precipitation events as well as seasonal and annual precipitation totals. Bergeron (1960) proposed the seeder-feede*r* conceptual model, which is illustrated in Figure 5.14, as an explanation for this enhancement. Precipitation increases around a hill because the hill generates a low-level stratus cloud (called a feeder cloud), that would otherwise be absent. In this model, the seeder cloud forms due to frontal lifting, which is independent of any terrain forcing.

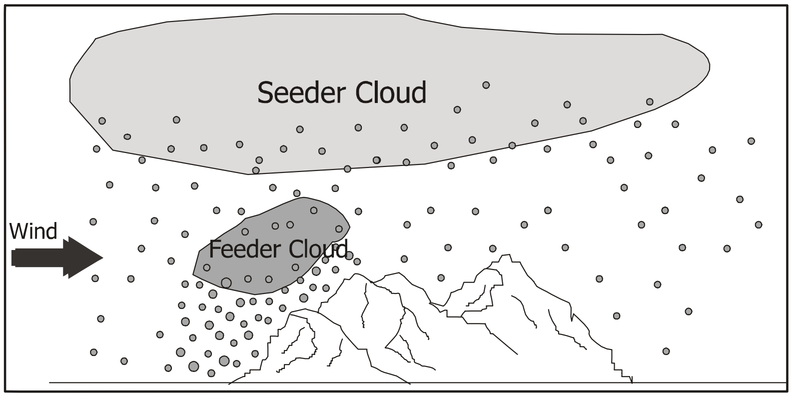


Figure 5.14- Seeder-feeder conceptual model for the enhancement of precipitation.

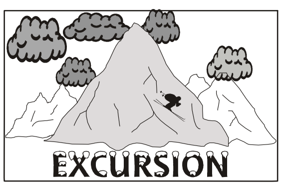
Once precipitation is initiated in the seeder cloud, it falls through the feeder cloud where it either gains mass due to collision-coalescence (or in the case of ice crystals-aggregation), or it evaporates within the feeder cloud, adding moisture. The Bergeron model is an idealization; we generally do not observe two separate clouds over this type of terrain. However, this model is applicable to a single thick cloud, where the upper region acts like the seeder cloud and the lower region plays the role of the terrain generated feeder cloud. The amount of precipitation enhancement varies from storm-to-storm and season-to-season, but an increase of 30-50% above the background precipitation is common for isolated hills lying on a flat plain. The distribution of precipitation in hilly terrain on the micro-scale, is in many cases independent of local topography. In other words, due to wind drift and the short distance between hills; leeward slopes and valleys frequently receive as much precipitation as the windward slopes. Keep in mind however, as a whole, hilly terrain receives considerably more precipitation than the surrounding plain because of the seeder-feeder mechanism.

<u>Stratiform precipitation over mountains</u>

Unlike the spatial distribution of precipitation commonly observed around hills and small mountains, <u>larger mountains</u> often exhibit a pronounced increase in precipitation from the base to some height on the windward slope, after which precipitation decreases with height toward the summit. This means that the precipitation maximum (storm, seasonal, or annual) occurs below the summit; this certainly holds true for coastal mountains located in the middle and high latitudes, but is less evident for mountains located in the interior of the continents. It is important to recognize the differences between the spatial distribution of precipitation on the local-scale from that which occurs on larger scales. For example, when Schermerhorn (1967) analyzed precipitation data taken from the central Washington Cascades along a west-east transect, he found that elevation alone was not the only factor determining annual precipitation. What he did find was that distance from the ocean was also a major contributing factor. Two stations that differed in elevation by 1000 m (3,280 ft), might have identical annual precipitation. The lower station however was located closer to the coast, so what it lacked in terrain forcing capabilities, it made up for by its larger moisture content. In contrast, Hjermstad (1970) reported that winter precipitation across central Colorado, increased with elevation. The seasonal precipitation ratio between Grand Junction and Vail Pass was 1 to 5.9, but varied according to wind direction and wind speed. On the local-scale, stratiform precipitation in the central and southern Rockies usually reaches a maximum near the tops of the mountains. This trend for precipitation to increase with height is due to the fact that the lowest kilometer of the troposphere in continental regions is quite dry. What moisture there is, resides well above the surface, hence considerable lifting has to occur before precipitation is generated.

An additional complicating factor in our understanding of the distribution of precipitation in mountainous terrain is that the majority of precipitation gages are located at lower elevations. This means that the data is biased toward lower elevations (valleys, passes through the mountains, etc.), and in many regions we have little idea how much precipitation occurs at higher elevations. Consider the following example as a case in point: In a study of June-October rainfall in the central Himalaya, Dhar and Rakhecha (1980) found that there were two zones of maximum rainfall. The first was located at the base of the foothills, just north of the Indian Plains. The second maximum occurred at an elevation of 2200 m (7,200 ft), after which rainfall diminished rapidly with height. Their analysis was based on 44 gages, 11 of which were located above 2000 m (6,500 ft), only two of which were higher than 3000 m (9,800 ft). This study indicates that mesoscale rainfall during the summer monsoon decreases as one moves into the interior of the Himalaya, but it also highlights the bias towards low elevation gage data. If more gages were available at higher elevations the data might reveal a more complex rainfall pattern such as the presence of a third maximum at some elevation above 3000 m (9,800 ft). Another factor which affects precipitation at a particular gage is the height of the surrounding terrain. For example, in the Nepal Himalaya, a gage located at 3500 m (11,400 ft) along the sides of a north-south oriented valley can receive considerably more precipitation than a gage located at the same elevation, but in a valley that is oriented east-west. In the former valley, moisture is easily transported up the valley, while in the latter case, the higher terrain blocks a considerable amount of moisture from reaching the gage. This 'terrain effect' is more important in areas where the bulk of the moisture is in the lowest levels of the troposphere, as it is in the eastern Himalaya during the summer.

ESTIMATING ALPINE PRECIPITATION



There are several ways of estimating high elevation seasonal precipitation without using a precipitation gage. One method is to try and estimate the depth of snow on the ground just before the summer melt season begins. This is much easier said than done, especially in areas with a deep snowpack. The first objective is to estimate the average snow depth over a given area (i.e.-basin). The second objective is to estimate or measure the average density of the snowpack. When the average snow depth and density are multiplied together, the result is a number which represents the weight of the snow per unit area. However, what we are really trying to determine is the water equivalent of the snowpack. The term *snow water equivalent* is used to designate how much water would be produced if the snowpack is melted. For example, if the average snow depth is 1.0 m and the average density is 250 kg m^{-3}, the resulting weight is 250 kg m^{-2}. Since fresh water has a density of 1000 kg m^{-3}, the water equivalency in this example is 0.25 m (250 kg m^{-2}/1000 kg m^{-3}). So if the entire snowpack is melted, it would produce a pond of water 0.25 m deep over the entire basin.

Snow depth and the snow water equivalent for alpine areas is collected by one of two ways: an automated snotel station or on a snow course. Both of these data collection points are run and maintained by the National Resource Conservation Service (NRCS). Snotel stations consist of a snow pillow, which looks like a square piece of metal about a meter and a half in length. In side is a fluid that is hooked up to a meter which reads the pressure of the fluid. When snow collects on the pillow the weight of the snow increases the pressure in the fluid, this value is then transmitted back to a central station via satellite. Hence the name SNOW TELEMETRY or snotel. Frequently, this stations also collect precipitation using a standard rain can. Snotel stations have been around since the late 1970's.

Since snotel sites are expensive to setup and operate, additional snow data is collected the old fashion way: by hand. Snow courses consist of a designated area where a series of from 5 to 10 stakes are set up, and snow depth and snow water equivalent are measured once a month. Snow surveyors use a specially designed snow tube which gives the depth and when weighed with a special scale, yields the water equivalent as well. Snow course data is typically gathered at the beginning or end of the month.

Another method used to estimate alpine precipitation is to monitor river runoff over the course of the year. In areas that do not have perennial snow or ice, this technique is pretty straight forward. In glaciated areas you have to factor in snow and ice melt that occurs from pre-existing precipitation. This can add considerable error, but nevertheless does provide a rough estimate of annual precipitation.

WEB: www.wcc.nrcs.usda.gov (NRCS web site)

The strongest control over orographic precipitation is without a doubt wind speed. Increasing winds have two interrelated affects on precipitation; first it increases the inflow of moisture, and; secondly, it increases the strength of the terrain forced lifting; the net result is an increase in precipitation. Hjermstad (1970) in his Colorado study, found a high correlation between increases in geostrophic wind speeds and increased precipitation, at least for elevations above 2300 m (7500 ft). Wind direction is also important because terrain lifting is strongest when the incoming flow is perpendicular to the long-axis of a mountain range.

Blocking can also enhance precipitation, however the enhancement often occurs upstream of the mountain, as well as over the lower windward slopes. In several documented cases, Peterson *et al* (1991) found that wintertime precipitation could be significantly enhanced 30-60 km upstream of the Gore Range of northern Colorado. To the east of Colorado's Front Range, a considerable amount of the region's winter precipitation is due to upslope flow (easterly surface winds) during blocking events. Without these blocking episodes more moisture would be transported into the Front Range, resulting in less snowfall in eastern Colorado.

In a region of very high mountains such as the Karakoram of Northern Pakistan, the few precipitation gages that are in existence are usually located at the bottom of deep valleys. As a result Flohn (1970) estimated precipitation in the higher elevations of the Karakoram using river runoff data. He found that valley gages only catch about 1/20 to 1/30 of the precipitation that falls at higher elevations. This means that the valleys are quite arid while the mountains contain some of the largest glaciers outside of the polar regions.

Convective orographic precipitation

It was noted earlier how small cumulus clouds typically form over ridgelines. However, as *Cu* grow into *Tcu* and *Cb*, they tend to become decoupled from the underlying terrain. On the local-scale, the distribution of precipitation from cumulus clouds is quasi-random, that is valleys as well as leeward slopes are as likely to receive as much precipitation as windward slopes. In a study conducted along the Front Range, Jarret (1990) found that summer (convective) 6-hour rainfall totals were considerably higher for elevations below 2300 m (7,500 ft). In other words, convective storms that produce large amounts of rain, are more frequent at elevations below 2300 m (7,500 ft). He also reported a latitude dependancy for this relationship. In New Mexico, for example, the largest rain producing storms occurred below 2400 m (7,800 ft), while in Idaho and Montana this level was around 1600 m (5,200 ft).

Tropical Mountains

When we discuss meteorological events in the tropics we often need to distinguish between equatorial (10^o N to 10^o S) and tropical (10^o N/S to 25^o N/S). In the equatorial zone the surface and mid-tropospheric synoptic winds are weak and from the east. The higher terrain in this zone lies in east Africa, the southwest Pacific, and the northern Andes. These three regions all have different climate and weather patterns, therefore we will not attempt to discuss each one. For a small country, Ecuador has a very complex climate regime, in large part due to the fact that it is sandwiched between the Amazon and the Pacific Ocean. Despite all of the local effects that occur in these regions, maximum precipitation in the equatorial zone generally occurs below 1500 m (4900 ft). This is a result of large amounts of water vapor found in the surface layer. This should not be interpreted to mean that it does not rain (or snow) at higher elevations, what it does mean is that the total amount of rainfall (storm, seasonal, or annual), does decrease substantially with increasing elevation.

The distribution of rain in these regions is also a function of the prevailing wind direction, which can cause significant changes in the supply of low-level moisture from one season to the next or from one stormy period to the next. Garreaud (1999) studied the temporal distribution of precipitation over Bolivia's Altiplano (average height 3800 m). His results show that during the December-March wet season, rainy periods and dry spells lasted roughly a week each. Since the Altiplano is located well interior of the South American coastline, its primary source of moisture is the Amazon Basin. During rainy periods easterly flow from the Amazon was considerable stronger than the southwest winds which occur during dry periods. Easterly flow transports large amounts of moisture over the eastern slopes of the Altiplano, while southwesterly flow has to cross the dry deserts and mountains of northern Chile, arriving over the Altiplano with little remaining moisture.

In the sub-tropical mountains of Hawaii, northern Chile, and North Africa, regional weather is dominated by the presence of subsiding air from sub-tropical high pressure systems. This subsiding air originates from the descending arm of the Hadley Cell, and produces a very pronounced temperature inversion which goes by the name of Trade Wind Inversion (TWI). It is no coincidence that the most of the earth's major deserts are located on the eastern sides of these sub-tropical highs. The TWI forms at the relatively low altitude of 2000-3000 m (6500-9800 ft) over the central Pacific and Atlantic Oceans (in both Northern and Southern Hemispheres). These strong inversions limit the vertical development of cumulus clouds, and in so doing, greatly restricts the availability of moisture and precipitation at elevations above the inversion. This is very evident on the Island of Hawaii, where Mauna Loa (4170 m) and Mauna Kea (4206 m) extend well above the TWI. The zone of maximum precipitation on these mountains is located at the base and along the lower slopes, while the summits are extremely dry [likewise for Haleakala (3056 m) on Maui].

An interesting aspect of orographic precipitation in the Hawaiian Islands is to compare precipitation on Kauai and West Maui, with that of the neighboring islands. Smith (1989) suggest that Kauai's extreme rainfall is due to a combination of factors; primarily the height of the mountains and the width of the barrier. The height of Kauai's mountains (1200-1550 m) are ideal with respect to the height of the TWI. This height relationship means that the moist northeast trade winds are able to ascend over the top of Kauai, which of course produces copious amounts of rain. If the mountains were another 500 to1000 m (1600-3200 ft) higher, they would block much of the flow, causing it to divert around the sides of the island, reducing rainfall at higher elevation. If the mountains on Kauai were lower, they would produce considerably less rain because of reduced lift. In addition, Smith points out the importance of width of the barrier to the production of orographic precipitation. Case in point, compare Molokai's seasonal precipitation (6 m) to Kauai's (10 m), both islands have mountains with similar heights and both are oriented perpendicular to the northeast trade winds. However, Molokai's mountains are very narrow compared to Kauai's. This narrow barrier produces considerably weaker lifting, and as a result subsequently less rainfall is produced over Molokai.

Summary

Table 5.3 is a cloud chart that summarizes cloud types and some of their attributes.

TABLE 5.3 Cloud Guide

Type	Height	Low Level Winds	Summit Winds	High Level Winds	Precip	Notes
Cirrus	8-12 km	NF	NF	ALL	NF	May indicate approaching front
Stratus	ALL	ALL	ALL	ALL	Pos-Prob	Reduces diurnal temperature range
Lee wave	Summit+	NF	Mod.	Mod-Stg	NF	Strong upper level winds
Cloud Cap	Summit+	NF	Mod-Stg	Mod+	Poss	Poor visibility
Banner	Summit	NF	Mod-Stg	Mod	NF	Strong summit level winds
Cumulus Humilis	1-2 km AGL	Lt-Mod	Lt-Mod	All	NF	Short lived, indicate fair weather
Cumulus Congestus	1-3 km AGL	Lt-Mod	Lt-Mod	All	NF	Aggregation of *Cu*
Towering Cumulus	1-12 km AGL	Mod	Mod	Lt-Mod	Prob	Heavy rain and hail possible
Cumulo-nimbus	1-12 km AGL	Mod	Mod	Lt-Mod	Prob	Hail and lightning, longer lived than *Tcu*
Warm Front	1-12 km Cirrus-Cirrostratus-Altostratus (precipitation possible) - Nimbostratus (precipitation probable). Time from first appearance of Cirrus to precipitation is 12-24 hours.					
Cold Front	1-12 km more cumulus than in a Warm Front, and also moves faster. Time from first appearance of Cirrus to precipitation is about 12 hours.					
Occluded Front	1-12 km Same as Warm Front.					

Key: NF=No Factor; All=All Possibilities; Lt=Light; Mod=Moderate; AGL=Above Ground Level; Poss=Possible (30-50% occurring); Prob=Probable (>50% occurring)

Remember that clouds not only affect the current weather but they can also give some indication of future weather as well. The wind at various levels is intended to give you a rough estimate of the speeds you would encounter if you were in or near the cloud. Some clouds only form when the winds are moderate to strong, while others form when winds are lighter.

Chapter Quiz

1. True/False: A cloud consists of many tiny liquid droplets and/or ice crystals held in suspension?
2. True/False: A wave cloud is only made-up of water droplets?
3. Cumulus clouds are often__________ than they are wider.
4. True/False: Lightning is really cool?
5. Supercooled liquid water exists at a temperature __________ freezing.
6. Orographic precipitation is primarily controlled by __________ speed.
7. As pristine ice crystals fall through a cloud and coalesce they are called __________.
8. True/False: Rainbows only form when the angle of the sun with respect to the water droplet is 22^{o} ?
9. In general, cooler air temperatures result in ___________ amounts of snowfall?
10. Cumulus clouds consists of updrafts and _____________?