PRACTICAL MOUNTAIN WEATHER

A GUIDE FOR HIKERS, CLIMBERS AND SKIERS

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CONTENTS

UNITS AND CONVERSIONS

		Metric units	English unit equivalent
distance			
km		Kilometers (1000 meters)	0.621 miles
m		Meter (100 cm)	3.38 ft or 39.27 inches
cm		Centimeters (10 mm)	0.3927 inches
mm		Millimeters	0.03927 inches
area			
km^2		Square kilometers	0.39 square miles
pressure			
mb		Millibars	1013 mb=29.92 inches mercury (sea-level pressure)
temperature			Conversions
^{o} C		Degrees Celsius	To convert to Fahrenheit: ((^{o} C) x 9/5) +32^o
^{o} K		Degrees Kelvin	^{o} C + 273^o
^{o} F		Degrees Fahrenheit	To convert to Celsius: ((^{o} C)-32^o) x 5/9

time

s Second

hr Hour

weight

gm Gram

kg Kilograms (1000 grams) To convert to lbs: (kg) x 2.2

misc

m/s meters per second

km hr^{-1} kilometers per hour

500 mb 500 mb pressure level

mph miles per hour

Introduction

This book is written for climbers, hikers, backpackers, backcountry skiers, and snowboarders; a group that we refer to as mountain travelers. By reading this book, not only will you gain an understanding of the complexities of mountain weather, but you will also appreciate the difficulties in forecasting it as well. The goal of the book is not to turn each and every reader into a weather forecaster *per se*; nevertheless, the more you understand about weather, the more likely you are to make well-informed decisions once you are in the mountains.

The mountain environment is no respecter of persons, thus the more the traveler knows about the environment they are passing though, the more likely they are to survive the journey. Extreme weather does directly kill a number of mountain travelers each year, primarily through lightning strikes and extreme cold (hypothermia). However, it is the indirect causes: whiteouts, severe winds, and heavy snowfall, to name a few, that frequently incapacitate mountain travelers. Applying what you have learned from this book will not, of course, eliminate the risks, but it will certainly stack the odds of survival in your favor.

Weather forecasting and analysis, and the study of mountain weather in particular, is not an exact science. Forecasting is part science and part experiential knowledge, with some randomness thrown in. Unlike a carpenter who builds a house exactly as the blue prints specify, the weather forecaster does not have a 'perfect' set of guidelines to base their forecast on. In other words, similar weather patterns do not necessarily produce the exact same weather. However, with a basic understanding of weather processes and with the aid of forecasting tools, a considerable amount of the mystery is removed.

Whether you realize it or not, everyone who travels in the mountains is a weather forecaster to some degree. Many technically proficient climbers, skiers, snowboarders and hikers have ventured into the mountains, only to meet disaster face-to-face because they either ignored a weather forecast or failed to recognize the signs of adverse weather. The person who ignores the weather, is still a forecaster, albeit a pretty dumb one. By choosing to ignore the weather, they are 'forecasting' that the weather is going to be survivable. This book is written to equip you with the necessary skills to make your own "on the mountain ' forecast as well as help you evaluate forecast that you receive from meteorological services.

If you have looked at the table of contents page, you probably noticed that this book can be divided roughly into two sections: the theory of mountain weather followed by a guide to weather and climate in various mountainous regions of the world. Some readers at the start may be 'put off ' by the more academic material, however, note that this chapters are loaded with hints on how to apply what you are learning. Note that this book is essentially a reference book and a guide book rolled into one. The power of knowing how processes work, in this case-how the atmosphere works-is that you do not have to know all the answers ahead of time. With a fundamental understanding you will be able to figure out why the weather is doing what it is doing.

We make no excuse for using the metric system as the primary units, with english units in parentheses. Metric is superior in every way shape and form! If you are not familiar with the metric system give it a try. In the body of the text we have <u>underlined</u> words, phrases or sentences that need emphasis-so take note. Words that are *italicized* are included in the glossary, located at the end of the book. In addition, from time-to-time we have inserted little side discussions called ‘Excursions’, which deviate somewhat from the main topic. They range from being very practical to a bit more off the wall!

Before proceeding however, take the following quiz to test your current knowledge of mountain weather. By the time you have finished reading this book you will be able to ‘breeze’ through these questions without hesitation.

Mountain Weather Quiz

1. Rising elevation on a stationary altimeter indicates: a) rising pressure b) decreasing pressure
   c) no change in pressure?
2) True/False: Air within a low pressure weather system generally moves toward the center of the low and upward ?
3) True/False: As a general rule of thumb:-wind speeds decrease with increasing height in the lower atmosphere?
4. What is the windiest season in the Presidential Range of New Hampshire: a) summer
   b) autumn c) winter d) spring?
5. Cloud-to-ground lightning has the highest frequency of occurrence between the hours of:
   a) 10 am-2 pm b) 3pm-7pm c) 10pm-1am d) 1am-4am
6. True/False: Wave clouds and a mountain cloud cap indicate high winds near the summit of a mountain?
7. True/False: Due to mixing of air in the atmosphere, a climber at 5000 m (16,400 ft) in the Alaska Range experiences roughly the same air temperature as a climber at the same elevation in the Himalayas?

8 Large thunderstorms typically develop over what time period: a) 14 hours b) 9 hours
c) 6 hours d) 1-2 hours?
9. During the summer, air temperatures _____ as a major low pressure storm approaches:
a) stay about the same b) cool down c) warm up ?
10. True/False: Most ‘ground blizzards’ occur <u>after</u> new snow has fallen?
11. On a night with no clouds and little wind, pick the location that will have the coldest morning temperature: a) top of a ridge b) half way up a ridge c) floor of a valley ?
12. True/False: Precipitation (rain or snow) always increases with increasing elevation?
13. True/False: The primary climbing seasons in Ecuador are May-September and January?
14. True/False: Climate statistics are not useful in expedition planning since the weather on any given day can be dramatically different then the long-term normals?
15. True/False: Water in the atmosphere <u>always</u> freezes when the air temperature is at or below 0º C (32º F)?
16. A large cumulus cloud generates the following types of ‘wind’: a) updraft b) downdraft c) horizontal d) all of the above ?
17. True/False: Wind chill temperatures increase with decreasing wind speeds?

Answers can be found in Appendix 1.

1

KILLER STORMS

Chapter Highlights:

- ✓ Analysis of two real storms that killed a large number of climbers.
- ✓ Learn the indicators of a developing/dissipating mountain storm.

This chapter illustrates the nasty side of mountain weather: real weather experienced by real climbers, no artificial ingredients added. Even though you may never set foot in the Himalayas or climb in the Andes, chances are good that sooner or later you will experience life threatening weather in your own local mountains, in which case the same principles apply. While reading this chapter, ask yourself: "what would I have done given these same circumstances?" Some readers will not understand the technical terms that are introduced in this chapter, do not despair, by the time you finish this book you will know them all too well! There are two reasons that we have started this book off with this chapter: first, to make the reader aware that even very experienced climbers make serious mistakes regarding the dangers of mountain weather, and secondly; to illustrate the point that what would be considered relatively moderate weather at lower elevations, can be deadly at higher elevations.

Everest: May 1996

In terms of world-wide media coverage, this event is "the mother of all" climbing tragedies to date. During the storm of May 10-12, nine climbers of various abilities died. These deaths can be attributed in large part to the adverse weather which embroiled the mountain beginning on May 10th. The specific climbing narratives that we have used are as follows: Into Thin Air (Krakauer 1997); The Other Side of Everest (Dickinson 1999); The Climb (Boukreev and DeWalt 1999); and High Exposure (Breashers 1999).

Since Krakauer has the most descriptive weather narrative, we have used his account to construct a May 10th weather timeline which is displayed as Figure 1.1. We know that strong winds raked the upper mountain throughout April and the early part of May. On May 9th for example, Breashers who was at Camp II in the Western Cwm at the time, reported strong winds above the South Col as evidenced by snow plumes on the upper mountain. Krakauer, who with a large number of other climbers, reached the South Col (Camp IV at 7930 m) on the evening of the 9th. He notes in his book that the winds died down around 7:30 PM. As the climbing teams assembled for their

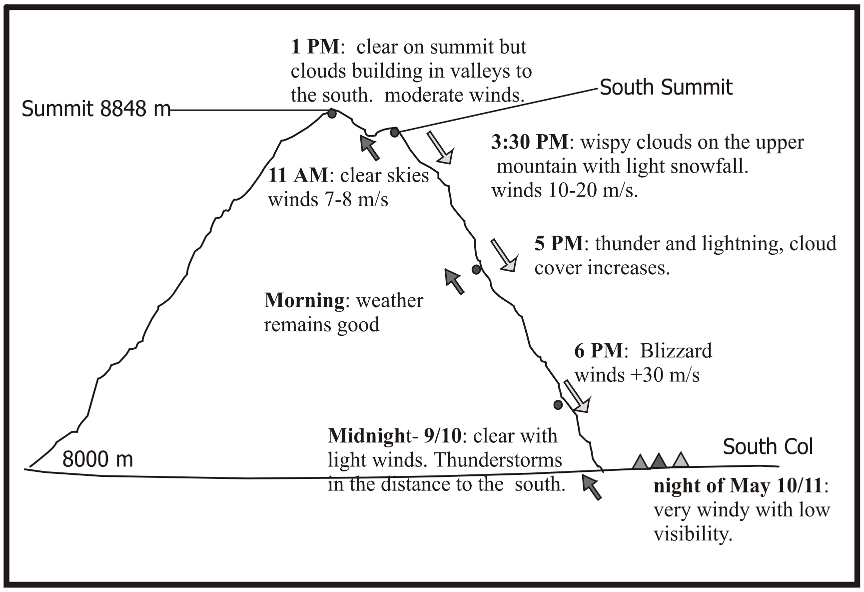


Figure 1.1- Mt. Everest climb May 10-11, 1996.

summit attempt sometime around midnight on the 9th/10th, thunderstorms could be seen over distant mountains to south. Let's pause here to ask an important question. Should the distant thunderstorms have been a warning of an imminent change in the weather? Not really, as you will learn later, pre-monsoon thunderstorms are common over northern India and southern Nepal.

The weather during the morning of the 10th was near perfect for an Everest summit climb, clear skies with light winds. Krakauer reached the South Summit around 11 AM, at which time he estimates the winds were roughly 10 m/s (22 mph). This wind speed is mild for Everest standards, however, it was strong enough to produce a snow plume over the upper Kangshung Face (east). Krakauer reached the summit around 1 PM. It was during his short stay on the world's highest point that he observed a rapid buildup of clouds in the valleys to the south and around Everest, while the region to the north (Tibet) remained relatively cloud free. The fact that clouds preferentially build over Nepal and not over Tibet simply means that moisture moves into the region from the south and southwest. Therefore, if you happen to be at high altitude in the Himalaya, don't bother looking into Tibet for some indication of cloud development, most of the action will be in your immediate surroundings and to the south.

When small cumulus clouds form over valleys and at low-elevations, especially when it occurs on a daily basis, it is generally not an indicator that a major storm is developing. However, the nature of the quick transition in the weather which occurred on the afternoon of May 10th, was a clear sign that a storm of considerable magnitude was developing. The tip-off was the rapid development of large cumulus clouds. In Chapter 4 you will learn that clouds that grow vertically (large cumulus clouds), are typically associated with strong updrafts. The speed and height to which these clouds developed, signified that a large amount of moisture was being "pumped" into the middle and upper troposphere.

By 3:30 P.M. Krakauer had returned to the South Summit where the winds were between 10 and 20 m/s (22 to 45 mph) and light snow was beginning to fall. As he descended the Southeast Ridge, the visibility continued to decrease while the winds steadily increased. Sometime around 4 PM, David Breashers who was at Camp II, noted that a mass of clouds was building to the west of Everest. Shortly thereafter, this cloud bank moved into the Western Cwm and obscured the upper mountain from his view.

It was nearly 6 P.M. when Krakauer reached the lower section of the Southeast Ridge, where he estimates the winds were around 32 m/s (70 mph). He was able to reach the tents on the South Col by 6:45 P.M, at which time a blizzard was raging and visibility was reduced to about 6 m

(20 feet). The winds remained strong throughout the night. At dawn on the morning of the 11th, the mountain was free of clouds but the winds were still very strong. By late morning, however, clouds once again engulfed the upper mountain.

There was a period of several hours in the early afternoon of the 11th, in which the winds decreased to about 20 m/s (45 mph). But by late afternoon the storm re-intensifies, resulting in some very strong gusts of winds (unknown speeds). Blizzard conditions continued throughout the night. By the morning of the 12th, the storm had abated enough for the survivors to descend into the Western Cwm. They arrived at Camp II around 1:30 PM amidst sunshine and light winds.

Figure 1.2 shows the 300 mb height field (in meters) for midnight on May 11 (if you don't know what this plot is showing-no fear, you soon will). Notice the trough of low pressure over northern India, extending well into Central Asia. This weather pattern produced strong westerly winds in the middle and upper troposphere (5-12 km or 3-7 mi). The area of strongest winds remained over northwest India for the duration of the storm, which was fortunate for the Everest climbers, since sustained winds exceeded 50 m/s (110 mph). Although not shown in any of the plots displayed in this section, the upper-level winds over the Khumbu region were quite strong before, and several days after the May 10-12 storm as well.

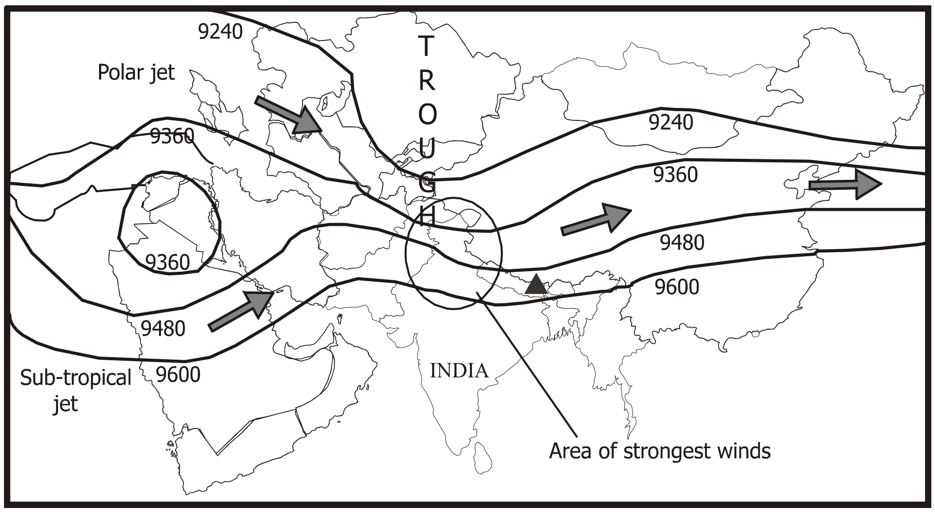


Figure 1.2- 300 mb geopotential heights for midnight May 11, 1996. Mt. Everest is denoted by a solid triangle. The strongest winds are over northern India (Kashmir).

Table 1.1 is an estimate of the actual and wind chill temperatures at several elevations during the storm, based on available weather data. It was obviously very cold on the South Col and above. The wind chill temperatures give some indication of the "apparent temperature" that exposed skin would be subject to, in a wind of 18 m/s (40 mph). In this extremely cold environment there was obviously little margin for error.

Table 1.1

Elevation (m)	Air Temp (C)	Wind Chill (C)
7,000	-19° (-2 F)	-49° (-56 F)
8,000 (South Col)	-27° (-17 F)	-62° (-80 F)
8,848 (Summit)	-33° (-27 F)	-71° (-96 F)

The next three paragraphs are some of our own comments on what transpired from May 10-12. First of all, the speed at which clouds developed on the afternoon of May 10 was an indicator that a major change in the weather was occurring. Cloud development, plus the increasing wind speeds should have been the signal to the guides to turn everyone around during the early afternoon and descend to the South Col. Since the winds over the summit of Everest were strong before and after May 10-12, you're probably wondering what produced the storm. The key element is the trough depicted in figure 1.2. In this book you will learn that there is strong large-scale vertical motion to the east of a trough axis (area of low pressure), which, if it happens to be coupled with a supply of moisture, will produce a very large mass of clouds and precipitation. In addition, the New Delhi (India) upper-air sounding shows a considerable increase in the moisture content in the 3 to 5 km (1.9 to 3.1 mile) layer in the early hours of May 10. Additional data also shows periods of modest southwest flow at 850 mb (1,400 m or 4,590 ft) into Nepal on May 10 and 11. In other words, despite strong winds before and after the storm, it was the development of the trough that produced strong vertical lifting, and allowed for the transport of moisture into Nepal.

Secondly, the clearing that occurred on the morning of the 11th, with little reduction in wind speed, is typical of high-mountain environments. The bulk of moisture in the troposphere is not transported at jet-stream levels. During the warmer months of the year, the primary source of moisture in the Himalayas comes from the Bay of Bengal, with some moisture transported from the Arabian Sea as well. The important point is that just because the clouds dissipated on the morning of May 11, it didn't mean the storm had ended. In this particular case, moisture was moving into the region in a series of impulses. In this type of situation it's very easy for climbers to get 'suckered' into thinking that the storm is over, when in reality it's only in a lull. Trying to determine whether a storm is truly over is a very tough call while hanging on the side of a mountain. Even a seasoned meteorologist, without the help of computer models and weather observations from surrounding regions, would have a difficult time deciding whether or not the storm was indeed over. The primary indicators that a storm like this is dissipating are: a major decrease in cloud cover not only on the mountain itself, but in the surrounding areas as well, and a <u>significant and sustained decrease in wind speeds</u>.

Thirdly, Krakauer's account of the descent from the South Col to Camp II on May 12 clearly illustrates how the winds in the Western Cwm are no indication of wind speeds on the upper third of the mountain. The upper third of a mountain like Everest is considerably windier than the lower two-thirds for two reasons: first, the upper third is closer to the jet stream; and second, because the lower two-thirds is sheltered, in large part, by the surrounding higher terrain of the Lhotse-Nuptse ridge. The South Col is a natural wind tunnel for winds from the west to northwest, as air is squeezed between Everest and Lhotse. No one knows how much <u>wind acceleration</u> occurs on the South Col, but at times it's probably substantial.

In terms of the severity of the May 10-12 storm, from a meteorological perspective, it could have been considerably worse. In fact, according to Everest storm standards, we would have to speculate that this was probably an "average" storm. This could have been a worse storm in several ways. The winds were strong, but they could have been considerably stronger. For example, if the jet streak (area of very strong winds) located over northern India/Pakistan, had moved over Nepal on the

11th, there probably would have been fewer survivors. In addition, the blizzard conditions and associated low visibility could have been continuous. The clear skies on the mornings of the 11th and 12th certainly were a much-needed break for the survivors who were on the South Col.

K2: August 1986

During the 1986 climbing season on K2 (8611 m), 13 climbers were killed, five of whom died while high up on the mountain during the storm of August 4-10. This event differs from Everest in 1996 in two respects: the climbers on K2 all had considerable high-altitude experience; and the storm that began on August 4 continued, except for several short periods, almost totally unabated for 6 days.

For a description of this event I have relied on the narrative of Jim Curran in his book- K2: Triumph and Tragedy (1987). I have also to some degree used Kurt Diemberger's-: K2, Mountain of Dreams and Destiny (found in- The Endless Knot. 1999).

On August 3, the day before the storm began, a team of Korean climbers reached the summit of K2. In the course of their summit climb, three meteorological events were occurring: first, wispy clouds were starting to form around the summit pyramid; second, wind speeds were increasing, and finally, a cloud cap was forming over neighboring Broad Peak (8010 m). The wispy cloud forming around the summit in and of itself is not much of an indicator of changing weather. The modest increase in wind speeds, is important, especially when it occurs in conjunction with additional factors. The developing cloud cap over Broad Peak signifies that the moisture content of the atmosphere was increasing.

These three factors in combination, should have been an indicator to those high on the mountain that the weather was probably going to deteriorate. The questions that the climbers should have been asking themselves were: if a storm develops, how long will it last, and how adverse will the conditions get? No one has ever studied the frequency or duration of major storms in the Karakoram or Himalayas. But most climbers, based on previous experience, know that a storm of moderate intensity generally lasts two to three days. Over the course of the climbing season however, it's not uncommon for these giant peaks to get hit by several storms that can last for a week. As it turned out, the August 4-10 storm was one of those where the only safe place to be was base camp.

Figure 1.3 shows the 300-mb height field (in meters) during the early days of the storm. A trough of low pressure had formed over the Caspian and Aral Seas around August 2. This trough intensified and moved south, reaching the Karakoram on the night of the 4th/5th. Over the next week, this system slowly moved to the

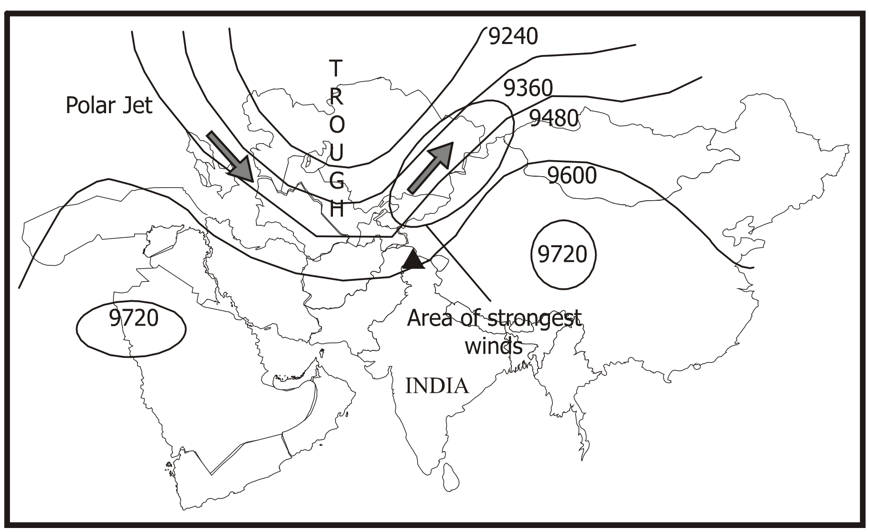


Figure 1.3- 300 mb geopotential heights for August 6, 1986. K2 is denoted by a solid triangle. The strongest winds are over Central Asia.

east. The area of strongest winds associated with the trough was located to the northeast of the Karakoram for most of the event. For example, the 200 mb (11,500 m or 37,700 ft) wind speeds exceeded 50 m/s (110 mph). By the 11th, the trough had almost completely dissipated, but strong westerly winds continued to blow over much of central Asia.

Middle and upper-level winds over the Karakoram are displayed in Figure 1.4 (this data is from standard meteorological analysis). Notice the weak winds at both 5800 m (19,000 ft) and 7500 m (24,600 ft). Jim Curran, who was at Base Camp (5500 m) during the storm, indicates in his book that the winds were periodically much stronger than what is shown in figure 1.4. Likewise, at higher elevations the winds were no doubt considerably stronger than what is depicted. Temperatures at 8000 m (26,000 ft) during this storm were about -18°C (0°F). Combined with a wind of 18 m/s (40 mph), that would produce a wind chill in the neighborhood of -47°C (-53°F).

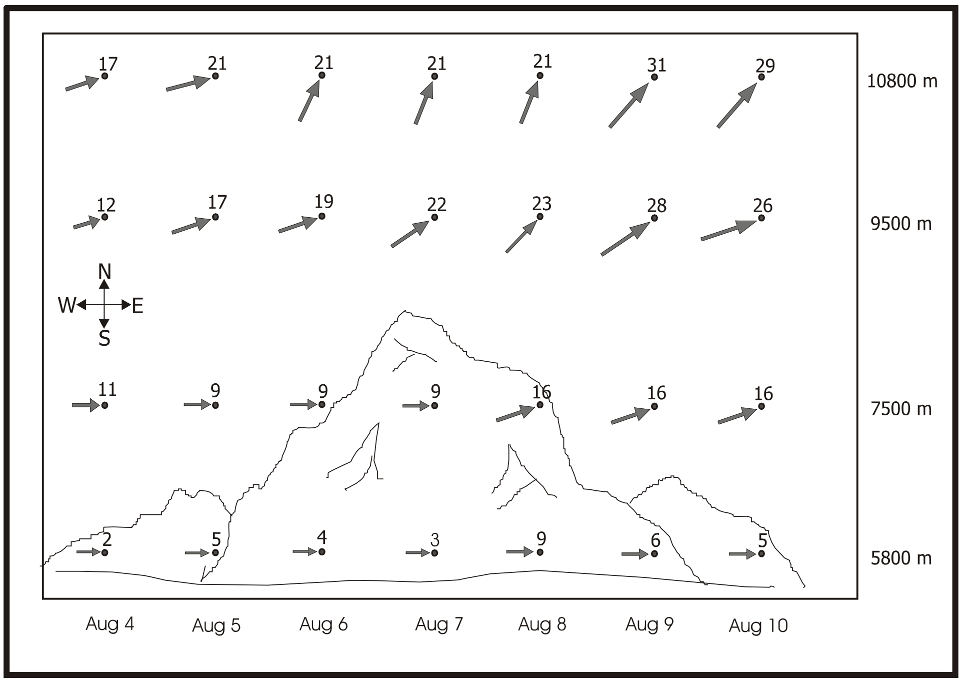


Figure 1.4- Winds over the Karakorum region August 4-10, 1986. Arrows indicate wind direction while numbers show wind speed (m/s).

Storm Awareness

The following is a list of weather parameters that you can use to monitor the evolution of mountain storms. Notice that these guidelines are pretty general, for example; “substantial increase in wind speed”. The reason for this is because conditions vary from storm-to-storm. Remember that meteorology is not an exact science. In some situations a 10 m/s (22 mph) increase in wind speed may be significant, in other cases it may not. Also, changes in one or two of these parameters may occur without the development of a storm. However, when a number of them change, beware! Where appropriate we have given some rough indication of the duration in which these parameters should change.

Pre-storm

- Drop in barometric pressure [rising altimeter] (> 4 hours)
- During the summer months, large storms transport cooler air-therefore a drop in temperature of 4º C (7º F) or more over a period of 6 to 12 hours is significant.
- Thick layer of clouds on horizon. Usually in the direction of the prevailing wind.
- Formation of cloud caps and lenticular clouds.
- The height of the cloud base continues to lower.

Storm Development

- Substantial increase in wind speed.
- Significant change in wind direction-may or may not occur. When a front moves across a region winds will frequently change direction. (< 2 hours)
- Rapid build-up of clouds. (1-2 hours)

Diminishing Storm

- An extended period of decreased winds, including a reduction in the frequency and intensity of wind gusts. (> 4 hours)
- Depletion of moisture- this will result in a reduction in cloud cover not only in your immediate surrounding, but across the region as well.
- Increase in barometric pressure [falling altimeter]- this indicates that the storm is weakening or moving out of the region. (> 4 hours)

2

METEOROLOGY 101

Chapter Highlights:

- ✓ Meet the earth's atmosphere: the weather maker.
- ✓ Learn why oxygen decreases with elevation.
- ✓ Weather fronts defined.
- ✓ How to use an altimeter for 'on-the-mountain' forecasting.

Earth's Atmosphere: The Basics

We will start our study of mountain weather by looking at the structure of earth's atmosphere. This is important because the phenomenon we call weather, is caused by simple changes in the condition of the atmosphere: that is changes in temperature, air pressure, moisture, and wind. The atmosphere is of course the layer of air that surrounds the earth which extends up to an altitude of about 800 km (500 miles).

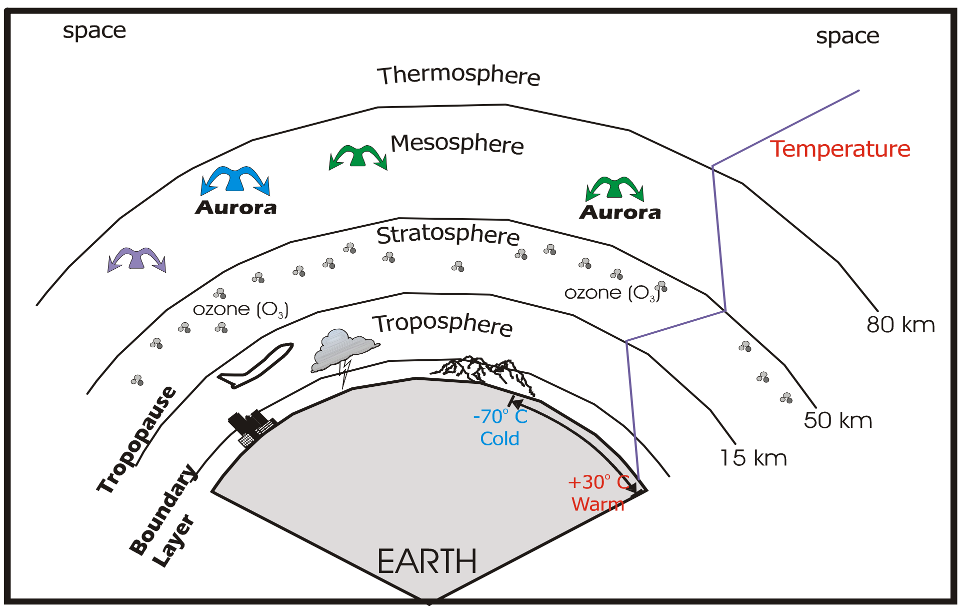


Figure 2.1- Schematic of the earth's atmosphere.

Meteorologists like to separate the atmosphere into four sub-regions, from the earth's surface upward they are the: troposphere, stratosphere, mesosphere, and the thermosphere (figure 2.1). This classification is based primarily on the change in temperature with height within each region.

The gas that we call 'air' is made up of 78% nitrogen, 21% oxygen, and 1% of trace gases like argon, carbon dioxide, water vapor, and helium to name a few. It's important to note that the percentage of these gases remains fixed throughout the entire troposphere and lower stratosphere. For example, the air on the summit of Mt. Rainier (4,392 m) or K2 (8,613 m) still contains 78% nitrogen, 21% oxygen, and 1% trace gases. What does change however is the density of air (Figure 2.2). Air density decreases as elevation increases. As a result, there is less air and hence less available oxygen for a climber to breathe as they ascend. In practical terms this means that on the summit of Mt. Rainier, a climber has approximately 57% of the oxygen available at sea-level, and on the summit of K2, only about 32%.

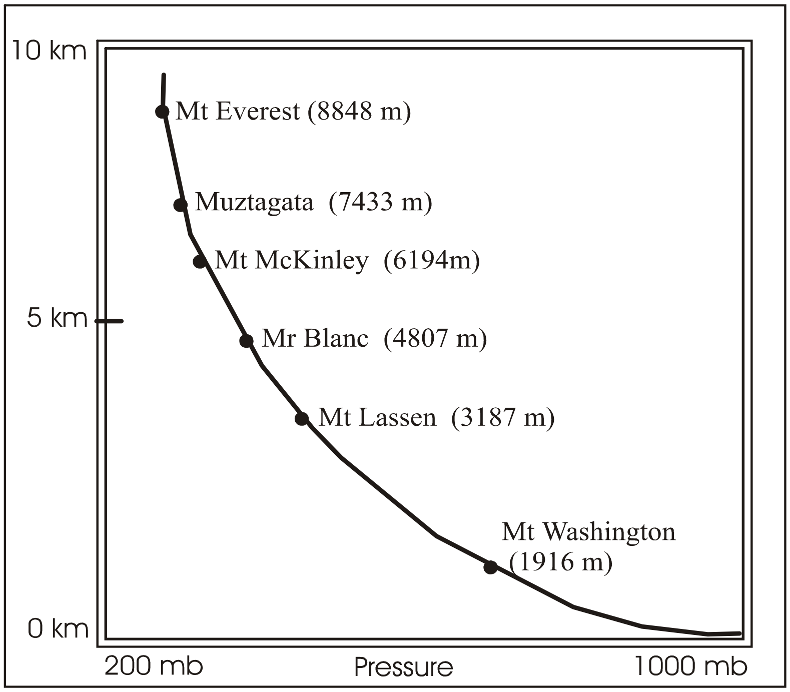


Figure 2.2- Change in pressure with change in elevation.

Troposphere:

The region that interests us the most is the lowest one, the troposphere; because it's only here that significant amounts of moisture and hence clouds are found. Above 16 km (10 miles) the air is too cold and too thin to contain very much moisture, so the weather that affects the earth is restricted primarily to the troposphere.

For the most part the troposphere is heated from the ground up. Since the surface of the earth consists of water, rock, dirt, vegetation, snow, etc. the air above these different materials can have widely differing temperature and moisture contents. This in turn leads to the generation of what we call *air masses* , very large-slowly moving-masses of air that have different temperatures, densities, and moisture. Polar air masses for example are cold and dry, while air masses that originate over the tropical oceans are warm and moist.

Since the troposphere is heated from the ground up, air temperatures within the troposphere generally decrease with height. The rate at which it decreases varies from about 4º C to about 10º C per kilometer (11º to 18º F per mile), and is termed the *lapse rate*. As you might imagine, the lapse rate at a given location changes as air masses pass through the area and as the air next to the ground heats up during the day and cools off at night.

Air masses move around and frequently collide with each other. Since cold air is denser than warm air, the colder air mass slides beneath the warmer air mass–with interesting results, as we shall shortly see. It's in regions of interacting air masses that most of our planet's stormy weather occurs.

It takes considerable energy to heat large masses of air, of course this energy comes from the sun. Considerably more sunlight reaches the tropics than anywhere else on the planet. In fact, if it weren't for the movement of air masses, the tropics would be considerably hotter than they already are, while the higher latitudes would be much cooler. Keep in mind that ocean currents also help distribute the heat from the tropics to higher latitudes. Nevertheless, a large temperature difference exists between the tropics and the mid-latitudes, which in turn produces an equator-to-mid latitude movement of air known as the Hadley Cell (Figure 2.3). The Hadley Cell sort of works like an air conveyor belt-heat energy from the tropics is transport to the mid-latitudes of both the northern and southern hemispheres.

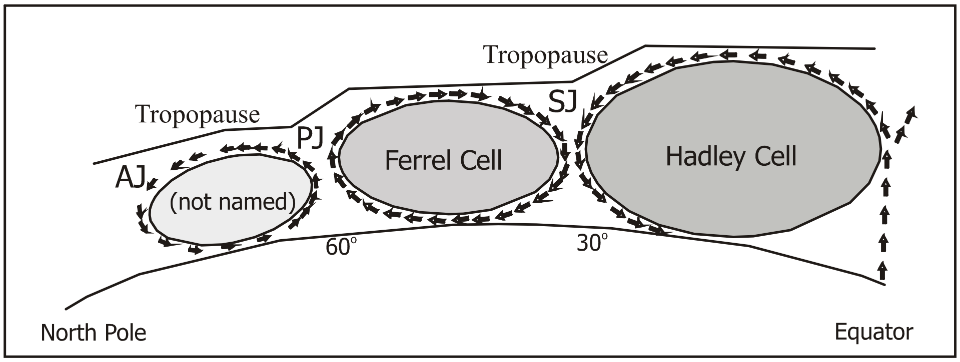


Figure 2.3- north-to-south cross section of hemispheric circulation systems.
AJ= Arctic Jet, PJ= Polar Jet, SJ= Sub-tropical Jet.

At the top of the troposphere, and dividing it from the stratosphere is a region called the tropopause. It's not a distinct boundary, but rather a transition zone. The tropopause is not located at a constant height over the entire globe; it varies with latitude and from season-to-season. The height of the tropopause ranges from about 17 km (10.6 miles) in the tropics to about 8 km (5 miles) in the polar regions. Although we won't be discussing jet-stream winds until a later chapter, it's worth noting that <u>some of the strongest winds found anywhere in the lower atmosphere occur in the vicinity of the tropopause</u>. As a result, a general rule-of-thumb is as follows: the higher you climb the more likely you are to encounter strong winds. In addition, taller mountains located at higher latitudes tend to experience stronger winds then a mountain of similar height located further south.

<u>Stratosphere</u>:

The stratosphere differs from the troposphere in a very significant manner; in this region temperatures increase with height. In addition, the stratosphere contains a significant amount of ozone that absorbs large amounts of the sun's ultraviolet radiation and so protects plants and animals from these harmful rays. The elevated warm layer of air above the troposphere is important because it acts as a lid on how much energy can be transported from the earth's surface to outer space. There is very little moisture in the stratosphere, but thin ice clouds do form from time-to-time. In short, weather in the stratosphere has little direct effect on mountain weather.

<u>Mesosphere and Thermosphere</u>:

These two regions are the domain of high-energy particles arriving via the solar wind, that is particles that travel from the outer layers of the sun. This is where the aurora borealis (Northern Lights) and the aurora australis (Southern Lights) are produced. Air in these regions has such a low density that virtually no weather is generated here.

Weather Fronts

In the lower troposphere, two neighboring air masses often have large differences in temperature. These pronounced contrast in temperatures (the technical term is- *temperature gradient*) are found along the boundaries between the two air masses. It's in these boundary zones, which we call *fronts*, that one air mass tends to be lifted over the other (Figure 2.4). This is important because rising air is essential for the development of clouds and rain or snow. The rate or speed of frontal lifting is small when compared to an air mass's horizontal speed, however it's still enough to generate clouds over large areas of the earth's surface.

Most fronts vary in length from about 500 km (300 miles) to over 1000 km (600 miles), while the width is a more modest 50-100 km (30-60 miles). A front is recognized as a change in wind direction, temperature, moisture, or sea-level pressure from one side of the front to the other. There are four types of fronts: cold, warm, stationary and occluded. Fronts typically form in conjunction with a developing storm (i.e. low-pressure system simply called a 'low'). They are named according to the direction in which the coldest air is moving with respect to the frontal boundary. It is important to note that 'cold' air and 'warm' air are relative terms. The important point is that there is a significant temperature difference between the two air masses. There is no criteria that says that 'cold' air has to be below a certain temperature, or that 'warm' air has to be warmer than a specified temperature.

A cold front is produced when cold air overtakes and slides underneath warm air which is ahead of it. A warm front occurs when warm air overtakes cold air, but in this case the warmer air moves over the top of the cooler air. Remember that cold air clings to the earth's surface because it's denser and therefore heavier then warm air. Sometimes warm and cold air move parallel to the frontal boundary, forming what we call a stationary front. As the name implies, there's very little relative movement between the two air masses. The fourth and final type is an occluded front. In this case cold air moves counterclockwise around the center of a mature storm, effectively undermining all of the warmer air that used to be near the ground.

A mature storm can have more than one front, in fact most have a cold and a warm front. The warm front is usually positioned several hundred kilometers to the east of the cold front. The four types of fronts we've just described occur most often during the winter months, in the middle and high latitudes. In the tropics, airmass fronts

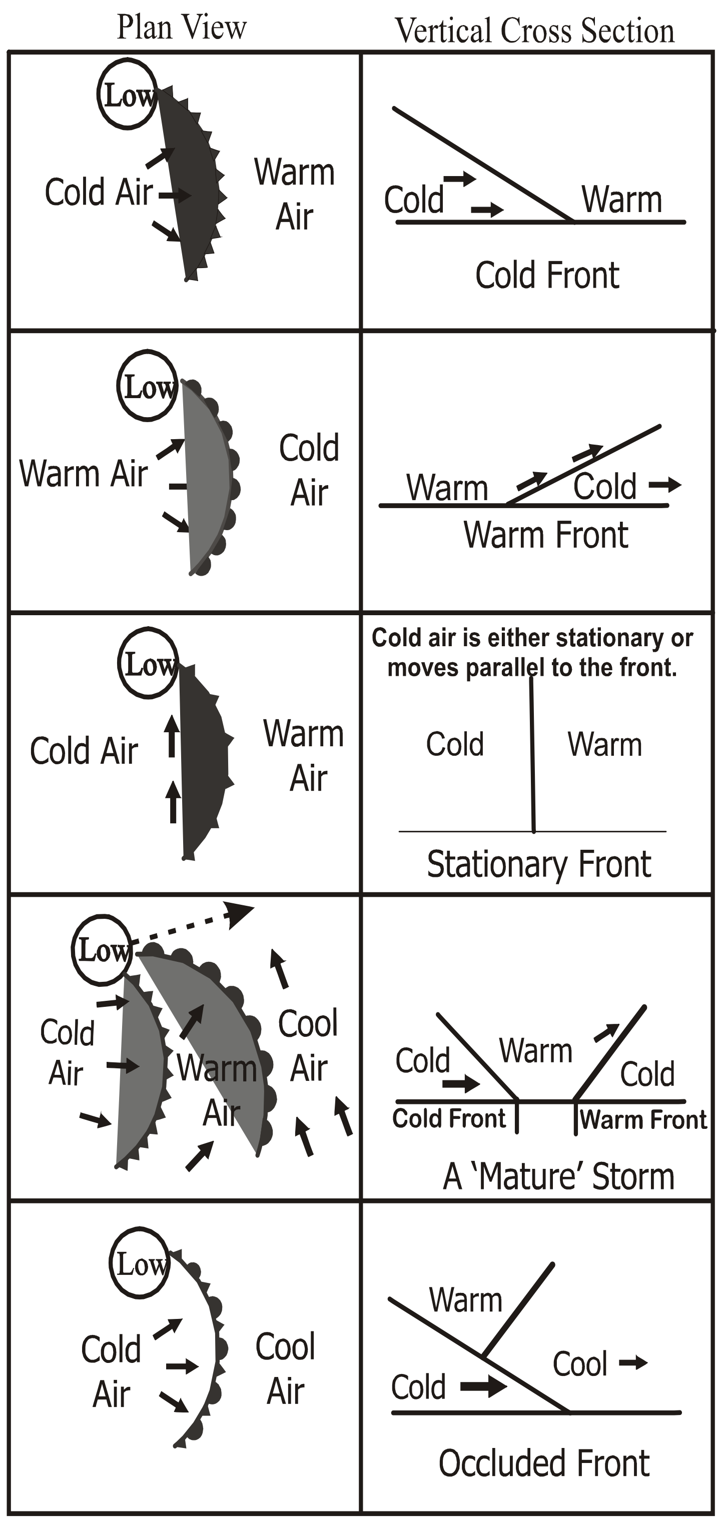


Figure 2.4- Surface Fronts.

do not often develop because there is very little temperature contrast across the uniformly warm tropical oceans.

Forces that Produce Weather

If you really want to understand how weather is generated, then you need to know what natural forces produce it. This topic can be a bit daunting to those readers who have never taken a high school or college level physics class, nevertheless, read through this section, you might surprise yourself with what you can learn.

The most easily identified force at work in the atmosphere is gravity. Here is an example how gravity helps produce weather. Imagine a parcel of air (some identifiable unit of air, suppose one meter on a side) located 3 km (1.9 miles) above the ground. This parcel has no horizontal or vertical motion and it has the same temperature as the surrounding air. Will the force of gravity cause this parcel to sink toward the ground? The answer is : no! In this scenario the downward pull of gravity is equally matched by an upward pressure force; the parcel will in fact remain stationary. However, if the parcel is warmer than the surround air, its density will be lower, and it will tend to rise. If on the other hand, the parcel is cooler then the surrounding air, it's density is higher and it will tend to sink toward the ground.

<u>Remember that warm light air rises; while cold heavy air sinks</u>. This turns out to be very important in the formation of clouds. For example, on a warm sunny day, air just above a large patch of bare dirt or asphalt will be heated to a temperature well above that of the surrounding air. Since it's warmer and hence lighter, the air above the dirt begins to rise above the ground; this process is called *convection*. A rising parcel of air which is heated in such a manner is often referred to as a *thermal.* As this warm parcel rises it eventually reaches an altitude where it's density matches the density of the surrounding air, at which time it comes to rest.

As you gain elevation in the mountains for example, atmospheric pressure– the weight of air above your head–decreases, as the density of air decreases. A good analogy is the pressure that a swimmer experiences on a dive to the bottom of a swimming pool. The water pressure at any given depth in the pool is the total weight of the water pressing down on the swimmer; the deeper the swimmer, the higher the pressure.

At sea-level, air pressure is on average around 1013 millibars (mb). At an altitude of 5.5 km (3.4 miles) it's only about half that. Table 2.1 shows the atmospheric pressure that a climber would experience while on the summit of a few selected mountains. Note that the decrease in pressure with increasing elevation is not linear (would not be a straight line if these values were plotted on a piece of paper and a line segment connected each value). The vertical distance between the 1000 mb and 900 mb levels is about 900 m (2,950 ft), while the distance between 500 mb and 400 mb is about 1600 m (5,250 ft). Incidentally, meteorologist usually measure vertical distances in the atmosphere in meters or kilometers. If they're referring to a specific level then they give the vertical coordinate in terms of the pressure level (i.e. temperature at 500 mb, or winds at 900 mb). You are probably wondering what millibars are- they are a metric unit of pressure (force per unit area), similar to 'pounds per square inch'.

Let's return to our discussion on important forces that are at work in the atmosphere. Two additional forces that should be mentioned are the *Coriolis force* and the horizontal pressure gradient force (the word horizontal is usually dropped since we know by the context that it's not the vertical pressure gradient that's being referred to). The Coriolis force is named after G.G. de Coriolis, a 19th-century French mathematician who discovered it. The Coriolis force causes a horizontally moving parcel of air to change direction (deflection) without any change in it's speed. In the Northern Hemisphere this deflection causes air move to the right of its original course, but in the Southern Hemisphere deflection is to the left. For example, a wind that is blowing from west-to-east across the USA will be deflected to the south. Likewise a wind that travels from the Gulf of Mexico toward South Dakota will be deflected toward the Ohio Valley.

TABLE 2.1- Pressure and percent oxygen for selected mountain summits. Oxygen values based on sea-level being 100%.

Mountain	Height (m)	Pressure (mb)	% Oxygen
(sea-level)	0	1013	100
Mt. Washington- USA	1916	805	79
Mt. Lassen- USA	3187	685	68
Mt. Rainier- USA	4392	580	57
Mt. Blanc- Fr/Sw	4807	555	55
Mt Foraker- USA	5302	520	51
Mt. Logan- Canada	5951	475	47
Mt. McKinley-USA	6194	455	45
Aconcagua- Chile	6958	410	40
Muztagata- China	7433	385	38
Nanga Parbat- Pakistan	8126	350	35
Mt. Everest- Nepal/China	8848	310	31

The pressure gradient force is not hard to understand. In Figure 2.5 two columns of air are displayed. In column **A**, the average temperature is 10 degrees colder than the air in column **B** (the actual temperatures and heights are not important). Since both columns stretch from the surface to a height of 20 km (12.2 miles), column **A** will have a higher average density than the air in column **B**. As a result, the pressure at any given height in column A will higher then the pressure at the same height in column **B** (this example falls apart near the top of the columns). If pressure were the only force involved, air would want to move from high pressure (**A**) to lower pressure (**B**). In

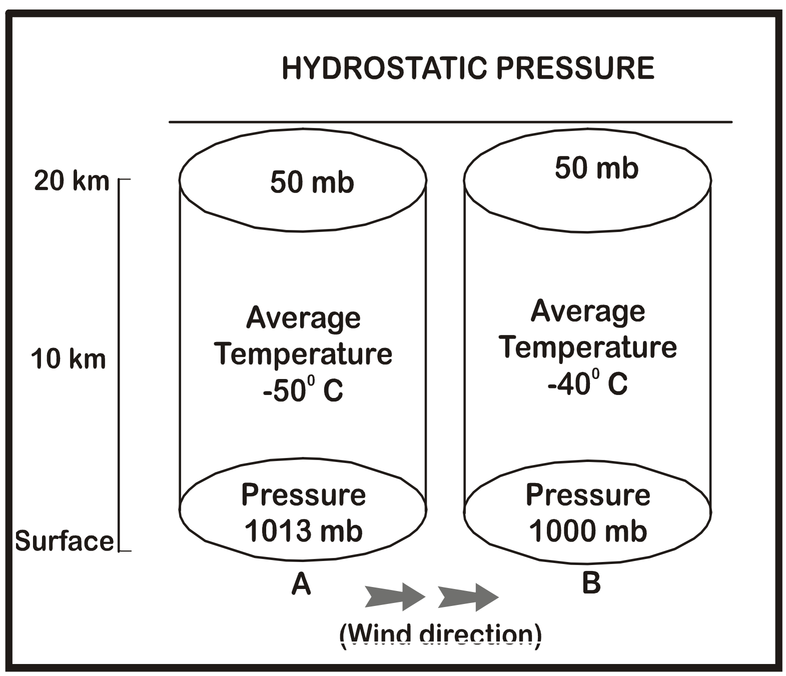
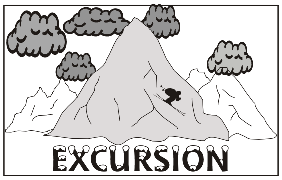


Figure 2.5- Pressure is a function of temperature.

reality however, since there are always additional forces involved, air does not move (wind!) directly from **A** to **B**, it almost always moves in a curved path.



Altimeters and Changing Pressure

Everyone knows that an altimeter measures changes in elevation. But what you may not know is that altimeters actually measure atmospheric pressure, which is then converted into a elevation for display. Because of this property, an barometric altimeter can be used to monitor changes in atmospheric pressure. For example, a 10 m (33 ft) change in elevation corresponds to a pressure change of about 1 mb. But make sure that you understand that pressure change is <u>inversely</u> related to changes in elevation. <u>When the reading on the altimeter decreases (when your descending for example) the pressure is rising, and when you ascend, the pressure is decreasing.</u> If you want to use an altimeter as a barometer, make sure that the altimeter is not physically changing altitude; in other words, this only works when you have stopped hiking or climbing for the day and the instrument remains at one elevation for period of time.

Here are some simple steps to follow if you want to use your altimeter as a barometer. ***Step One:*** *Read your altimeter when you first stop and set-up camp, you may want to write the value down.* ***Step Two:*** *In the morning, compare the current reading with the value from the previous evening. If, for instance, the altimeter <u>rises</u> 60 m (197 ft) during the night, you simply divide 60 by 10 (remember its 10 m or 33 ft per 1 mb). The result would indicate a pressure <u>drop</u> of about 6 mb. So what does this tell you about the weather? Well, during a typical 12-hour period atmospheric pressure will rise or fall 2 or 3 mb (20-30 m or 65-100 ft) on its own. A pressure change of 3 mb or more is required before any real significance can be attributed to it. A pressure drop of 6 mb in 12 hours is respectable, a 10 mb drop over the same period is pretty large. Since most major storm systems develop in association with low-pressure systems, this <u>may</u> indicate the approach of a storm and it's associated fronts. Keep in mind however, that a change in pressure is only one piece of the puzzle.* ***Step Three:*** *Observe additional meteorological elements such as changes in wind speed and wind direction, as well as the current and past cloud types. If the 6 mb decrease in pressure occurs at the same time that clouds are on the increase and the winds are either changing direction or increasing in speed, then you have enough information to deduce that a front is approaching your location. Armed with this information you can make an intelligent decision about whether to continue your climb/hike, remain where you are, or retreat.*

How Winds are Generated

In order to help you understand how winds are generated consider the following scenarios. The first scenario takes place near earth's surface, where winds are a product of three interacting forces: the pressure gradient force, surface friction, and the Coriolis force. The net result of these forces is a surface wind that blows <u>into a region of low pressure</u> and <u>out of an area of high pressure</u>. As Figure 2.6 illustrates, these winds tend to move across isobars (lines of equal pressure that are drawn on a surface weather map), at an angle of about 30 degrees. Recall that the winds do not directly move from an area of low pressure to high pressure, <u>friction and the Coriolis force cause the</u>

wind to have a curved path. Surface friction is a result of air moving over and through obstacles such as trees, mountains, buildings, etc.

The movement of air into an area of low pressure and out of high pressure has important implications for the development and dissipation of a storm. Air that moves into a region of low pressure has to continue in motion, otherwise air in the low would pile up and the pressure would increase. What actually occurs is that air moving into the center of a low is forced to ascend, which in turn produces clouds and precipitation.

Contrast this with regions of high pressure (a high), where air descends from higher levels within the atmosphere. This results in air moving out from the center of a high. Descending air tends to be dry, so few clouds are produced in regions of high pressure. By the way, it's important for you to understand that surface winds are frequently weaker and do not necessarily blow in the same direction as the winds in the middle and upper troposphere.

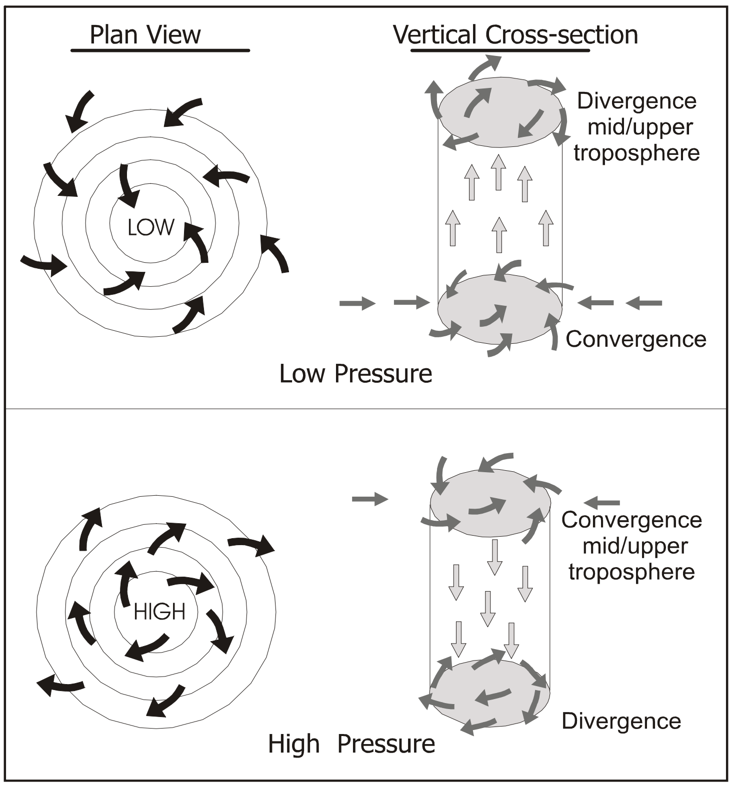


Figure 2.6- Air spirals into a LOW in a counterclockwise direction, while air spirals out of a HIGH in a clockwise direction. In the Southern Hemisphere the direction of rotation is the opposite.

In our second scenario, lets move above the influence of surface friction and into the middle and upper troposphere, where the two most important forces that generate wind are the pressure gradient force and our new friend, the Coriolis force (technically one has to also consider centrifugal forces but we will ignore them for now). Since the pressure gradient is balance by the Coriolis force, the wind moves parallel to the isobars (this is called the geostrophic wind). Above the earth's surface however, we no longer use isobars; we measure the height of a given pressure level above mean sea-level. This sounds ominous, however it is quite easy to use. Since the units are in meters, the term 'heights' is used when we refer to a pressure level.

A three-dimensional example is illustrated in Figure 2.7. You'll notice in this figure that the top of the box represents a constant elevation above sea-level, while the thin solid lines represent the height of the 500 mb pressure level above sea-level. Large heights (5480 m in this example) represent higher pressure, while lower heights (5120 m) represent lower pressure. In other words, the 500 mb pressure level in this example is not parallel with the earth's surface; it's higher in the south, and lower in the northwest. The spacing of the height lines gives an indication of wind speeds. Closely spaced height lines mean that the pressure gradient is large and the resulting wind speeds are high, while height lines that are widely spaced indicate lower speeds. We have added the arrows to indicate which way the winds are blowing. But you already know that wind travels counterclockwise around low pressure systems, and clockwise around highs, hence you won't need the help of arrows in the future. Keep in mind, however, that Figures 2.6 and 2.7 show highly idealized wind systems, although they're pretty accurate approximations in most cases. As you can see from the examples illustrated in this chapter, most large-scale winds follow a wavy (sinusoidal) course as they move

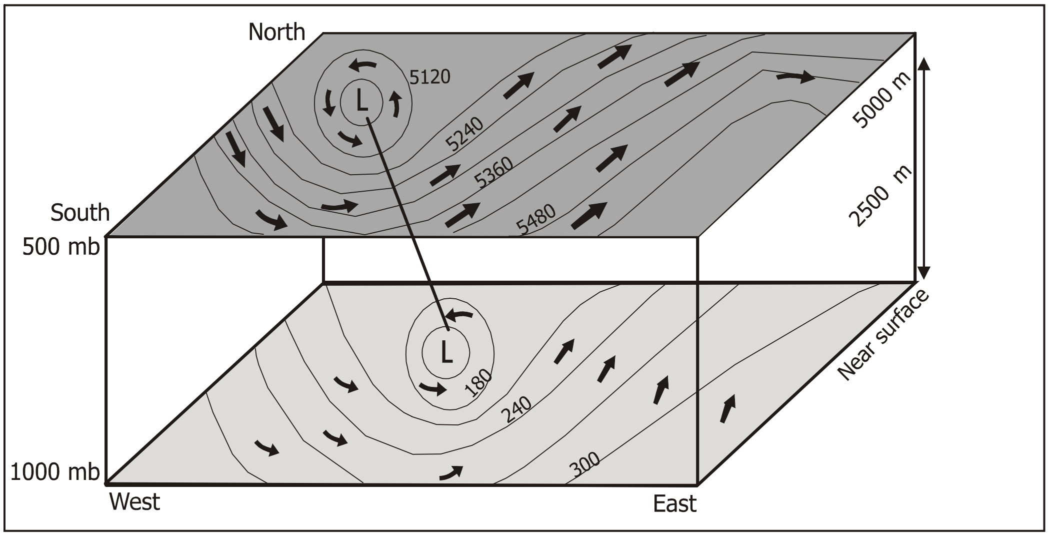


Figure 2.7- Wind vectors (arrows) and geopotential heights (solid lines in meters), for a typical low positioned in the mid/high latitudes. Note that the top pane represents the winds on the 500 mb pressure level, the bottom pane the winds at 1000 mb. The dot-dashed line indicates the western tilt of the low.

around the globe. This turns out to be important because these waves help to transport warm air from the tropics and cold air from the polar regions.

Getting the names right- time for a quick word about nomenclature:

Low-pressure systems (i.e. storms) are called *lows* or *cut-off lows* if the isobars completely encircle the area of lowest pressure, as depicted in the lower panel of Figure 2.7. Sometimes a region of low pressure forms, but the isobars are not closed; these features are called *troughs*. Areas of high pressure are called *highs* or *anticyclones* if the isobars are closed, or *ridges* if not closed. In your reading you may encounter the term cyclonic, which in the Northern Hemisphere refers to the counterclockwise motion around a low or trough, while anticyclonic refers to clockwise motion around highs and ridges. Frequently, a cut-off low will form in the lower troposphere but has the structure of a trough in the middle and upper troposphere. Note that 'tropical cyclone' is another name for a hurricane or typhoon. These storms are deep low-pressure systems that are generally only found in the tropics and sub-tropics.

Pressure Systems: the Highs and the Lows

In the tropics and polar regions, winds predominately blow from east-to-west, these types of winds are called *easterlies*. In the mid-latitudes (30º - 70º), winds blow from west-to-east and are commonly called *westerlies*. In meteorology the wind direction indicates the direction which wind is blowing out-of. Since most highs and lows form in the mid-latitudes, these features tend to move along with the westerly winds. However, there are certain regions on earth that favor the formation of highs and lows, at least during certain times of the year. In winter in the Northern Hemisphere, for example, there are two dominate lows and three dominate highs. The lows are the Aleutian low, which is centered near the Aleutian Islands, and the Icelandic low, which usually southwest of Iceland. These lows won't necessarily show up on the surface weather map every day in winter, but over the course of the whole season they are frequently found near these two areas.

Two of the highs are located over the sub-tropical oceans, centered near latitude 30º N, therefore they are called sub-tropical highs. One of these lies off the coast of California, while the other is located in the central Atlantic, west of the Azores. The third high is located in central Siberia at about 50º N, and is therefore called the Siberian High. It, unlike the sub-tropical highs, is produced by very cold air that is trapped in the lower troposphere, hence it forms during the winter months.

The sub-tropical highs are directly linked to the Hadley Cell circulation. In fact these highs are produced by sinking air that originated near the equator. These large regions of subsiding air tend to be very dry and hence relatively cloud free.

During the summer, the Aleutian and Icelandic lows weaken, while the sub-tropical highs strengthen and move slightly north and west of their winter positions. These semi-permanent lows and highs play a very important role in the overall climate of many regions. Areas to the west of lows tend to be cold and dry during the winter, while areas to the east are typically wet and cool. Regions to the west of the sub-tropical highs are dry and warm while areas to the east are very dry and hot. Its no coincidence that the largest deserts on earth are located to the east of sub-tropical highs.

In later chapters you will read about the weather and climate in various mountain ranges; note how the local weather is, in large part, controlled by its location with respect to one of these features. In the Southern Hemisphere, there are semi-permanent highs and lows as well. For example, the sub-tropical high located off the west coast of South America, is the main reason that most of the central Andes are quite dry.

The Size of Weather

This chapter is going to conclude with what many of you will think is a very dull subject-the scale at which weather occurs. Meteorologists typically use four scale sizes to help classify weather features: planetary, synoptic, mesoscale, and local-scale. First, note that these scales are used as <u>approximate</u> delineations. In other words, scale is relative, not exact. Second, the scale size is usually inversely proportional to the resolution. When you look at weather on the planetary and synoptic scales, your looking at the gross features over a large area. When you deal with the mesoscale and local-scale, your looking at finer details over a limited area.

A good way to understand these scales is to imagine a camera orbiting the earth, which takes photographs with four different lenses (Figure 2.8). The planetary scale would be equivalent to a picture of the earth's atmosphere taken from space with a wide-angle lens. It would capture the largest weather features, but it's not going to resolve thunderstorms developing over the Wasatch Mountains of Utah.

If the wide angle lense is replaced with one with a medium focal-length, the new image would be equivalent to the synoptic scale. This scale has a horizontal dimension of several thousand kilometers. It by the way is the scale used for most weather analysis. Additionally, a map of the continental USA that depicts surface pressure is another example of weather analysis on the synoptic scale.

The mesoscale on the other hand, covers a much smaller region, roughly several hundred kilometers on edge. At this scale, terrain becomes a significant weather factor. A forecaster who is writing a forecast for the northern Cascades, would first analyze the weather on the synoptic scale, then adjust it for the terrain (more on the forecast process in a later chapter).

The smallest scale is the local-scale, or what can also be called the microscale, which covers a distance of 100 km (60 miles) or less on edge. Most forecast for urban areas, where it may affect millions of people are made at this scale. In rural areas, most forecasting is done on the mesoscale. This means that when you read a forecast for a mountainous region, it typically only gives general wind, temperature, and precipitation information. In other words, there is going to be considerable variations in weather across that zone, due to changes in elevation, and rain shadow effects to name a few, which are not typically covered in the forecast.

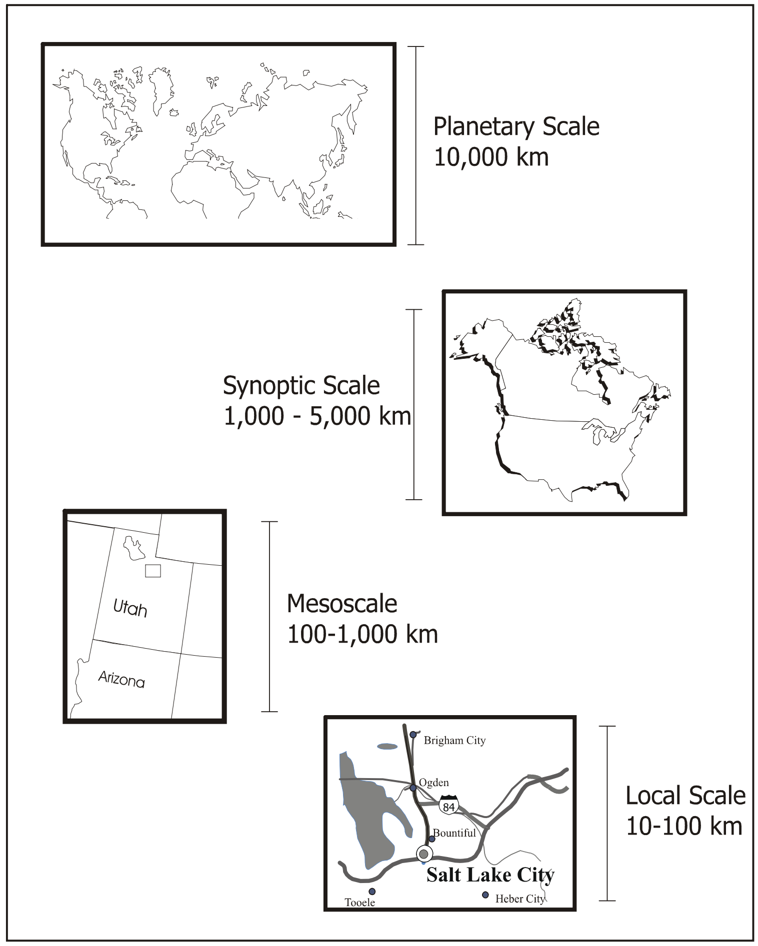


Figure 2.8- The four scales used in meteorology.

The main point of this section has been to point out that weather occurs on a variety of spatial scales. In general, it is easier to forecast on the larger scales then it is on the smaller scales. In addition, it is easier to forecast weather in the middle and upper troposphere then it is near the earth's surface. This is a result of the fact that friction and turbulence to name a few, add a lot of complexity to weather dynamics.

1. Name the four types of fronts: ______, ______, ______, ______.
2. True/False: The Coriolis force only changes wind speed, not wind direction.
3. Oxygen is the: 1st, 2nd, 3rd, 4th most abundant gas found in a sample of air.
4. A rising altimeter signifies what?
5. Air moves in a ________direction around an area of low pressure (Northern Hemisphere).
6. True/False: more oxygen is available to a climber on the summit of K2 than to a climber

on the summit of Mt. McKinley.

7. A cold air mass moves: a) over b) through c) under, a warm air mass.

8. A westerly wind is moving in what direction?

3

RADIATION AND TEMPERATURE

Chapter Highlights:

- ✓ Learn the difference between shortwave and longwave radiation.
- ✓ Excursion- Why is the sky blue?
- ✓ Temperature cycles explained.

There is no free ride in the earth's atmosphere: energy is required to produce weather. As you probably already know, the sun provides the earth with energy that drives the atmospheric circulation system. The sun emits massive amounts of energy in the form of electromagnetic radiation (i.e.- photons), of which the earth receives only a fraction. The energy level of this radiation is directly related to its frequency; the higher the frequency the higher the energy level (Figure 3.1). It turns out that the frequency of energy emitted by the sun is a function of its temperature; higher temperatures produce radiation at higher frequencies. The sun, like any other star consists of a large quantity of very hot hydrogen gas. The temperature of the hydrogen in the sun varies significantly, gas near the core is thousands of times hotter than gas in the outer layers. Therefore the sun does not emit radiation at a single frequency, rather it emits radiation across the entire electromagnetic spectrum. However, since the layer of hydrogen in the sun's outer atmosphere is about 5,000^{o} C, many of the photons it emits are in the visible portion of the spectrum.

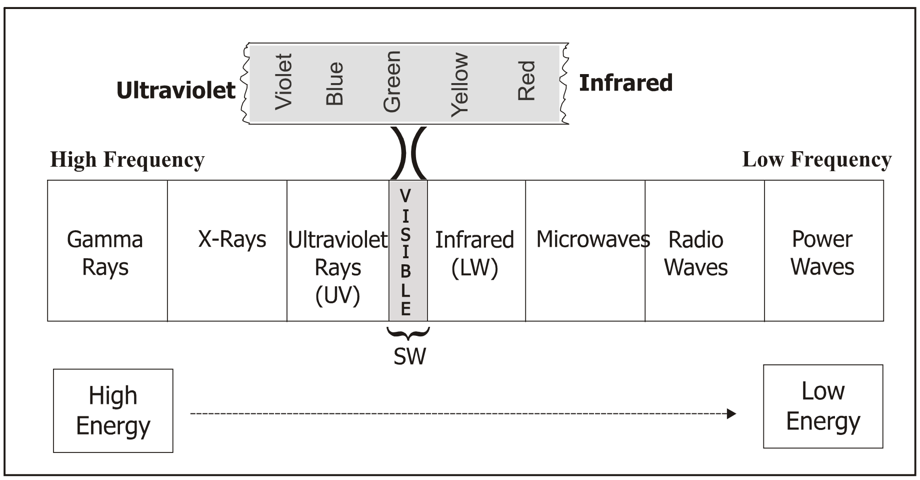


Figure 3.1- The electromagnetic spectrum. The human eye only sees the visible portion of the spectrum. Light appears white because it contains equal amounts of all colors.

Shortwave Radiation

In meteorology we use the terms shortwave (SW) and longwave (LW) radiation to designate the visible and infrared portions of the spectrum, respectfully. Visible radiation has shorter wavelengths (higher frequency) than infrared radiation, hence the names shortwave and longwave. Since the earth is on the average 150 million km (93 million miles) from the sun, it only intercepts a very small amount of the sun's total energy output. The amount of shortwave radiation that reaches the top of the earth's atmosphere is called the *solar constant*, it has a value around 1370 Wm^{-2}. It does vary slightly from season-to-season since the earth's orbit is not a circle, but rather an ellipse. The solar constant also varies with sunspot activity, however these variations have little effect on daily weather patterns, although they may have some influence on the earth's climate

Over the course of let's say a year, on average about 70% of the shortwave energy that enters the top of the earth's thermosphere, is transmitted through the atmosphere and reaches the ground (i.e. sunlight). The 30% that is 'lost', is reflected back out of the atmosphere off of cloud tops and highly reflective surface features such oceans, lakes, glaciers and snow covered regions of the planet. On any given day however, much more than 70% can reach the ground while on other days much less. The actually amount of course depends on a number of factors.

As sunlight passes through the atmosphere some of the photos are absorbed by nitrogen and oxygen molecules. These molecules then re-emit a photon at the same frequency as the original, however it can be emitted in any direction. Since this process occurs billions of times per second, photons are moving in every conceivable direction, giving the sky its bright appearance. Some photons however travel through the atmosphere without being absorbed. Sunlight therefore has two components: diffuse-the part that is absorbed and re-emitted, and direct-the part that is not. On an overcast day there is no direct sunlight, its all diffuse, hence it is difficult or impossible for an object to cast a shadow.

Not all areas of the earth's surface receive the same amount of shortwave radiation on any given day; the amount received is primarily a function of latitude and secondarily on cloud cover. This dependency on latitude is a result of earth-sun geometry. The path of the sun across the sky is confined between the Tropic of Cancer (23.5° N) and Tropic of Capricorn (23.5° S). Over the course of a year the tropics of course receive far more shortwave radiation than any other region on the planet.

The amount of energy that is incident on the earth's surface can be approximated by the following little equation:

shortwave radiation   X cosine(θ)

where θ is the angle that the sun makes with the vertical, called the sun angle. Since cosine(0°)=1 and cosine(90°)=0, any unit area of the earth's surface receives the largest amount of radiation when the sun is directly overhead (on the zenith). You may be asking yourself the question, "what

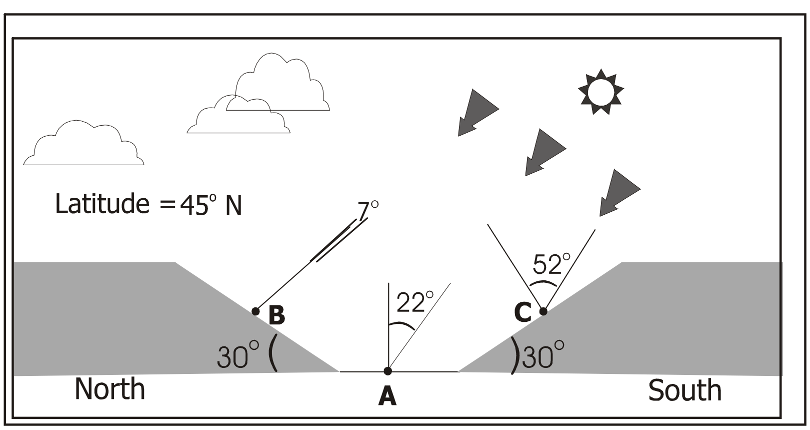


Figure 3.2- Hypothetical shortwave radiation scenario.

happens if the ground is not horizontal or the sun is not directly overhead"? Have no fear, Figure 3.2 shows what occurs in a valley that is positioned at 45^o N, on June 21st.

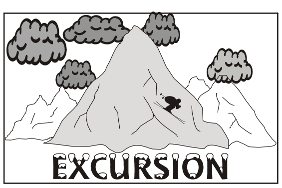
At solar noon the sun's zenith angle is 45^o-23^o= 22^o, remember that since we are in the Northern Hemisphere that the suns' path will be an arc across the southern portion of the sky. If we allow the symbol SW to stand for the maximum amount of shortwave radiation that would reach any surface in this example, then the following holds true.

at: **A**	shortwave radiation X cosine(22^o) = SW X 0.93
at: **C**	shortwave radiation X cosine(30^o + 22^o) = SW X 0.61
at: **B**	shortwave radiation X cosine(30^o - 22^o) = SW X 0.99

It turns out that the south facing aspect of the northern slope (**B**) receives the largest amount of instantaneous shortwave radiation. If position **B** was on a slope of 22^o instead of 30^o, at solar noon it would receive 100% of the available shortwave radiation. Keep in mind that this scenario only holds true at solar noon. At other times of the day the sun angle is a little more difficult to calculate.

Now if you can imagine making this calculation every minute over the course of a day, include shading of the ground due to clouds, mix in different types of vegetation and surface characteristics, and it should not be difficult to understand why certain terrain features heat up or cool down much faster than adjacent ones.

Why is the preceding discussion important? Because our weather is driven by temperature differences that exists between one region and the next. This occurs at all scales, from a small plot of dirt to continents. This differential heating of the land becomes very important when we consider local-scale winds generated in mountainous terrain (Chapter 4). The heating of snow covered south facing slopes in the spring is important to mountain travelers because the additional heating and subsequent production of melt water weakens the bonding within the snowpack, increasing the potential for avalanches. Furthermore because the troposphere is heated from the ground up, any temperature differences that exist at the surface are often transferred to the lower troposphere. If the earth was covered by a uniform surface type such as water, or light brown dirt, grass, or pink flamingos, it's weather would be much less energetic then it currently is. The net result would be less frequent and less intense storms, while the earth's climate would be considerably more uniform.



Why is the Sky Blue?

As shortwave radiation travels toward the surface from the top of the atmosphere, as noted above, it strikes molecules of nitrogen and oxygen. These molecules absorb the radiation and immediately re-emit it in process called scattering. It turns out that nitrogen re-emits shortwave radiation that is rich in the blue wavelengths, and since nitrogen is more than three times as abundant as oxygen, the sky appears blue. Nitrogen selectively scatters blue wavelengths due to the size, spin, and vibrational modes of the molecule (Rayleigh scattering).

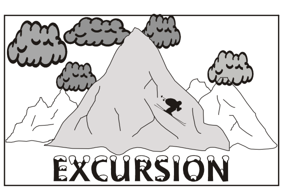
High altitude climbers know that the sky turns increasingly darker shades of blue as they ascend. The explanation for this is as follows: as a climber gains elevation, air pressure and hence the density of nitrogen and oxygen molecules decreases. As a result, the amount of scattering occurring at these higher altitudes is also reduced. As scattering decreases, the sky becomes *progressively black because the only source of illumination is the narrow shaft of direct sunlight. The extreme example occurs on a place like the moon where there is no*

atmosphere, and hence no scattering. Photographs of the astronauts walking on the moon show the "sky" as being black with the sun appearing as a bright disk in the background.

Longwave Radiation

Once shortwave radiation is absorbed by the earth's surface (dirt, rocks, vegetation, water, etc.), the surface warms-up and emits longwave radiation (note that the terms longwave, infrared, and terrestrial radiation are synonymous). Since the temperature of the earth's surface ranges somewhere between -50º C and +65º C (-60º to 150º F) , the longwave radiation it emits has less energy than shortwave radiation emitted by the sun. Unlike shortwave radiation which is transmitted through the atmosphere without very much loss of energy, longwave radiation is absorbed by clouds and many atmospheric gases like water vapor, carbon dioxide and methane. This results in most of the longwave radiation which is emitted by the surface being absorbed within the atmosphere. Due to the nature of atmospheric gases, those that are good absorbers of longwave radiation also have the tendency to re-emit it as well. When clouds and gases re-emit longwave radiation, most of it stays in the lower troposphere. As a result, the total radiation balance (SW and LW) of the earth and atmosphere is pretty complex.

Because clouds are great absorbers and emitters of longwave radiation, the coldest temperatures over the course of a year at any given location generally occur on cloud free nights (or days). A layer of low-level clouds can cause the air temperature near ground level to be roughly 5º C (9º F) warmer than under clear sky conditions. Probably the best way to illustrate the total radiation balance is to consider a high elevation site that is also arid, such as the Tibetan Plateau or the Altiplano of Bolivia. On a typical summer day with few clouds to interfere with incoming shortwave radiation, the surface of the ground heats up considerably. At night the surface emits large amounts of longwave radiation, but since there are no clouds and there is little moisture in the air, this radiation moves higher into the atmosphere. As a consequence, air temperatures during the day are warm for the elevation, but cool rapidly during the evening, and are quite cold at night. Therefore these high elevation sites experience a larger day-to-night (diurnal) temperature range than air in the free atmosphere does at the same elevation.



Conduction and Convection

<u>Conduction</u> is the transfer of thermal energy (heat), from a hot to a cold object. If a warm motionless air mass resides over a cold body of water, heat is transferred from the warmer air to the cooler water. Given enough time, the two temperatures become equal and reach a state called thermal equilibrium. ***.*** *Conduction only occurs when the two bodies (air masses in our case) are in direct physical contact with each other, and there is a transfer of heat energy from the warm body to the cooler body.*

<u>Convection</u> is more abstract than conduction, it signifies the transfer of thermal energy through the movement of a fluid. Consider a patch of dirt that absorbs large amounts of shortwave radiation and becomes warm in the process. There are a number of possible ways for heat to be transferred to

the atmosphere. First, if the air is initially motionless, then conduction transfers heat from the warmer ground to the cooler air. Once the air near the ground is heated to a temperature that is warmer than the surrounding air, it becomes less dense and positively buoyant. As a result the air begins to rise, and new air moves into replace the air (parcel) that is positively buoyant.

In a slightly different example, suppose air is moving over a warm surface. The air molecules that are in direct contact with the surface are heated by conduction, but the carrying away of energy as they move past the heat source is convection.

Conduction and convection are not only fundamental to the atmospheric energy balance, but they are critical in many other geophysical processes. Consider the formation of ice on a mountain lake. In Autumn, as air temperatures decrease, the upper layers of water in the lake lose heat to the atmosphere through longwave radiation, conduction and convection. Fresh water reaches its maximum density at +4º C (39º F), so once the water at the surface of the lake cools to +4º C, it sinks because it is negative buoyancy. During this process, warmer water moves to the surface to replace the sinking water. Ice does not form on the lake until the upper water layers have cooled to the freezing point. Deep lakes take longer to form a layer of ice than shallow lakes or ponds.

Once a layer of ice forms the rate of continued ice growth is a function of the rate at which the water directly beneath the ice layer can lose its heat. Water loses its heat by conduction through the ice, the ice in turn has to lose this heat to the atmosphere. The thicker the ice layer the slower its growth rate since the layer of ice insulates the water below. Since snow is a very good thermal insulator as well, a layer of snow on top of the ice will further inhibit ice growth.

Air Temperature

Since the troposphere is heated from the ground up as noted earlier, on average, tropospheric temperatures decrease with height. This decrease is not constant or uniform, it varies both in space and time. Air temperatures above the earth's surface have been traditionally measured via weather balloons which are released from weather stations all over the globe, two times per day (at 0 and 12 Greenwich Mean Time). Attached to the bottom of a weather balloon is a light weight package of instruments (called a radiosonde) which measure the temperature, humidity, wind speed and wind direction. A plot of temperature versus height from a balloon flight shows the environmental lapse rate (solid line in Figure 3.3), which typically varies between 4º and 10º C km^{-1} (2º to 5.6º F per thousand feet). When the environmental lapse rate is less than about -9º C km^{-1} (5º F 1000^{-1} ft), the atmosphere is considered *stable*, at higher lapse rates (>-9º C km^{-1}) it is *unstable*. Stability refers to how easy it is for a lifted parcel of air to become positively buoyant. A stable atmosphere restricts vertical motion, so that even if a parcel is forced to flow over a mountain or up a frontal boundary, it will have a difficult time maintaining positive buoyancy. When the atmosphere is unstable, positively buoyant parcels may ascend to the upper troposphere without difficulty.

As a parcel ascends it cools independently of the environmental lapse rate. Likewise, when a parcel descends it warms independently of the environmental lapse rate. The rate at which the parcel cools or warms, depends on the amount of moisture contained within the parcel. When a parcel is saturated (contains a maximum amount of water vapor), it cools at a rate called the *saturated adiabatic lapse rate*, which is around 6.5º C km^{-1} (19º F $mile^{-1}$), but varies with moisture content. If the ascending parcel is not saturated, the rate of cooling is around 9.8º C km^{-1} (29º F $mile^{-1}$), which is referred to as the *dry adiabatic lapse rate*. The term *adiabatic* means that a parcel does not exchange any mass or energy with the ambient air it is moving through. This is an assumption, but it turns out that it is a pretty good approximation of how the atmosphere works.

A parcel cools when it rises because it expands, like a helium balloon set adrift (the balloon eventually bursts as it ascends because it expands beyond the elastic limit of the rubber). Parcels of air and balloons expand because air density decreases with height. Descending parcels experience an increase in temperature because they are being compressed Therefore, descending parcels warm at either the dry or saturate adiabatic lapse rate, depending on their moisture content.

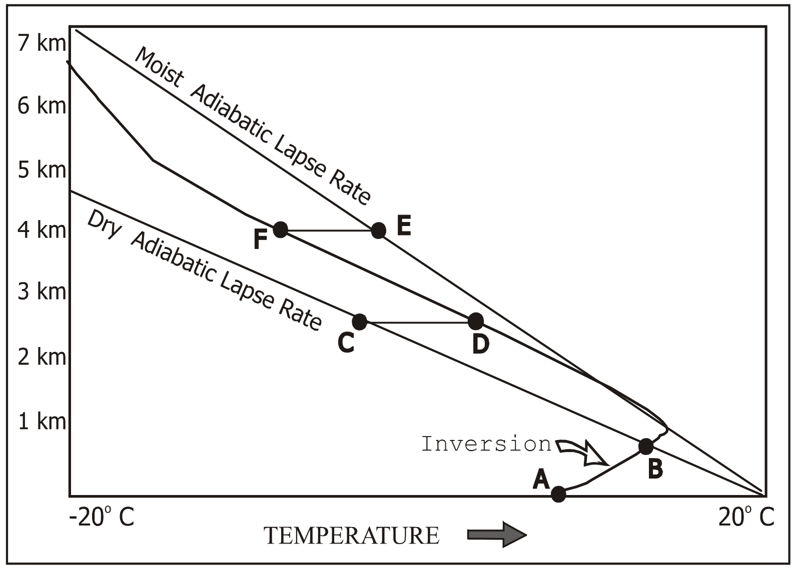
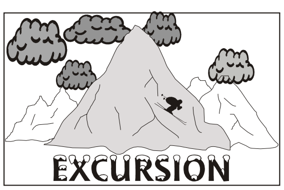


Figure 3.3- Hypothetical vertical temperature profile.

In Figure 3.3 a parcel at **C** is cooler then the ambient air (**D**), therefore it is more dense and negatively buoyant. A parcel at **E** on the other hand is warmer than ambient air (**F**), therefore it is less dense and positively buoyant. The cooling of air as a parcel moves higher works regardless whether it is a thermal or air being forced up the side of a mountain range.

At saturation, water vapor begins to condense and form water droplets. Because condensation releases heat into the parcel (latent heat of condensation), the amount of cooling is reduced. As a result the saturated adiabatic lapse rate is considerably less than the dry adiabatic lapse rate. Likewise when an initially saturated parcel of air begins to descend and warm, water droplets begin to evaporate, this process requires energy for the change of phase from liquid to vapor (latent heat of vaporization). This heat is taken out of the parcel, therefore the parcel warms at a slower rate than when it is not saturated.



Wind Chill

Modern wind chill charts are constructed from data collected in laboratory experiments. You will notice that the wind chill temperature given on a chart is a function of not only the true air temperature, but the wind speed as well. When skin is exposed to air, it not only 'feels' the actual air temperature, but it also experiences an additional heat loss due to convection. The wind chill temperature is a measure of both the actual air temperature and that part which is due to convective heat loss via the wind. For example, you are in an environment where the ambient air temperature is -15^o C (5^o F), and the wind speed is 15 ms^{-1} (33 mph), then the combination of wind and cold is equivalent to being in a windless environment where the air temperature is -40^o C (-40^o F). If you happen to be all bundled up, you should only 'feel' the actual air temperature. Since most outdoor clothing is not 100% wind proof, the temperature you 'feel' may lie somewhere between the actual and wind chill temperature.

Credit for wind chill temperature concept belongs to two Antarctic explorers, Siple and Passel. In the late 1940's they measured the time it took for 0.25 kg (9 oz) of water to freeze, under various *wind and temperature regimes. After considerable experimenting, Siple and Passel were able to develop a equation that related the rate of heat loss from exposed human skin to wind speed and ambient temperature. This was the beginning of the wind chill chart.*

The work of Bluestein and Zecher (1999) has shown some of the deficiencies of the original Sipel and Passel study. This newer work suggest a revision of the wind chill chart in such a manner that wind chill temperatures would <u>*increase*</u> *by roughly 2-3^o C (3-5^o F). This increase in temperatures in the chart is a result of the fact that most 'surface' wind data is actually collected on a tower 10 m (33 ft) above the ground. The net result is that wind speeds at the height of a human above the ground, is considerably less than at 10 m (33 ft).*

<u>Temperature Inversions</u>

There are times the temperature of the lower troposphere increases with height, which is of course opposed to the general trend in the troposphere for warm air to be located <u>below</u> cooler air. When this occurs it is called a 'temperature inversion' or just 'inversion'. In the vicinity of an inversion the atmosphere is very stable (resists vertical motion) often trapping pollutants, haze particles, or small cloud droplets. Inversions can be located at any height in the troposphere, although they are frequently within 1-2 km (6,600 ft) of the surface. Often the tops of marine stratus clouds for example, occur at an inversion.

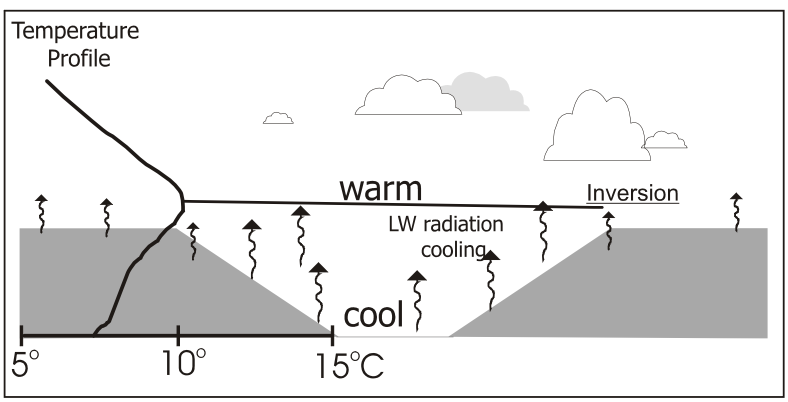


Figure 3.4- Formation of a temperature inversion in a valley.

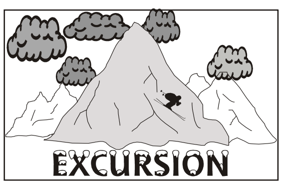
There are two ways which inversions form: the first is by the radiative cooling (longwave) of air closest to the ground; and secondly, by warmer air moving over cooler air (or cooler air moving under warmer air). Inversions that form due to radiational cooling are very common in mountainous terrain. Inversions form in small alpine valleys as well as large intermontane valleys. In a small mountain valley they may form in the evening and dissipate by mid-morning as the surface begins to warm (Figure 3.4). In intermontane valleys, like those in the central Rockies, inversions that develop early in the winter often persist for many weeks or even months. In these large valleys, inversions can be several kilometers deep and can only be destroyed when a new airmass moves into the region. Inversions that form when warm air moves over cooler air usually occur in association with frontal boundaries.

<u>Day-Night Temperature Trends</u>

Everyone is familiar with the diurnal (day and night) change in air temperature that occurs near the surface of the earth. Near the ground the diurnal temperature cycle is a result of daytime heating and nighttime cooling, with the maximum temperature occurring sometime in the afternoon, and the minimum in the early morning hours. There are many times when this scenario does not hold true; for example when a front moves through, when there is considerable cloud cover, or during extended periods of high winds to name a few.

On mountains that have a permanent snow or ice cover the temperature regime can be more complicated because of the presence of low lying inversions induced by the snow and ice. The work of Ludecke and Kuhle (1991) on the lower slopes of K2 and Everest (between 5000 and 6000 m) indicates that the diurnal temperature range is larger over bare ground (dirt and rock) than over snow or ice. However, the diurnal regime over snow and ice still exceeds the regime in the free atmosphere because at night the longwave cooling of the snow exceeds the cooling of the free atmosphere. During the day near surface, the temperature of the air may or may not exceed the free atmosphere, depending on the presence of inversions and the strength of the wind.

In summary, <u>air temperatures near the ground will tend to be more extreme than free atmospheric temperatures at the same elevation</u>. Actual temperatures are site specific and are a function of surface characteristics and the prevailing weather. Temperature extremes occur on days and nights with little cloud cover and little moisture. In addition, dry climates (deserts) experience larger diurnal and seasonal temperature regimes than wet climates. At night in wet climates, the presence of large amounts of moisture in the lower troposphere effectively blocks outgoing longwave radiation emitted by the surface. During the day clouds block a portion of the incoming shortwave radiation, and in addition a large part of the shortwave energy that reaches the surface is used for evaporation instead of heating the air.



How Warm Do You Feel?

Next time you are in the mountains on a sunny day, notice how warm it feels when you are in direct sunlight. If the sun goes behind a cloud or if you walk into the shade, you experience a rapid drop in skin temperature. Do you think the air temperature changes this rapidly as well? Here is an brief explanation of what occurs. When you are standing in direct sunlight, your body absorbs considerable amounts of shortwave radiation. Hence you warm-up a lot more than the air. While in the shade, you experience a rapid cooling simply because you are no longer absorbing direct sunlight and because your emitting large amounts of longwave radiation (yes-people emit longwave radiation!). Meanwhile, the air temperature has remained pretty much constant.

This heating/cooling pattern is very noticeable to any one who travels on a glacier on a sunny day, especially when the wind is light. The intense heat that you feel is due to the large amounts of shortwave radiation that is being reflected off of the snow and ice, and absorbed by your body (and its dark clothing). In reality the air temperature is not as warm as you think it is based on how warm you feel, in fact the air temperature is considerably cooler than you would guess it to be.

This illustrates an important point about the proper way to site an outdoor thermometer. We see lots of the thermometers mounted in the sun, there is one word for this: BOGUS. This also applies to those little mercury key-chain jobs that most of us carry. Keep it out of the sun for a accurate measurement. Real thermometers, that is those used to collect weather data, are placed in some type of ventilated shelter which allows good air flow but keeps the instrument shaded.

Chapter Quiz

1. True/False: The sky appears blue because oxygen decreases with elevation?
2. _______________ is the transfer of heat from a hot to a cold body.
3. Clouds absorb and emit ______________ radiation.
4. True/False: In the Northern Hemisphere, south facing slopes generally receive less shortwave radiation than north facing slopes?
5. For a given air temperature, the wind chill temperature ______________as the wind speed increases.

4

WINDS

Chapter Highlights

- ✓ *Learn about the all-powerful jet stream*
- ✓ *Understand how wind interacts with mountainous terrain.*
- ✓ *Become an expert on mountain, valley and glacier winds.*

Strong winds are probably the most inconvenient and potentially life threatening meteorological phenomena that you will encounter in your travels among the mountains. Strong winds can occur under a very broad range of conditions; basically in combination with any other type of weather scenario. In addition, by virtue of the mountain environment, high winds can generate extremely poor visibility by lifting dirt or snow from off of the ground, or blowing snow horizontally during a storm. As was noted in the previous chapter, strong winds are also responsible for the 'apparent' reduction in air temperature that mountain travelers frequently experience (i.e.-wind chill). Being able to recognize high wind situations and avoid them if possible is a skill that you should attempt to master.

Jet Stream Winds

As a mountain traveler you should care about the jet stream for two simple reasons: first, wind speeds frequently exceed 70 m/s (150 mph), in short, they are the strongest persistent winds found in the lower atmosphere. Secondly, almost all big storms develop, mature, and dissipate in the vicinity of the jet stream. We have been using the singular so far, however there are actually three jet streams: the arctic, polar (Figure 4.1), and sub-tropical . The good news is that these 'brutes' are for the most part confined to elevations above 5,000 m (16,400 ft or

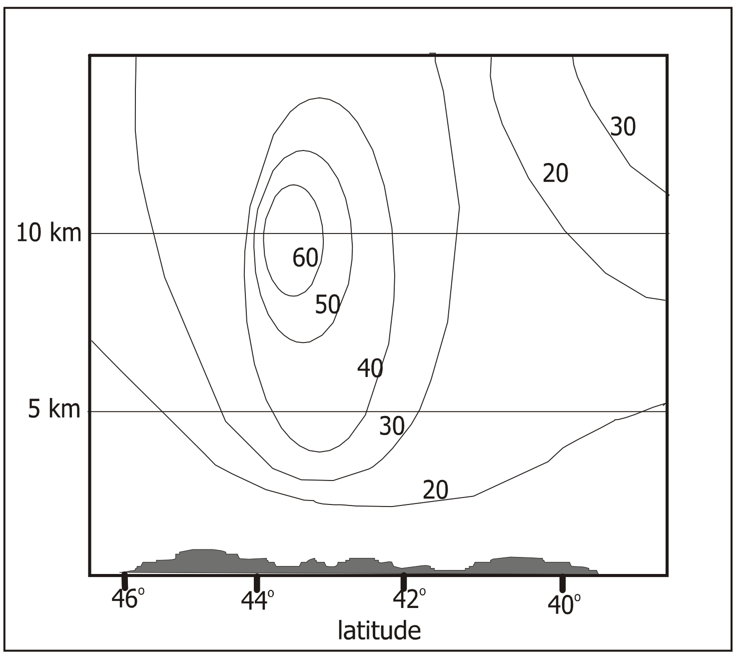


Figure 4.1- Idealization of the polar jet. Contours represent wind speeds in ms^{-1}, with air moving into the page.

500 mb). Furthermore the current suite of computer weather models that are in use, forecast the position and strength of the jet streams quite well. Now for the bad news: you do not want to be at a high elevation when one of these jet streams is in the neighborhood.

The arctic jet (we often drop ‘stream’) is the least understood and the most recent one to be discovered. It is generally located between 60º-80º N, and a elevation of 500-400 mb This is not a permanent feature, in other words there are days when the arctic jet is present, but many days when it is absent. In fact, it is absent more days then it is present. Likewise, it is not continuous around the globe, it may appear in the Canadian Arctic but not over Scandinavia, or vice versa. Wind speeds in the arctic jet stream typically range from 30-45 m/s (60-90 mph).

The polar jet is the one that most people are familiar with in large part because it has a very large influence on mid-latitude weather. However, the polar jet is not fixed in any one position, it makes a sinuous path across the Northern Hemisphere (there is also one in the Southern Hemisphere), but is most frequently found between 35º-60º N. In addition, it is weaker in the summer months when the temperature difference between high-latitudes and mid-latitudes weakens. There are two regions of the globe where winds in the polar jet reach maximum speed; off the east coast of Asia, near Japan, and off the east coast of North America, near Newfoundland. During the winter both of these regions have very large temperature contrasts due to the proximity of very cold land masses to the west and relatively warm oceans to the east. The level of strongest winds within the polar jet stream occur between 8-10 km (300-200 mb). Wind speeds range from 40-90 m/s (90-200 mph), and on occasions even higher.

The sub-tropical jet develops between the mid-latitudes and the tropics. It generally lies between 20º-30º N and is an important factor in the generation of thunderstorms over the southern USA in the summer and it also produces strong winds over the Himalaya during the cooler months of the year. The sub-tropical jet forms in the descending portion of the Hadley cell, therefore the level of maximum winds tends to be centered around 10-11 km (~200 mb). Wind speeds are similar to those of the polar jet, but can at times exceed 110 m/s (240 mph) <u>over a limited area.</u>

One of the more difficult concepts to grasp is the fact is that all three of these jet streams vary in strength and position from week-to-week or day-to-day. In fact if you were to examine daily weather charts of upper level winds, you would find that on some days only one of the jets is discernable while on another day two or even three of the jets are evident. By way of example, on any given day the polar jet may be strong over the western North Pacific, while the arctic jet is absent and the sub-tropical jet is weak over the North Pacific but strong over the Gulf of Mexico. Two days later the polar jet may have weakened considerably while the sub-tropical jet has moved further north and intensified. As you can imagine, there are many different possible combinations.

The horizontal dimensions of these jets vary greatly, they are typically 500-800 km (300-500 mi) wide and vary in length from a thousand kilometers to several thousand kilometers. Embedded within the jet stream are moving zones of maximum winds known as *jet streaks*. I<u>n a statistical sense, the higher you climb the more likely you are to encounter strong winds</u>. When the sub-tropical jet lies over the Himalaya in the winter, strong winds are the rule not the exception. However, if strong winds are present at 7,000 m (23,00 ft) for example, it does not necessitate strong winds at 4,000 m (13,100 ft) There are a number of factors that determine whether jet stream winds extend to lower elevations.

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<u>Discussion</u>: Suppose you are on a three day climb of Mt. Rainier, and your' camped at 3400 m (11,150 ft) on the Tahoma Glacier. You poke your head out of the tent at dawn on summit day, and you observe upper level cirrus clouds moving rapidly towards the northeast-your clue that upper level winds are strong (polar jet). But what about wind conditions on the Summit, 1000 m (3,280 ft) above? Your best indicator is to look for blowing snow higher up the mountain, if you do not observe any, then the winds are probably light, at least at present.

Keep in mind that upper level clouds in the preceding scenario act as tracers. Blowing snow, dust, or fast moving clouds simply help us monitor the wind at a distance. Strong winds can develop without any cloud development of course, but when clouds are present use them to estimate mid-level wind speeds. When you are in the mountains it can be difficult determining whether or not a jet is in the neighborhood. One indicator is the presence of long rows or streaks of cirrus clouds. In this case the jet is parallel to the long-axis of the clouds. It is also very helpful if you know <u>a prior</u> what the major storm patterns are for the area you are in. For example, in the Cascades major storms move in from the northwest-to-southwest, it would be rare to have persistent strong winds from the southeast. The Mountain Weather Survey chapters of this book are intended to provide you with this type of information.

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Interaction of Wind with Terrain

<u>Blocking</u>

Mountains of any size impede or disrupt the movement of air, the larger the mountain or mountain range, the larger it's impact on the flow of air over, through, or around them. One very common and important class of terrain-wind interaction is *orographic* blocking, or simply blocking. Essentially what occurs is that a mountain or mountain range acts as a barrier to low and mid-level winds. In many cases, low-level (roughly within 1 km or 3,280 ft of the ground) air is not capable of being lifted over the top of the range. As a result either the low-level air moves around the sides of the range or it slows down at the base of the mountains forming a pool of air. This does not mean however that all the air on the windward side of the range is blocked. As Figure 4.2 indicates, <u>mid-level</u> air is often able to flow over the top of the blocked layer as well as the mountain range.

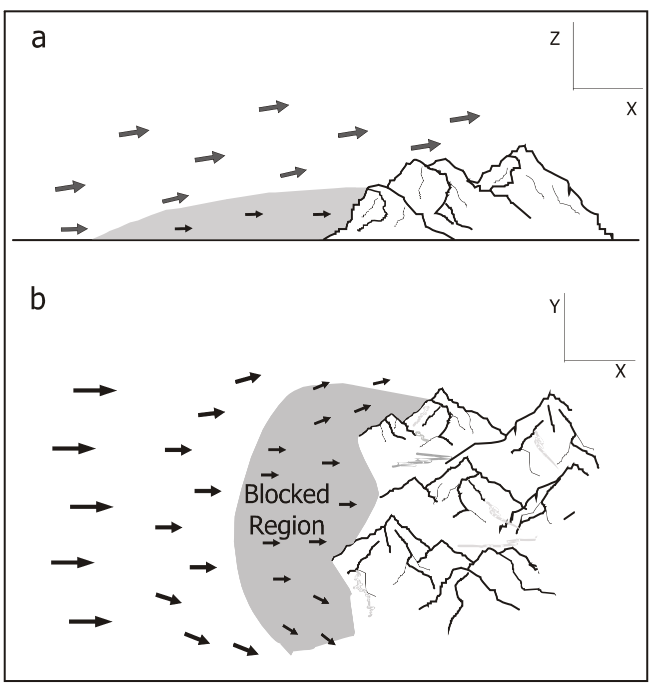


Figure 4.2- a) vertical cross-section of a blocked layer. B) Plan-view. The length of the arrows indicates relative wind speed.

The upstream extent to which

blocking occurs is a product of several factors: the stability (change in temperature with height) of the lower troposphere, the height of the mountains, and the speed of the approaching winds. Furthermore if the troposphere is stable and the air is unsaturated, as it usually is, air will cool at the dry adiabatic lapse rate as it rises up the windward slopes of the range. This produces a pool of cooler air at the base of the mountains, which is referred to as a 'blocked layer'. Air within the blocked layer is usually only a fraction of its upstream speed, and its direction is often altered from the original inflow direction as well. In cases of very strong blocking, air within the blocked layer can reverse direction and flow back upstream. The most favorable situation for blocking to occur is when we have a very stable lapse rate (it takes a lot of lift to get air over the mountains), a high mountain range, and weaker winds moving directly toward the range.

If you happen to be hiking on the windward slopes of a mountain range when blocking is occurring, you should expect winds to be <5 m/s (<10 mph) within this layer. Above the blocked layer however, wind speeds return to their upstream values. If you were to hike through the upper regions of a blocked layer, you will experience a sharp increase in wind speed as you ascend.

You may be wondering what happens to the flow as it moves to lee-side of the mountain range depicted in Figure 4.2? Air that has moved around the edges of the barrier can often curve towards the backside of the mountain in what are called leeside vortices. Air that has gone over the top of the barrier can descend to the surface in smooth layers, or if the lee slope is steep, the air often becomes turbulent, forming rolls and eddies. In rather special cases, the flow accelerates as it descends the lee-slope creating a downslope windstorms. Many of these features will be discussed in the following sections.

As you read through the text and view the drawings you should keep in mind that what is being described are idealized scenarios. In real mountainous terrain we rarely find nice symmetrical mountains, instead we find a series of irregularly spaced ridges interspersed with jagged mountains. All of this adds up to produce some very complex air flow patterns.

Barrier Jets

One of the consequences of blocking is the formation of a zone of strong mountain parallel low-level wind called a barrier jet. Barrier jets do not form during every blocking event, in fact well developed barrier jets may only develop several times per year in a given mountain range. In order to understand how this wind develops, imagine a layer of air which is moving perpendicular to a mountain range, at some height above the ground as drawn in Figure 4.3 As a result of blocking, a mesoscale area of higher pressure (meso-high) forms at the base of the windward slopes (essentially a cold pool of air). Keep in mind that this is a separate feature from the synoptic-scale pressure gradient which generates the winds in the first place.

Now the speed and direction of the flow upstream of the range is a balance of the pressure gradient and Coriolis forces as well as friction. The incoming flow slows down due to the local increase in pressure. Since the Coriolis force is proportional to the speed of the wind, as the wind decelerates in the blocked zone, the wind turns toward the direction of low pressure, which is to the left in Figure 4.3. However, the winds begin to re-accelerate because of increase in pressure that occurs in the cool pool. Even though the Coriolis force tries to shift the winds back towards the right, the mountain range is in the way, as a result the wind remains parallel to the mountains.

The formation of a barrier jet typically takes about 4-6 hours from the onset of blocking. They are fairly common along the west slopes of the Sierra Nevada and along the eastern Rockies. They are much more common during the winter when the lower troposphere is stable. Winds speeds in these jets vary considerably, in very strong cases speeds on the order of 15-30 m/s (35-70 mph) have been reported (Parish 1982). Barrier jets can extend out from the mountains as much as 50-100 km (30-60 miles).

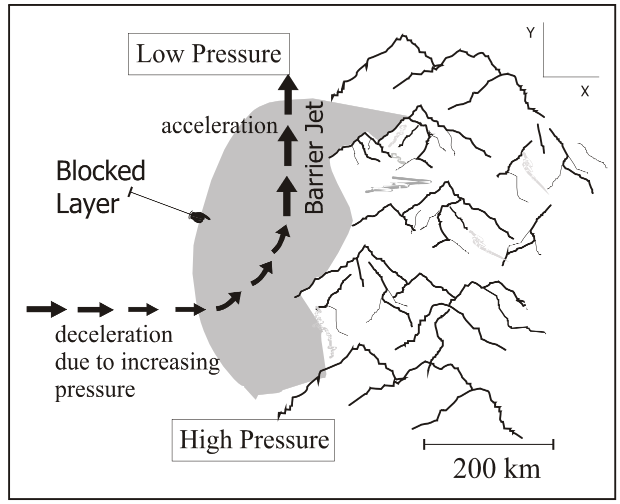


Figure 4.3- Example of a barrier jet.

Flow Over and Around Large Mountains

We will now consider what takes place when air moves over and around a large mountain or mountain range. If the summit lies well above the blocked flow or if no blocking occurs, and if the troposphere has a stable lapse rate, wind speeds across the summit will often increase from their upstream free atmospheric values (in other words the winds are accelerated). In order to understand why this occurs, we are forced to consider the propagation of *gravity waves*. Simply put, gravity waves are regions within the troposphere where the air oscillates up and down, in similar fashion to swells on the surface of the ocean. Gravity waves are created because air is forced up and over mountainous terrain, however they only occur when the troposphere is stable (Figure 4.4). Once gravity waves are produced, wave energy propagates away from the mountain in all directions. As a result, the path of the air moving over and downstream of a large mountain or mountain range often has a wave-like structure.

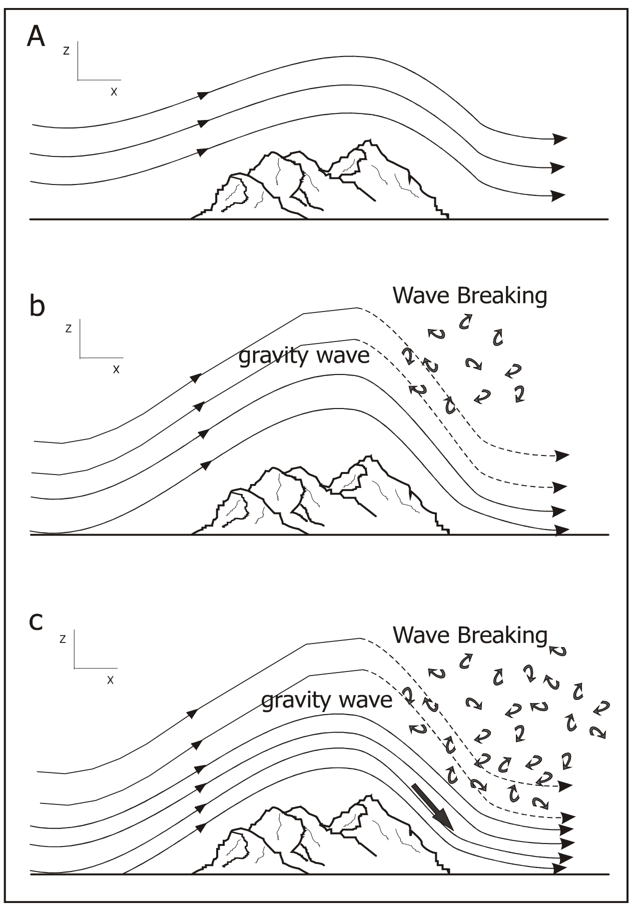


Figure 4.4- Evolution of gravity waves and wave breaking.

There are times when the amplitude of gravity waves becomes so large that they overturn or break, like a ocean wave steepening and breaking as it moves onshore. When wave breaking occurs it disrupts the horizontal flow by forming a turbulent region within the middle troposphere. As a result, subsequent air that flows over the mountain is effectively “squeezed” through the space between the summit and the turbulent layer above (Figure 4.4c). Since the volume of air passing between the summit and turbulent region remains unchanged, the flow must accelerate as it crosses the summit and moves down the lee-slope. Wave breaking is not the only way that air is accelerated over the summit of a large mountain, an inversion located several kilometers above the summit, will have the same effect. In this case air is squeezed through the space created by the inversion and the mountain below.

The type of flow described in the preceding paragraph and illustrated in Figure 4.4c is an extreme case. Nevertheless, anytime the middle and upper troposphere is highly stable, gravity wave propagation above a mountain can cause the upstream winds to be accelerated as they move over the mountain. It would be instructive to compare upstream free atmosphere winds speeds with speeds across the summit of a large mountain. However, since observational data of this nature is rare, a computer model of tropospheric flow was used for analysis. Using model terrain that ranges in height from 2,000-4,000 m (6,660-13,200 ft), and starting the model with a broad range of initial wind speeds, the results show that winds accelerate from 30% and 80% as they move over the summit.

We also need to consider under what tropospheric conditions there is little or no amplification of the winds over the top of a mountain. Table 4.1 highlights five flow regimes where amplification of the wind is minimal.

Table 4.1 Flow regimes with little summit-level wind amplification.

Atmospheric State	**Explanation**
Weak upstream flow	If upstream wind speeds are below some critical value (roughly 10 m/s or 20 mph), gravity wave development above the mountain is weak.
Unstable mid/upper troposphere	Gravity waves cannot form or propagate in a region where the lapse rate is unstable.
Inversion located at or below summit level	This limits the amount of air that can flow *over* the mountain from low-levels. In addition, at inversion level the winds are often quite weak.
Deep low-level blocking	If the depth of a block layer extends to near summit level, a considerable amount of the flow will be diverted around the mountain, speeds may also be reduced.
Upstream obstacles	If an upstream mountain or ridge generates turbulence or flow deflection, wind amplification will usually not occur over the second summit. This is highly dependent on terrain configuration (distance between ridges), the height of each summit with respect to each other, etc.

In addition to flow over the summit of a mountain, a considerable amount of air can flow around the sides. Additional computer modeling work suggests that in cases of extreme blocking, where the majority of the air approaching the mountains flows around the sides of the barrier, winds on the flank of a mountain can also experience large accelerations (Olafsson & Bougeault 1996). Wind amplification, whether it occurs near the summit or on the flanks of a mountain, is constantly changing as the inflow direction, stability, and upstream wind speed change.

Gap and Gorge Winds

If you have ever been hiking through a mountain gap or pass where the winds were considerably stronger in the gap, when compared to the area just outside of the gap, then you have first hand experience of gap winds. These are the classic type of winds that most mountain travelers envision when meteorologist talk about the acceleration of winds in an mountainous environment. In this book we will refer to a gap as a small elevated divide on a ridge or as a pass through the mountains; where the width is roughly equal to the length. Winds that are funneled through gorges and valleys, will be referred to as gorge winds. [Note that the term valley wind is reserved for a different class of wind discussed later].

Let's start this discussion by considering an example from hydraulics that will serve as an analogue to gap and gorge winds. Many readers are familiar with the "venturi effect", in which a fluid accelerates as it is forced through a constriction in a pipe or hose. By way of illustration consider water flowing through a solid piece of pipe that is 10 cm (4 in) in diameter as depicted in Figure 4.5. What happens to the flow of water if the end of the pipe narrows down to 5 cm (2 in)? If the pressure on the upstream end of the pipe is constant, then water is forced to accelerate through the constriction. Basic physics shows that the speed of the water in the 5 cm diameter section is roughly 4-times faster than in the 10 cm (4 in) diameter section. Most of the acceleration occurs in the transition section of the pipe.

In the example cited above, the constriction in the pipe plays the role of a mountain gap. As air flows through a gap it is accelerated because higher pressure forms near the gap entrance. Gap winds are generally not 4-times faster than the upstream wind speed because of other effects not discussed, nevertheless gap winds can be quite substantial.

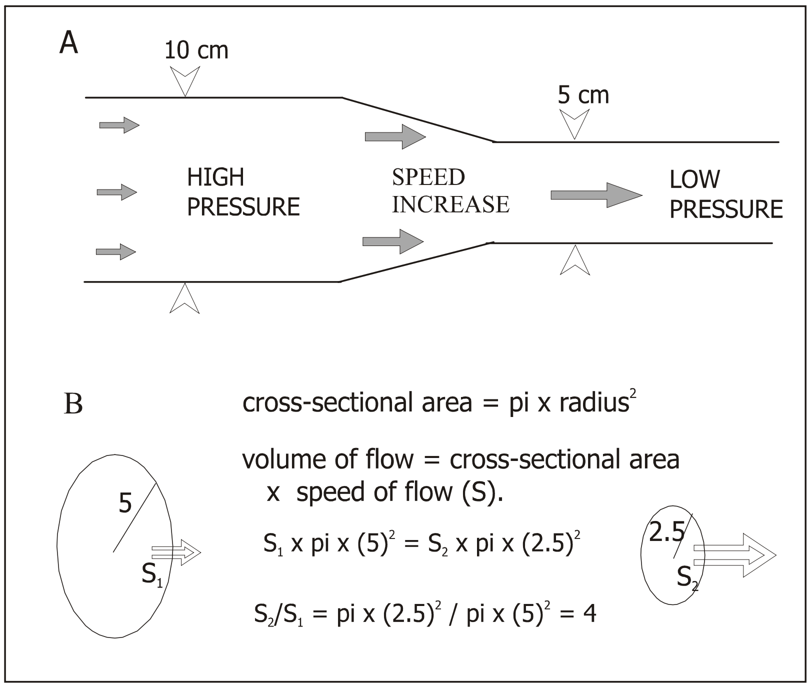


Figure 4.5- Flow of water through a pipe.

The example given in Figure 4.5 is an 'instantaneous' snap-shot of the flow, but what happens if we observe the flow over a period of time? In essence the pressure in the pipe or gap is not steady in time, this in turn produces surges in the flow. For example, in Figure 4.5, if the volume of the water passing through the constriction is smaller than the volume of water which enters the pipe on the upstream end, then the pressure in the wide section of the pipe is forced to increase. This increase in pressure in turn produces an extra acceleration in the flow through the constriction. This 'tug-of-war' between water pressure and the speed of the flow is what generates the surges of water in the pipe. If we extrapolate this example to air moving through a mountain pass, then it is not difficult to understand why the winds are often gusty.

Gorge winds are similar in nature to gap winds in many respects, however, due to the greater length of gorges, the flow is not uniform down the long axis of the channel. In areas where the gorge narrows, flow usually accelerates while in wide reaches it often decelerates (Figure 4.6). It is also fairly common to find hydraulic jumps within a gorge flow. Hydraulic jumps occur when the wind speed decreases and the depth increases over very short distances, resulting in a wave-like feature or abrupt 'jump' in the depth of the layer of air. Hydraulic jumps form in rivers as well and are generally avoided by river runners since large jumps are highly turbulent.

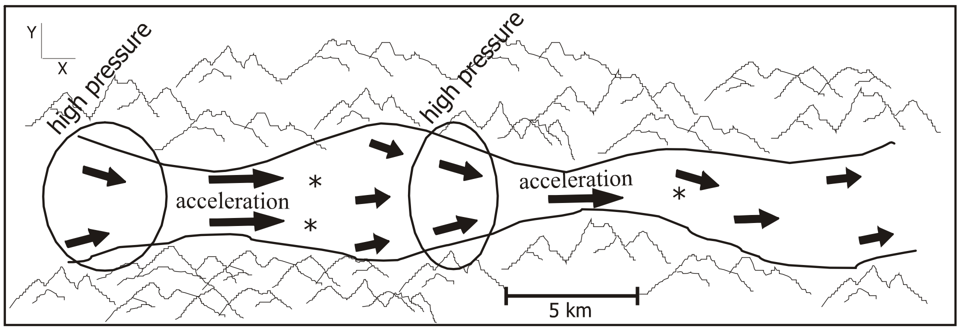


Figure 4.6- Gorge winds. In locations marked with a * it is common to find hydraulic jumps- areas where the wind rapidly decelerates.

Some of the more notable gorge winds are found in the Pacific Northwest, especially in the vicinity of Fraser Gap and Columbia Gorge. These gorge winds develop during the winter months when a ridge of high pressure produces cold arctic air over southern British Columbia and eastern Washington and Oregon. Jackson and Steyn (1994) found that the strongest winds are within the first 500 m (1650 ft) of the surface and that the depth of the cold air decreases down the long axis of the gorge. The cold air in a gorge is often bounded by an inversion, which in this case acts like a rigid lid forcing the flow beneath it to accelerate. At the outlet of gorges, wind speeds are often two or three times what the ambient pressure gradient alone would have generated. Typical speeds of 20-30 m/s (44-66 mph) were reported by Mass *et al* (1995) for Fraser Gap winds, with some observations of gusts as high as 45 m/s (100 mph) over the San Juan Islands.

<u>Air Flow over a Ridge (micro-scale)</u>

Understanding how air moves over a ridge should be of particular interest to climbers, skiers, and snowboarders; since it has important implications on the distribution of snow. This section should help explain why windward slopes often have a thin snowcover, while leeward slopes are buried in snow.

As air moves over a ridge, it usually reaches its maximum speed at the crest of the ridge, as shown in Figure 4.7. Essentially this case is no different than the gap winds described in the previous section. In this new scenario however, the 'gap' is formed by the ridge itself and the stable layers of air some distance above the ridge. This means that air which is moving up the windward slope is forced to converge at the summit. To the lee of the summit air moves into a region where the flow is divergent (i.e. expands), hence the speed decreases. Furthermore as a result of the decrease in speed, a localized area of high pressure forms to the lee of the ridge. If the angle of the leeward slope exceeds 15^o-20^o, then air moving down the slope no longer remains in contact with the surface.

At times air is redirected back up the slope in the form of a lee-side eddy, due to the presence of the localized area of high pressure. These eddies often contain significant amounts of turbulent air, nevertheless their speeds are a fraction of the wind speed in the free atmosphere above. This whole process is referred to as flow separation, which reflects the fact that a fairly homogenous flow regime develops into two different regimes because of the steep leeward slope.

We will take a closer look at the distribution of snow by wind at the conclusion of this chapter, however this conceptual model of flow over a ridge helps explains why snow is frequently transported from windward slopes, where wind speeds are high, to the leeward slopes, where speeds are much slower. In fact the amount of snow that the wind can transport is a function of the speed of the wind. High winds can transport large amounts of snow, once the speed decreases however, the carrying capacity of the wind decreases as well. As a result, large quantities of snow are typically deposited on leeward slopes in areas where wind speeds diminish and eddies form. The complicating factor is that the wind does not always blow from the same direction. For example, snow can also be deposited in gullies and couloirs when the wind blows parallel to the long-axis of a ridge, in a process called cross loading. In this case the upstream side of the gully plays the role of a ridge, snow is deposited at the base of the upstream slope as wind speeds start to decrease.

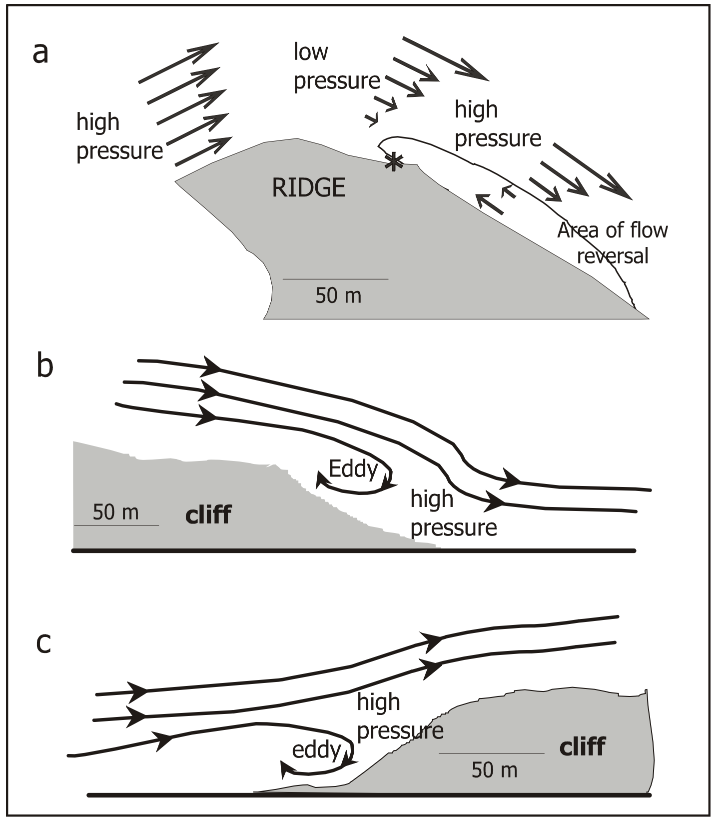


Figure 4.7- a) flow over a small ridge. b) flow over a cliff. c) flow up a steep slope.

Flow separation can occur over a variety of topographic features as illustrated in Figure 4.7 b,c. As air encounters a steep cliff or mountain face, the air decelerates at the base of the cliff and forms a eddy due to an increase in local air pressure at the base of the cliff. Flow separation also explains why wind breaks, like a small grove of trees or a outcrop of rocks, work as well as they do. As a rule of thumb, wind breaks are effective to a distance of about 2 or 3 times their height. For example, if a group of trees or rocks are 10 m (33 ft) high, the zone of reduced winds extends out to a distance of about 20-30 m (65-95 ft), as measured from the base of the barrier.

Downslope Winds

These types of winds occur when air flowing over a mountain barrier accelerates on its descent down the lee-slope, as illustrated in Figure 4.8. Downslope windstorms occur in virtually all moderate to large mountain ranges and are much more common during the cooler months of the year than during the warm months. They are also associated with rapid increases in lee-side temperatures, leading to rapid snow melt. Downslope windstorms are commonly know in North America as *Chinooks* and in Europe as Foehns.

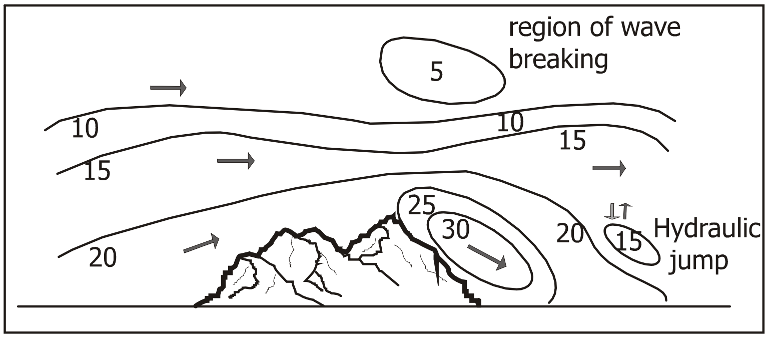


Figure 4.8- Downslope windstorm with contours of windspeed (isotachs) in ms^{-1}.

In North America downslope windstorms are well documented, large events occur on average several times per year in the western USA. They are most common in the Front Range of the Colorado, along the east side of Glacier and Waterton Lakes National Parks, and in the Canadian Rockies, just east of Banff-Jasper. They are also fairly common on the east side of the Sierra Nevada's, near Bishop. In Europe, foehns are common on both the north and south side of the Alps, depending on the direction of the geostrophic winds. When downslope winds occur, strong gap and gorge type winds may also be produced as air flows out of the mountain passes and valleys.

In order for a downslope windstorm to develop, the following requirements need to be meet (Brinkmann 1974): 1) A mountain range has to be high enough and large enough to disrupt the air flow over its crest. Isolated mountains (like stratovolcanos), even large ones are not prime candidates for downslope windstorms because air often flows around the sides instead of over the summit. 2) There must be moderate-to-strong winds (>20 m/s or 44 mph) at mountain crest level.
3) The direction of the wind should be nearly perpendicular to the long axis of the barrier. 4) The air over the mountain must have a stable lapse rate. The steeper the leeward slope, the easier it is to generate strong downslope winds. However, the steepness of the windward slope is not important.

The January 11-12, 1972 downslope windstorm in Boulder, Colorado (Klemp & Lilly 1978) was noteworthy because it produced a considerable amount of damage in the area. In this particular storm wind speeds of 50 m/s (110 mph) were observed at ground level near the base of the Front Range. One common feature of Front Range wind storms is the presence of an inversion, roughly 2 km (1.2 miles) above the highest terrain. When a downslope windstorm does occur, the area of strong winds usually does not extend much more than 30-50 km (20-30 miles) from the base of the mountains. In addition, it is common for the downstream end of the strongest winds to terminate abruptly in a hydraulic jump.

As you can see from this short description, a number of tropospheric variables, primarily the stability, wind speed and wind direction have to be 'in alignment', before a downslope windstorm can occur. It turns out that this alignment occurs less frequently than you might imagine. As a result, areas known for downslope winds might typically experience three to five significant events and 10 to 15 minor events during the course of a year. Of course there are many mountainous regions around the globe where these events go unnoticed because there are no systematic weather observations.

Discussion: As a mountain traveler take note that the strongest winds in downslope windstorms extend from about halfway down the lee slope to the base of the slope. The good news is that most downslope windstorms last less than 12 hours. They can occur in clear air or in association with clouds. In many cases clouds and precipitation are produced on the windward side of the barrier and over the crest.

On the lee-slope however, air is drying as it descends, so partly cloudy skies are the norm.

The leeward temperatures are generally warmer than the corresponding temperatures at the same elevation on the windward slopes. The reason for this is that due to the release of latent heat as moist air ascends the windward slopes, the air over mountain range is relatively warm. As the air moves down the leeward slope it is compressed and hence it warms even further.

If you are in the mountains and you think you are in a situation where a downslope wind event is beginning, you will first notice an increase in wind speed and rising temperatures. There are no real precursors or atmospheric parameters that you can monitor from the field that will suggest a downslope wind event is going to develop. About the only thing you can do is find a sheltered location and wait it out.

There are a few places in the world where downslope winds lead to a <u>cooling</u> on the lee-side of the range, these events are called boras and are common in the Dinaric Alps (western Slovenia and Croatia), as well as the region between the Pyrenees and the Alps (Smith 1979). Boras from when a deep layer of cold air accumulates on the windward side of the mountains. As air flows over and down the lee side of the range it of course warms . However, what warming does occur is not enough to off-set the initially cold air temperature. Therefore at the base of the leeward slope, temperatures become cooler instead of warmer. Boras occur in ranges where the mountains are of modest height. If the mountains are too high, cold air can never become deep enough to spill over the top, although some air will flow through gaps and gorges producing localized areas of cold air drainage.

Some of the diagrams in this section may have given the reader the impression that downslope winds only occur on the leeward slope of an isolated mountain range. In reality, they can occur to the lee of any range that meets the stated criteria. It is possible to be in the middle of a large mountain range and still experience downslope winds. As was noted earlier in this chapter, the type of flow that occurs in the middle of a mountain range depends on the distance between the mountains and their relative heights. In addition, if the intervening valleys are full of cold air it is unlikely that a downslope windstorm will develop, even if all other factors are favorable.

<u>Lee-Waves and Rotors</u>

There is a special class of mountain induced gravity wave phenomena called lee-waves which are characterized by a series of troughs and crests in the flow directly above and downstream of a mountain barrier (Figure 4.9). This type of flow has much in common with downslope wind storms, however there are some significant differences as well. When flow over a mountain generates lee-waves, it means that wind speeds near the summit and above are quite high. However, unlike downslope windstorms, surface winds on the leeward slope and at the base of the mountain are usually pretty light. The

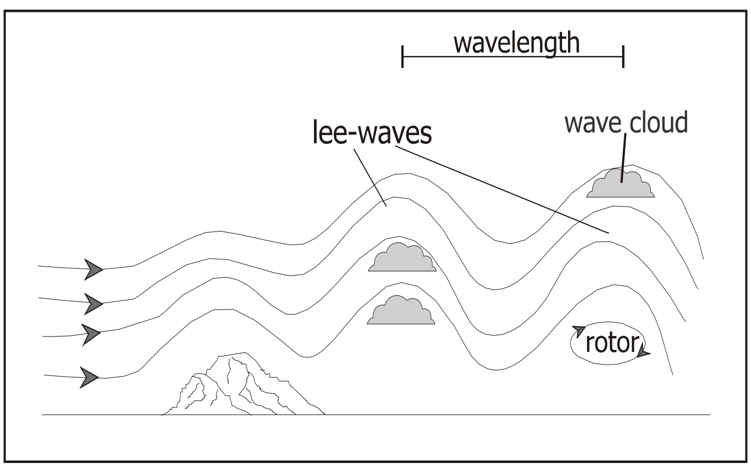


Figure 4.9- Lee-waves with wave clouds and rotor.

generation of lee-waves depends on the vertical profiles of wind speed and atmospheric stability as well as the height and width of the mountain.

Lee-waves stay fixed with respect to objects on the ground even when strong winds are blowing through the waves. This occurs because lee-waves are a collection of gravity waves that have their own individual characteristics. When these gravity waves are combined to produce lee-waves, they form a stationary pattern in the air above the mountain. The height of lee-waves is strictly a function of mountain width and height; where a narrower and steeper mountain produces higher and thicker lee-waves.

The wind speed and stability on the other hand, determines the horizontal wavelength of the lee waves (distance between successive waves). Durran (1986) states that most lee-waves have horizontal wavelengths between 5-25 km (3-15 mi) and that the minimum speed of summit-level winds needed to generate them is on the order of 10-15 m/s (25-35 mph). It is quite common to have a wave form directly above the summit of a large mountain as well. If sufficient moisture is available in the middle troposphere, clouds may form in the crest of these waves (lenticular clouds), which is the only way an observer on the ground knows of their presence. A pilot on the other hand may experience moderate to heavy turbulence when flying through or near lee-waves.

Rotors form on the downstream side of mountain ranges when mid-tropospheric winds form an eddy-like feature. Rotors are like a continuous series of ocean waves, that break in one specific spot as they move toward the beach. They most often form in conjunction with lee-waves or downslope windstorms. They may reach the ground for short periods of time but more often remain aloft. Clouds usually do not form in rotors, although if a rotor is located near the ground it can stir up large quantities of dirt, and be a serious hazard to aircraft.

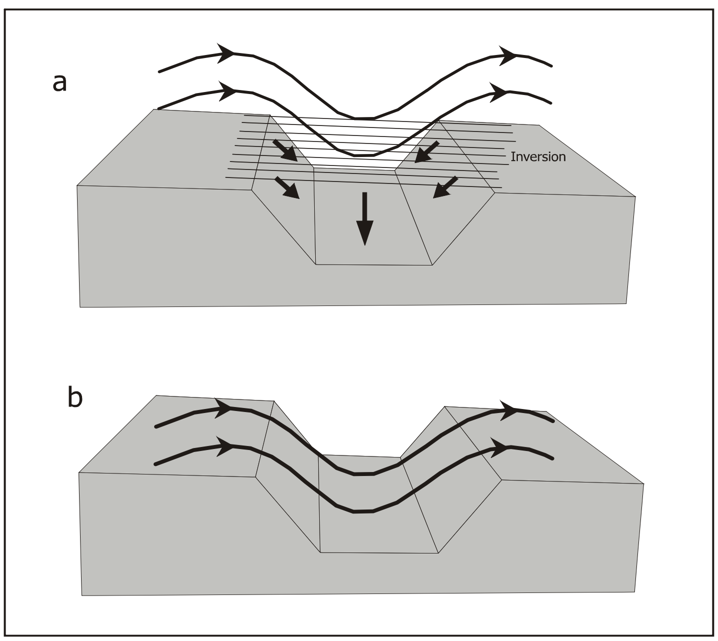


Figure 4.10- a) Due to an inversion, geostrophic winds cannot enter the valley. b) Without an inversion winds descend deep into the valley.

Wind Flow over a Valley

We have already discussed air flowing down the long-axis of a valley, but what happens when synoptic-scale winds blow over the top of, and perpendicular to the long-axis of a valley? Figure 4.10 shows several possible flow scenarios. Not all larger-scale winds descend into the valley, in fact they often do not. The determining factors are:

* Valley width
* The height of terrain upstream of the valley
* Stability of the air in the valley
* Presence of an inversion at the top of the valley
* Strength and depth of thermally generated winds within the valley

* Speed of the large-scale (geostrophic) winds

Thermally Generated Winds

This class of winds should be of great interest to mountain travelers, because they have a high frequency of occurrence and are found in every type of mountainous environment. Since all winds are ultimately thermally generated, you are probably wondering what specific type of winds we are talking about? Thermally generated winds are essentially near surface winds that develop in response to differential heating/cooling of the earth's surface. You may recall from chapter 3, that heating of the ground by shortwave radiation and the subsequent cooling by longwave radiation occurs on the local-scale, basically from slope-to-slope. This is why sun angle, slope aspect, latitude, time of year, and physiographic properties of the surface determine when and where thermally generated winds will form. These types of winds usually only develop when the low-level geostrophic winds are light
(< 5 m/s or 10 mph).

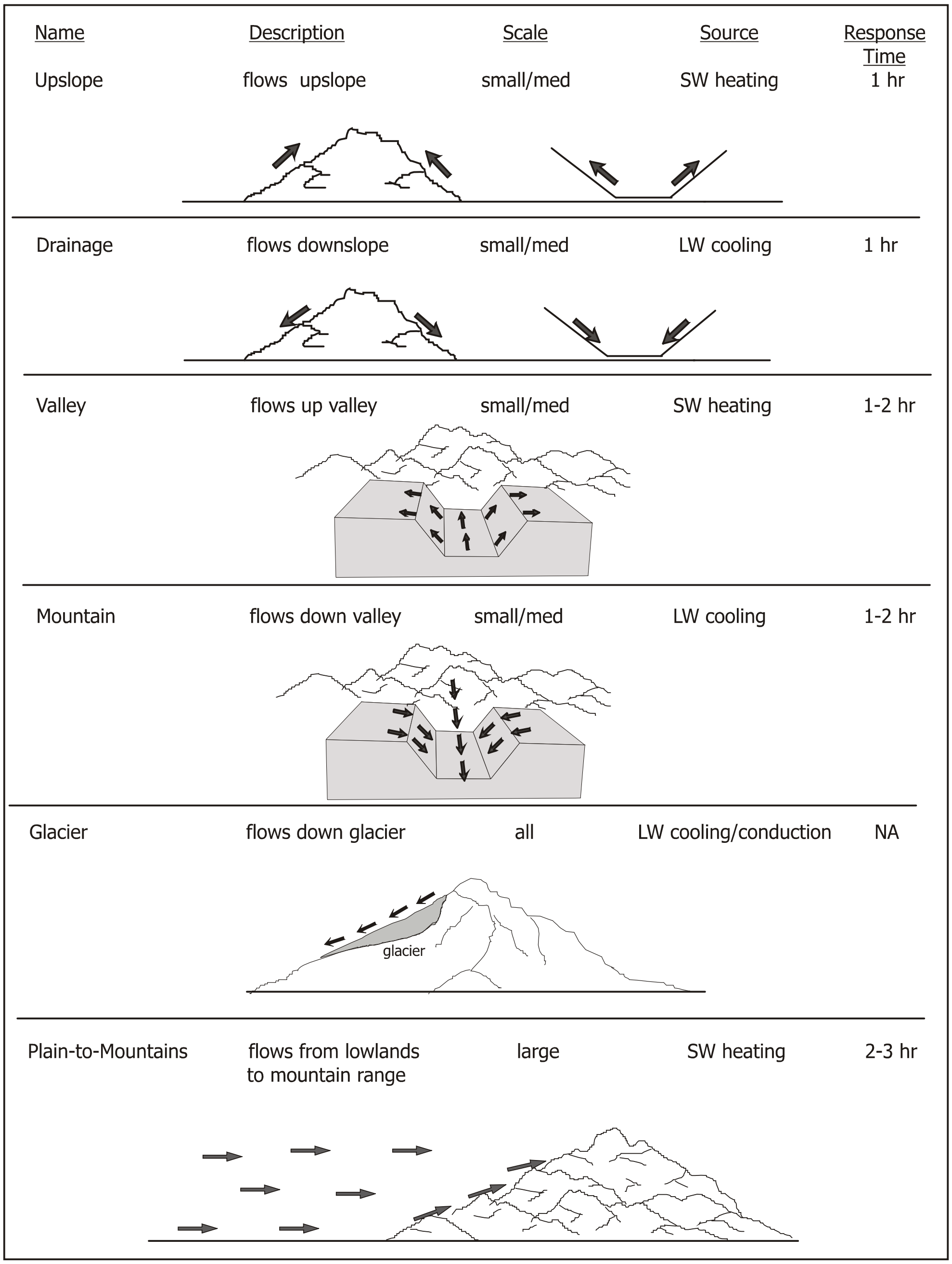


Figure 4.11- Various examples of thermal winds.

Figure 4.11 illustrates the various types of thermally generate winds which will be discussed in this section. In addition, this figure also gives an overview on the spatial scale at which they occur, what process(es) are responsible for their development, and the amount of time it takes for them to develop.

As a rule of thumb, small-scale winds (those that involve a smaller volume of air) develop on the order of 30-60 minutes, while large-scale winds take 1-3 hours. For example, in small alpine valleys, upslope and drainage winds can begin to develop 20-30 minutes after sunrise or sunset, respectively. On the much larger plain-to-mountain scale, thermally generated winds may take several hours to develop since considerably more air has to be heated. Glacier winds on the other hand develop independently of the solar cycle, and can blow for days on end if the geostrophic winds remain light.

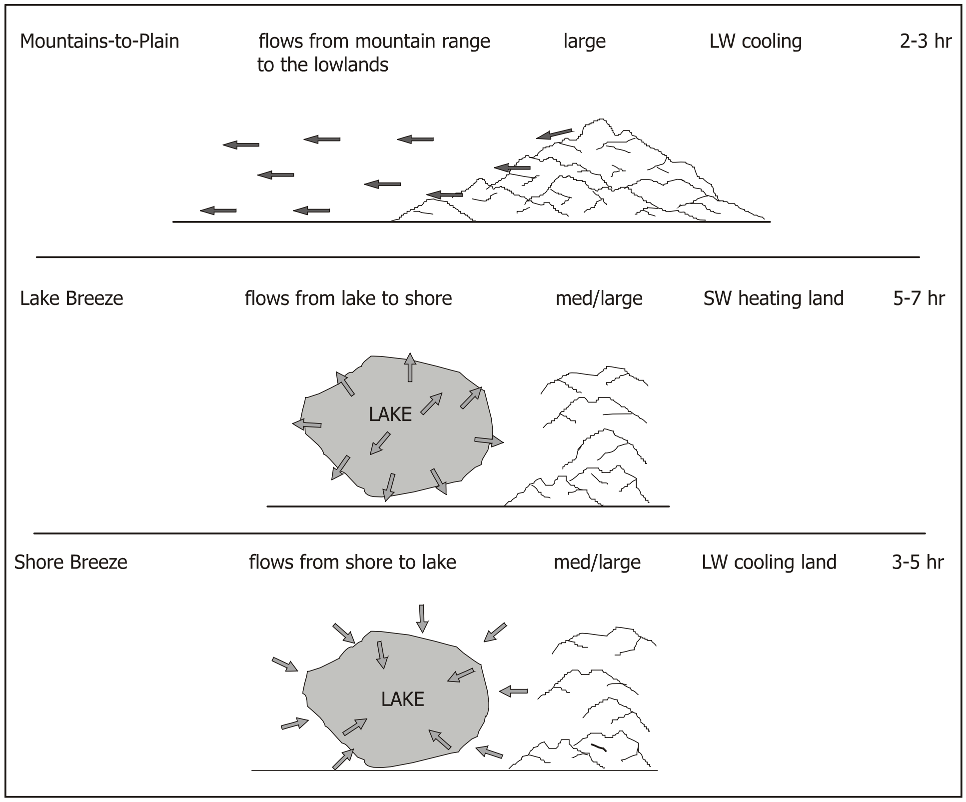


Figure 4.11 continued

Slope, Mountain-Valley, Mountain-Plain Winds

In order to understand how these winds develop, consider the idealized two-dimensional valley configuration in Figure 4.12. We will assume there are no ambient winds, the ground is snow-free, and clouds are not a factor. Figure 4.12a shows what the thermally generated wind structure might look like in the early morning, shortly after sunrise. The air next to the valley floor and slopes heats up because of its contact with the rapidly warming ground. The air in the middle of the valley on the other hand is considerably cooler than the air near the sunlit slopes or valley floor. Since warm air is less dense than cooler air, it rises. Slopes that are in the shade may still have cold air moving downslope at this time, but by late morning, upslope winds are usually fully developed throughout the entire valley. At night, because of extensive longwave cooling of the surface, the air nearest the ground cools more than the air in the middle of the valley atmosphere (Figure 4.12b). This results in air moving down the slope in the form of a drainage wind. Studies have shown that the speed of a drainage flow is proportional to the steepness of the slope, as well as the temperature gradient between the air near the ground and the air at the same height but some distance out from the slope. The larger the temperature gradient and the steeper the slope, the faster the flow.

Both upslope and drainage winds have typical depths on the order of 10-30 m (30-100 ft), and speeds from 2-5 m/s (5-12 mph). From time-to-time however both the depth and speed of these winds can be much larger. You should note that upslope and drainage winds develop above many types of sloping terrain, it does not matter if the sloping terrain is on the flanks of a mountain or the sides of a valley. What does matter to a large degree is the type of ground cover- a wet heavily vegetated slope will take a lot longer to warm up then a dry rocky slope. As we will see below, a slope covered by snow or ice will not heat up at all. It the sky is overcast, these winds may not develop at all.

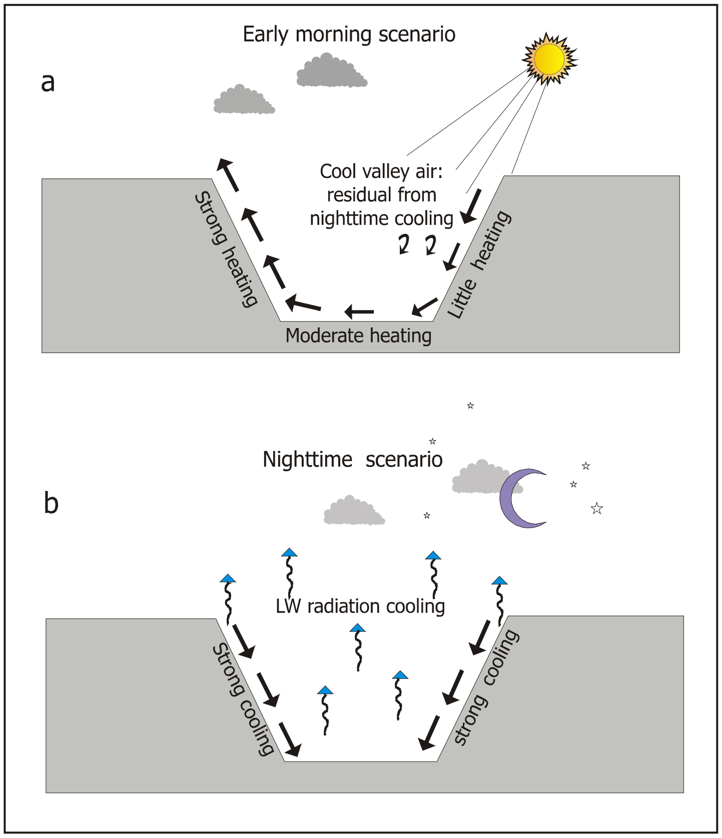


Figure 4.12- a) Generation of upslope flow in the morning.
b) Drainage flow at night.

The next step is to consider winds that form parallel to the long-axis of a valley. During the course of a day, air within a valley heats up considerably more than the air at the same elevation outside of the valley over the lowlands. This produces a valley wind that moves up the valley towards the mountain (Figure 4.13). At night, since air in the valley cools more than air over the lowlands, the valley air becomes cooler and more dense. The resulting flow moves down valley in the form of a mountain wind. These types of winds have a depth on the order of 50-200 m (160-650 ft) for typical high alpine valleys, and speeds on the order of 4-7 m/s (10-15 mph).

In large valleys these winds can be much deeper, for example, Vergeiner & Dreiseitl (1987) report valley winds 1-2 km deep in the Inn Valley of Austria. Since valley and mountain winds involve a larger volume of air than either slope or drainage winds, they typically require several hours of heating or cooling of the valley atmosphere before these winds are initiated. In addition, numerous observers have reported finding a layer of winds located just above valley or mountain winds. These winds move in the opposite direction as the valley or mountain wind, and therefore are called anti-valley or anti-mountain winds.

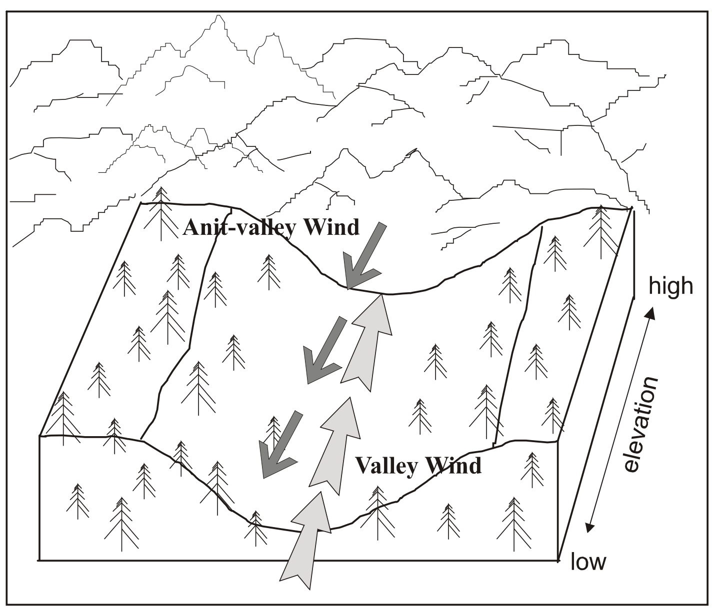


Figure 4.13- Valley wind with anti-valley wind above (return flow).

On a larger scale, a plain-to-mountain circulation can be established during the daylight hours, which reverses direction at night. This type of circulation system is basically a larger scale version of the mountain/valley winds. During the day, the air adjacent to the mountain slopes becomes considerably warmer than air at the same height over the plain, which creates low pressure over the mountains and draws in air from the plains. At night, air over the mountains becomes cooler than air over the plains, so cold air flows down from the mountains. Plain-to-mountain winds generally develop three to four hours after sunrise, while mountain-to-plain winds develop an equal number of hours after sunset.

What we have been discussing so far in this section are idealized models of how thermal winds develop. In the real world they tend to be much less homogeneous in space and steady in time. Added complications arise due to the non-homogeneous nature of terrain and physiographic features. For example, variations in slope angles, ground moisture, vegetation types, snowcover, etc. all have important effects on thermally generated winds. Further complications come from the development of inversions during the night, the interaction of geostrophic winds, and cloud development.

The following example illustrates that our conceptual models are essentially correct, however real world flows tend to be much more complex. Case in point, the world's deepest valley, the Kali-Gandaki of Western Nepal (located between Dhaulagiri and Annapurna massifs), was the site of two short field programs that studied the notorious local mountain-valley winds, during the summer and early autumn seasons. Researchers at one point used a motorized glider to monitor these winds (Neininger *et al* 1985, Egger *et al* 2000). Near the valley floor they measured valley winds that had sustained speeds of 10-15 m/s (20-35 mph) with gusts as high as 25 m/s (55 mph). Wind speeds were not uniform through the length of the valley, they tended to be highest in the central section.

At times valley winds would persist through the night, albeit in a much weaker state. On other nights however, the valley wind would be replaced by a weak mountain wind. The high speeds of the valley wind are attributed to the massive heating which takes place on the Tibetan Plateau during the day. The heating of the surface of the plateau creates low pressure over Tibet, which draws higher pressure air located over central Nepal, through the valley into Tibet. This is a prime example of a mesoscale plain-to-mountain flow combining with a valley wind to generate exceptionally persistent and strong winds.

During a separate field program, Ohata *et al* (1981) spent an entire year observing thermally generated winds in a unnamed valley of the Khumbu Himal region of eastern Nepal. One curious phenomenon they observed during the monsoon season (June-September) was the persistence of valley winds long after sunset. There was significant cloud formation in the late afternoon and early evening which caused the valley atmosphere to remain warmer due to the release of latent heat, than the time of day alone would suggest. Their data also revealed a weaker thermal wind regime during the monsoon season, compared to the pre- and post-monsoon seasons. Weaker winds during the monsoon season were attributed to an increase in surface moisture at low elevations, greater snow cover at higher elevations, and greater cloud coverage.

Buettner & Thyer (1966) spent several summers studying thermally generated winds in the 5 km (3 mi) long Carbon River Valley located on the northwest flanks of Mt. Rainier. They observed that the typical depth of a valley/mountain wind was from one-quarter to one-third the depth of the valley. The overlying anti-wind had about the same depth. Valley winds reached maximum velocity in early afternoon while mountain winds peaked just before sunrise. Typical speeds for both wind regimes was on the order of 3-6 m/s (7-13 mph). They also noted that speeds were rarely constant and seemed to vary with a period of about 20 minutes.

Glacier Winds

When a stagnant layer of air lies over a large mass of snow or ice, such as a glacier, there is considerable cooling of the air due to conductive heat loss to the ice below. In time this layer of air becomes significantly colder than the ambient air at the same elevation in the free atmosphere. As a result, cold air moves down slope in the form of a glacier wind. Like all other types of thermally generated winds, glacier winds usually only form when the geostrophic winds are light. Glacier-type winds also develop over snowfields (snow breezes), the air adjacent to the surface does not care whether it is in contact with snow or ice. Glacier winds and snow breezes can also develop over horizontal ground, as long as a large temperature contrast is maintained between the glacier (snow) and the non-glaciated (snow-free) ground.

Unlike other thermal winds which often change direction at sunrise and sunset, glacier winds and snow breezes are generated independently of the solar cycle. This should not be taken to mean that glacier winds maintain constant speeds. In fact most of the glacier winds that have been studied display some type of diurnal wind speed fluctuation. The speed of a typical glacier wind ranges from 5-10 m/s (12-24 mph), with a depth generally not exceeding 100-200 m (300-600 ft). Speeds are not constant with height, there is usually a distinct speed maximum, which can occur however, at just about any height. Due to the partial dependency on longwave cooling of the air adjacent to the ice/snow, glacier winds tend to be fully developed on days/nights which are predominately cloud free.

Over the past 40 years there have been a considerable number of studies on the subject of glacier winds. What all these studies show is that there is significant variation from one glacier to the next. This testifies to the wide range of conditions and spatial scales under which glacier winds have been studied, and more importantly it tells us that we cannot assume that all glacier winds have identical properties.

Glacier winds, like all thermally generated winds, exhibit speed fluctuations due to turbulence. They can also extend well beyond the terminus of the glacier. During the day, a glacier wind may flow down valley and converge with a valley wind, creating a zone of vertical motion that is a preferred area for cumulus cloud formation. At night, glacier winds can reinforce mountain and other drainage winds as they all move down valley. In general, glacier winds will reach maximum speeds on or near the terminus, since beyond the terminus they encounter an increase in surface roughness (trees, large boulders, etc.), which decelerates the flow. As the studies cited in this section indicate, glacier winds show considerable variation in depth, speed, and extent. This illustrates the difficultly of trying to construct a conceptual glacier wind model that fits all cases.

Lake Breezes

Since thermally generated winds result from the differential heating of the earth's surface, the heating/cooling contrast between a lake and adjacent land is sufficient to generate a lake-to-land breeze during the day (lake breeze), and a land-to-lake breeze at night (land breeze). The development of these breezes over a large lake is a result of the different thermal properties of water versus land (dirt or rock). As a starting point for this discussion we will assume that the water and the land have the same temperature at sunrise, we will also assume that over the course of the day equal amounts of shortwave radiation is absorbed by the water and dirt (we are neglecting reflection off of the water). As a result, equal amounts of energy are absorbed by each square meter of the respective surfaces. However, if we were to measure the temperature of each surface during the early afternoon, we would find that the land is considerably warmer than the water, why? Water conducts heat much better than dirt or rock, hence heating occurs deeper within the water column.

In addition, since the density of water is temperature dependent, small circulation cells are initiated within the upper few centimeters (inch or less) of the water layer. The net result is that the energy absorbed by the water is spread over a larger volume of material as compared to what occurs in dirt or rock. In fact most of the shortwave radiation absorbed by the land, stays in the uppermost layer (few millimeters), giving it a very warm surface temperature. This has an important effect on the temperature of air overlying each surface. By afternoon the air over the land is warm and the air over the lake is still relatively cool. This results in low pressure developing over the land with slightly higher pressure over the water. This pressure gradient in turns produces a light wind which blows from the lake to the shore.

During the evening and early night time hours, both the water and the land cool. Since the land does not retain very much heat (low thermal heat capacity), it cools rapidly due to emission of longwave radiation. On the lake the upper 10 cm (4 inches) of the water still has a considerable amount of heat in storage, which is by the way slowly released over the course of the night. As the surface of the water cools due to the emission of longwave radiation, heat energy from the water below is conducted to the surface, replenishing the heat loss. Therefore, at night the air over the land is cooler than over the water, and a land-to-water breeze is generated.

Lake (or land) breezes typically range from 2-3 ms^{-1} (4-6 mph) and only occur on days with considerable shortwave radiation input. As you might imagine, on days with extensive cloud cover or when the land is moist, these types of breezes are suppressed or at least delayed.

Wind and Snow

We have already briefly discussed the transport of snow from the windward to leeward side of a ridge, this section will however elaborate on the processes involved. We will use the term snow "particles"to designate snow that is already on the ground or picked up and transported by the wind, and reserve the term snow "crystals" for snow which is falling from clouds as original precipitation.

Movement of Snow by the Wind

There are three ways that snow particles can be transported: rolling, saltation, and suspension (Figure 4.14) In reality the same forces are at work for each of these three mechanisms, the difference being how far above the surface the snow particles are lifted. It is important to consider the degree of bonding between the individual particles. Shortly after being deposited on the ground, snow particles quickly form bonds with adjoining particles. The stronger the bonds and the larger the particles, the more difficult it is for the wind to move them. The degree of bonding is a function of the past and current air temperatures. Cold dry snow is more susceptible to wind transport than wet heavy snow, which has a temperature near freezing. Research has shown that the bonding strength of particles increases as the temperature of the snowpack approaches 0^{o} C (32^{o} F).

Rolling occurs when wind speeds are not strong enough to lift a particle off of the ground. Due to wind stress as air moves over and around a particle, the upstream side of a particle is subject to higher pressure than the leeward side. This is what causes particles to either roll or slide across the surface. Angular particles tend to have sharp corners broken off, so that after a few minutes of rolling, most particles have a rounded shape.

Saltation occurs when the wind speed is large enough to create a small amount of aerodynamic lift, which carries the particle into the air stream. The lift is created in the low pressure zone on the back side of the particle. However, due to its weight, the particle starts to fall back towards the surface, following a parabolic trajectory. When a particle hits the surface, it can loosen surrounding particles and make them more susceptible to lifting, since the impact weakens or destroys ice bonds. Some snow scientists (the folks who get paid to play in the white stuff) suggest that the impact of a falling particle actually works like an elastic collision (the transfer of momentum from the particle that is landing to the one that is being ejected), sending new particles into the air stream. One of the best ways to understand the saltation process is to observe it on a windy day at the beach. If the wind speeds are moderate, dry sand can be lifted above the surface, only to travel a 10-20 cm (5-10 in) before crashing into the surface and ejecting other sand particles into the air stream. There is a minimum wind speed at which the saltation process begins in snow, the exact value of course depends on the amount of bonding. Mellor (1965) for example, found that speeds from 6-10 m/s (13-22 mph) were high enough to initiate saltation in dry snow. Empirical data also shows that the mass of snow transported increases exponentially with increasing wind speed. The depth of a typical saltation layer is on the order of 0.5-1 m (20-40 in), with a typical horizontal trajectory of less than 2 meters (~ 6 ft).

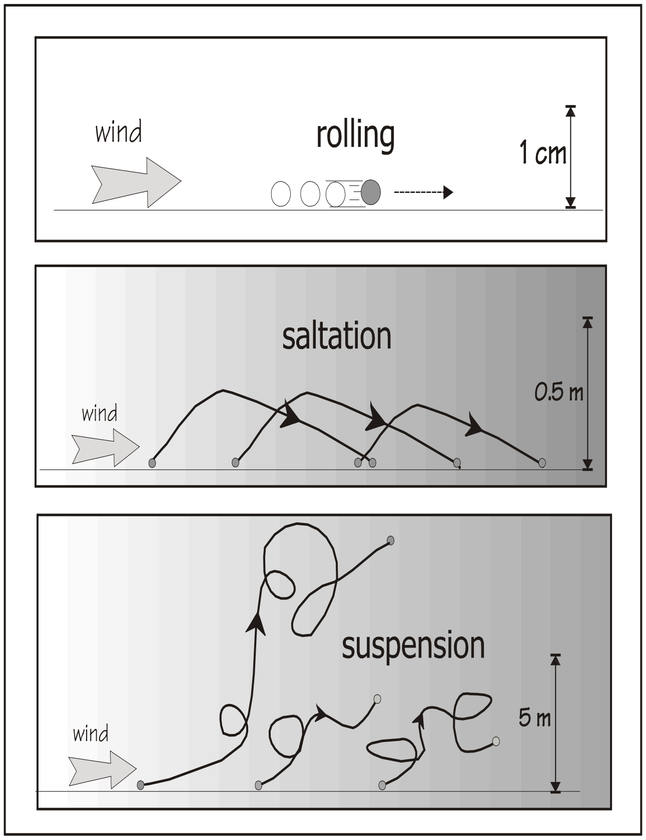


Figure 4.14- Three modes of snow transport.

The third and final way that snow particles are moved is by suspension, also called turbulent diffusion by some authors. When wind speeds are high (approximately > 15 m/s or 32 mph), it is possible for particles to be carried horizontally for 10-100 meters (30-300 ft) without making contact with the surface. Snow that is transported across a ridge undergoes saltation or suspension, or both. The determining factors are wind speed, particle size, and degree of bonding.

By virtue of continued movement, wind transported snow particles become rounded, typical diameters range from 0.1-1.0 mm (0.004-0.04 in). Snow drifts and snow slabs have a density range of 200-400 kg m^{-3} (12-24 lb ft^{-3}), where glacier ice for comparison has a density of 700-900 kg m^{-3} (43-55 lb ft^{-3}) and fresh water 1000 kg m^{-3} (62 lb ft^{-3}). The tight packing of rounded particles in wind slabs results in a layer with considerable mechanical strength (it can take a lot of stress without breaking). However, these dry particles <u>do not</u> bond well with older types of snow that lie underneath or next to the slab. The mechanical strength is a mixed blessing. The slab itself is quite strong, however, pressure waves can travel great distances through it, so when failure does occur, it is generally catastrophic (involves a much larger volume than it would in unconsolidated snow).

As backcountry skiers know all to well, skiing across wind blown snow is often difficult due to the presence of sastrugi and dunes. The steep side of sastrugi point in the direction of the prevailing wind during the time of formation (these are created by the removal of snow). In addition, when skiing from unconsolidated snow to area of wind blown snow, skis often slow down and feel “sticky” because wind blown snow is very dry. Wind blown snow is dry because surface moisture was removed or shed while the particle was being transported.

Cornices form when snow is blown to the crest of a ridge. Snow collects on the very edge and forms a horizontal protrusion. In time, the pre-cornice grows vertically as well as horizontally. Due to the strength of the bonds among the particles, large volumes of snow are able to hang out over the edge of the ridge and not break off. Over the course of the winter, due to its weight, the cornice begins to deform and take on the classic curl shape

___

Discussion: *If you have ever camped in a tent during a blizzard, you know that the rate of snow accumulation on and around a tent can be phenomenal. For a given rate of precipitation (the amount of snow falling from the clouds), the amount of snow that accumulates on a tent increases dramatically with increases in wind speed, up to a point. What happens is that the wind blows snow into the sides of a tent at a much faster rate than it would accumulate if the wind was calm and the snow was falling straight down. If the winds are very strong (> 25 ms^{-1} or 55 mph) during a blizzard, then chances are not very good that very much snow will accumulate, because most of it is blown away. One of the keys to successfully enduring a major snowstorm is to be mentally prepared to occasionally leave the tent to clean off of it, or expend the energy to dig a snow cave*.

___

Snow Fences

Marring the landscape as they do so well, snow fences play a key role in helping man attempt to tame nature. Snow fences are more commonly used in the Alps than in any mountain range in North America, in large part because the Alps sustain a much larger population base.

The sole purpose of a snow fence is to control the deposition of snow. There are several different types of snow fences, two of the more common ones are blower fences and collecting fences. Blower fences are designed to accelerate the wind in a given area, so that its snow load will be deposited farther downstream, much like wind accelerating through a mountain gap. Collecting fences on the other hand are used to reduce wind speeds by generating turbulence, so that snow is deposited directly to the lee or around the fence. If the fence can disrupt the flow, wind speeds will decrease, and the carrying capacity of the wind will be reduced. Snow fences are used to reduce the amount of snow accumulating across roads, railroad tracks, buildings, etc., which saves removal costs and reduces potential hazards (Figure 4.15). Collecting fences may also be used to increase the amount of snow accumulation near a lake or pond, in hopes of increasing spring runoff into the lake.

Collecting fences are most often constructed as non-solid barriers (they have open spaces between the slats), made from wood or metal and are from 2-3 m (6-10 ft) in height. Slats can be placed either horizontally or vertically. The height, density of the slats (open area versus solid area), and the height of the bottom gap are all important in determining how efficiently the fence collects snow. By altering the height, density, and bottom gap, the designer can regulate to some degree, how the snow will accumulate behind the fence. If you have seen very many snow fences you will have noticed how they are often not aligned vertically. There is little loss of efficiency if they tilt in either direction from the vertical. In order to take greater advantage of the upstream decrease in wind speed produced by a single row of fences, a series of snow fences are often used

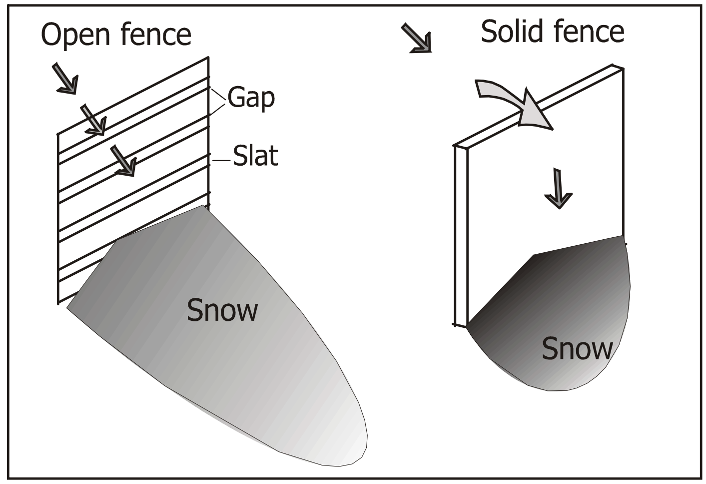
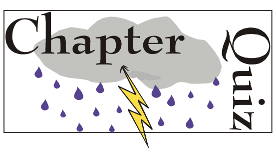


Figure 4.15- Two types of snow fences.

In terrain that has a high avalanche potential, fences are often seen high on the ridges. Strategic placement of fences is critical if they are going to be effective at decreasing the amount of snow that accumulates in the avalanche starting zone. The designer needs to know the direction(s) of the prevailing winds during snowstorms, and what slopes have the highest potential to slide.



1. Name the three jet streams?
2. True/False: Glacier winds only occur at night?
3. Barrier jets form when the lower atmosphere is ____________.
4. True/False: Downslope wind storms only form when the upstream wind is nearly parallel with the mountain range.
5. True/False: Valley winds develop faster than slope winds?
6. True/False: The stronger the bond between ice particles, the more difficult it is for the wind to transport it.
7. Slope winds take about ________ minutes to develop after sunlight 'heats' the air?
8. True/False: Downslope wind storms are common in the western Rockies?

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).

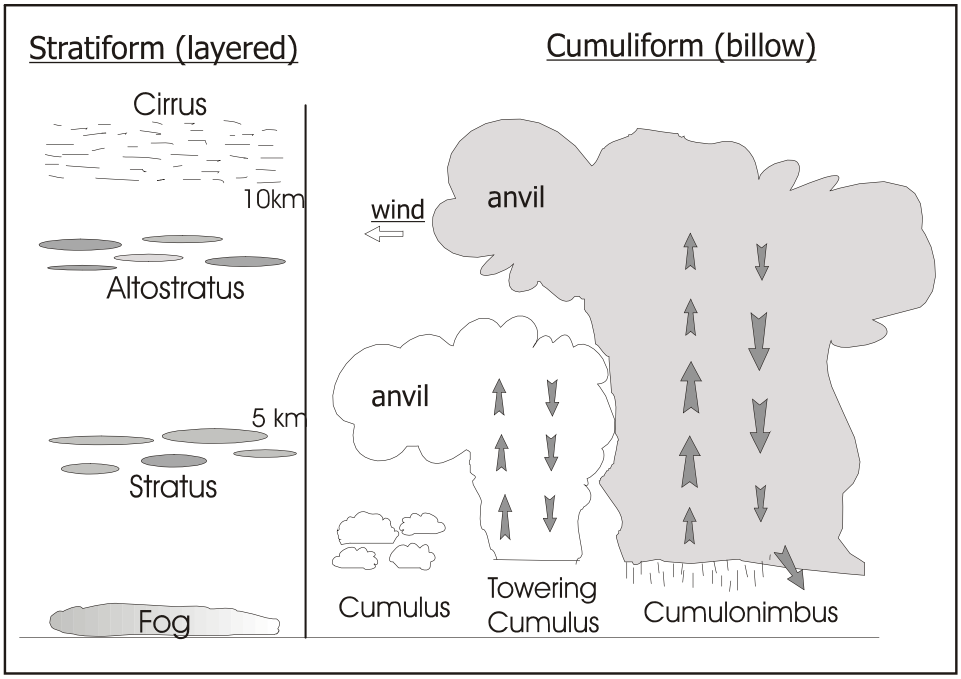


Figure 5.1- Basic cloud types.

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.

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

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**),

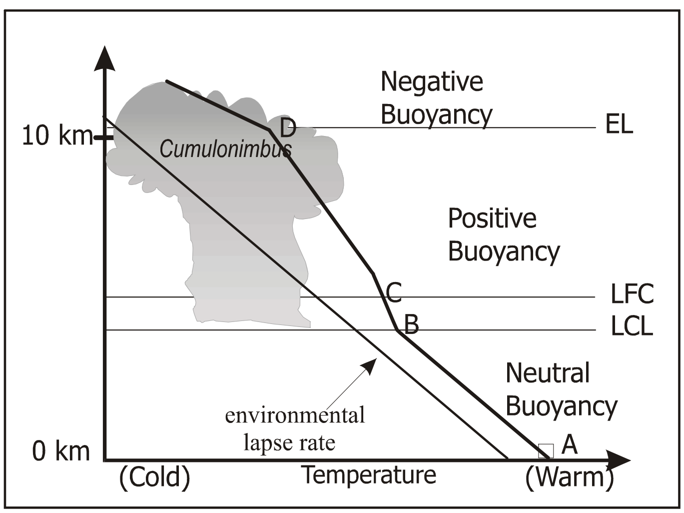


Figure 5.2- Temperature profile for idealized cloud.

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.

<u>Stratiform Clouds</u>

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.

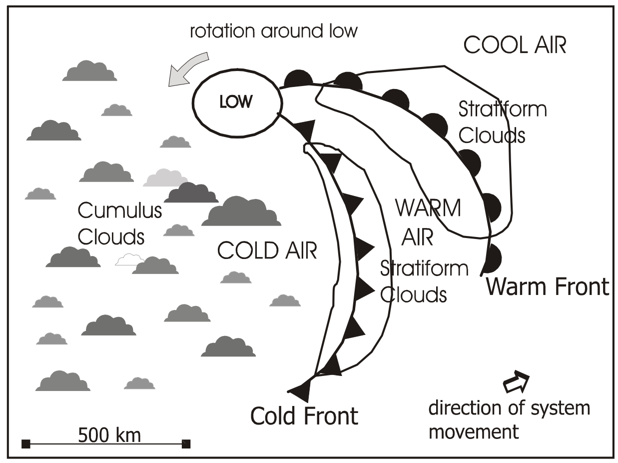


Figure 5.3- Plain-view of low pressure system with its fronts and associated clouds.

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.

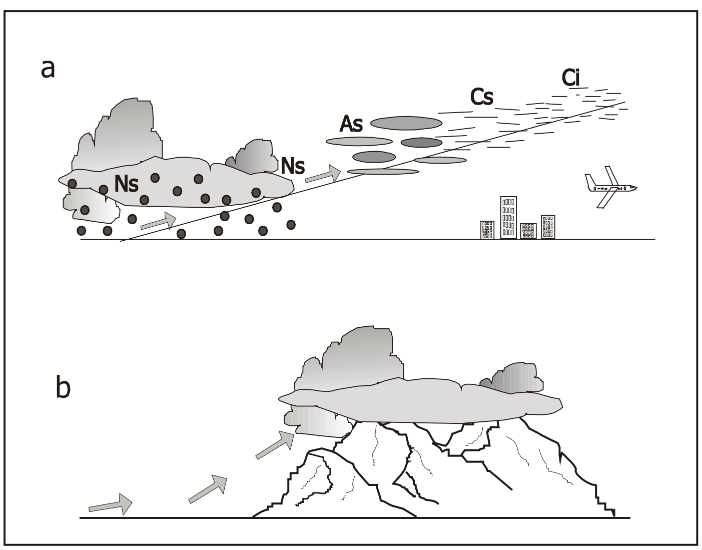


Figure 5.4- Two ways in which air is lifted to create clouds: (a) frontal (b) orographic.

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.

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.

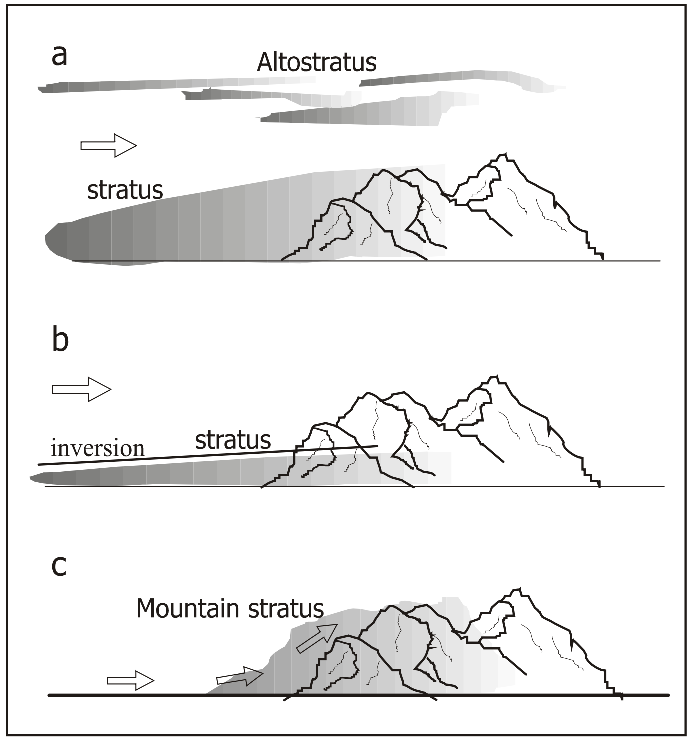


Figure 5.5- Three stratus cloud configurations.

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.

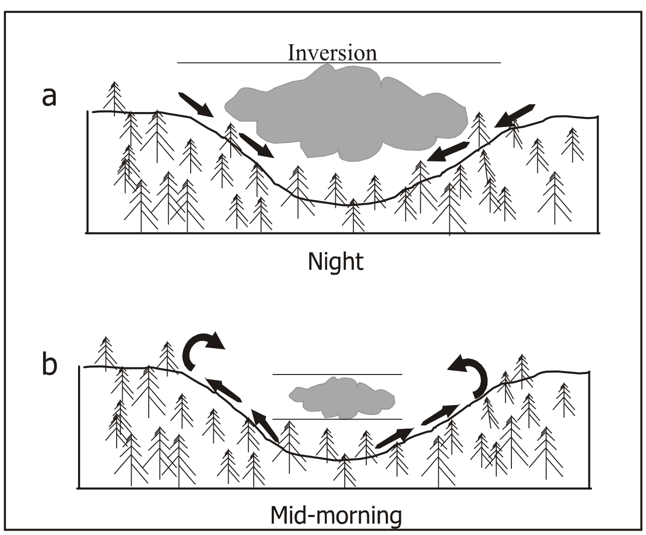


Figure 5.6- (a) Stratus cloud forms below inversion during the night. (b) Cloud dissipates as dry air from above inversion mixes moist valley air.

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.

Banner Clouds

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.

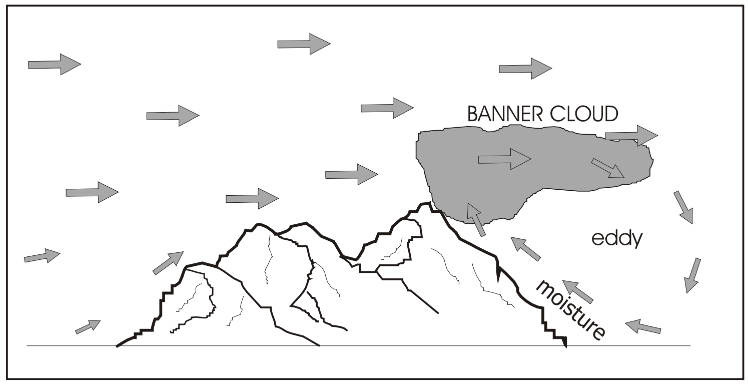


Figure 5.7- Banner cloud forming downwind of the mountain.

Banner clouds form quite frequently over some of the higher Himalaya peaks as a result of strong sub-tropical jet stream winds.

Wave Clouds

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.

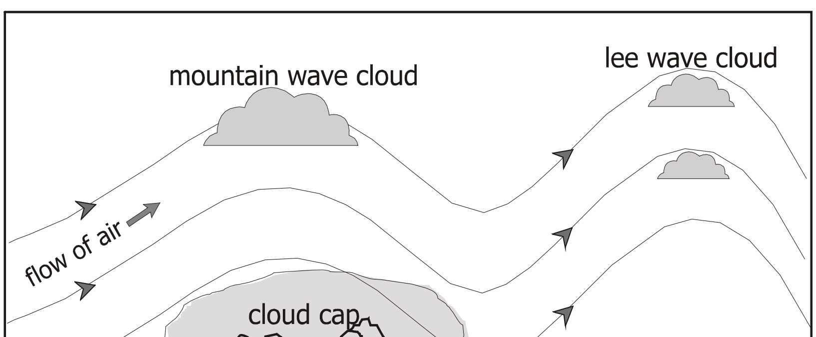


Figure 5.8- Three types of mountain wave clouds that fall under the generic category of lenticular clouds.

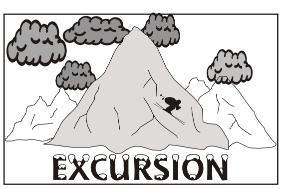
The typical Elk Mountain cloud cap ranged in thickness from 200 -2000 m (660-6600 ft), while the leading edge was located some 2-7 km (1-4 mi) upstream of the summit. The residence time for air parcels (the time it takes air to move from the upstream edge through the cloud to the downstream edge), ranged 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.

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.

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.*

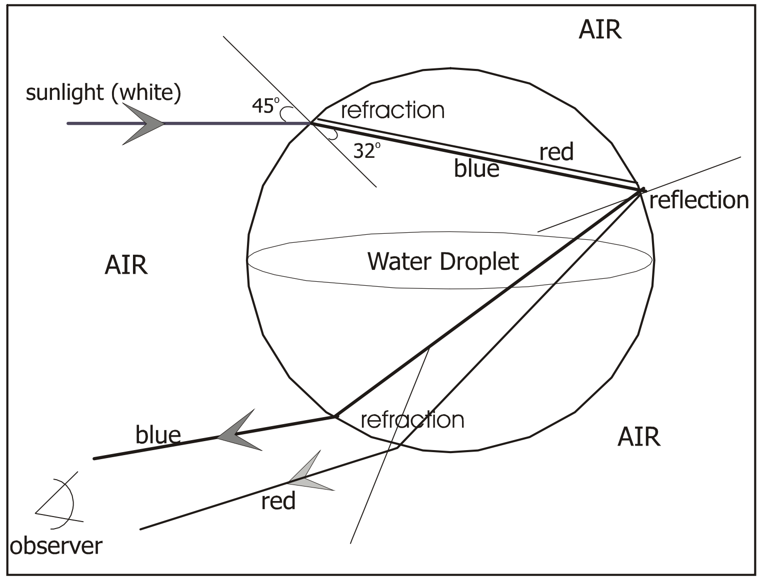


Figure 5.9-- Separation of colors as sunlight travels through a water droplet.

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 which have a tendency to 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.

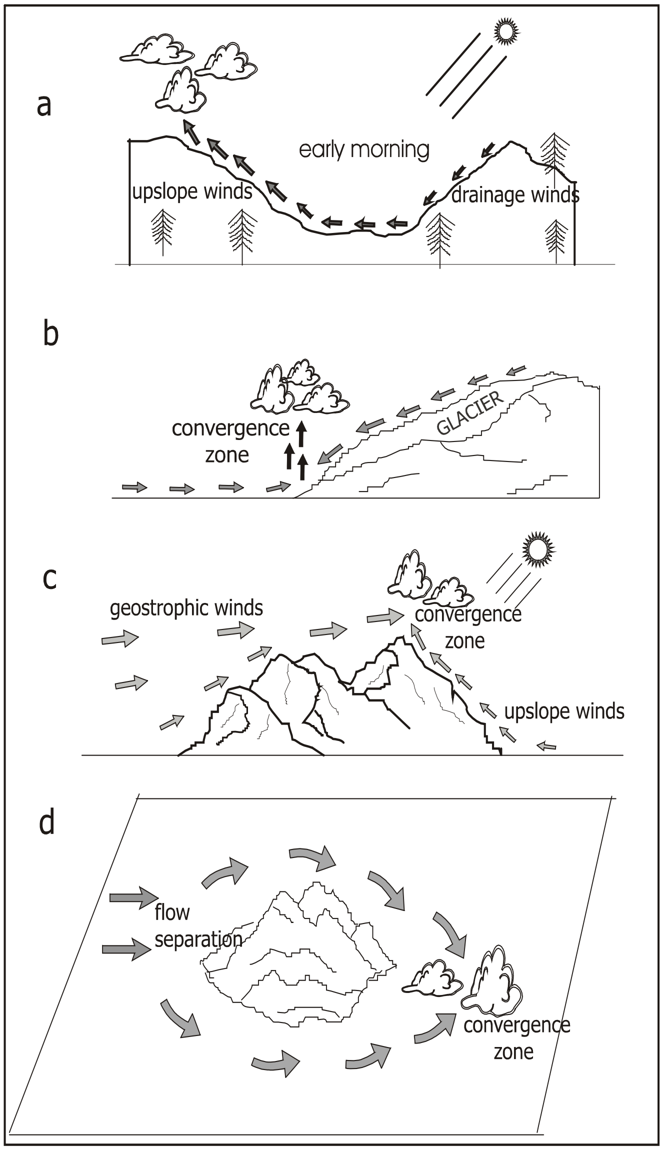


Figure 5.10- Development of cumulus clouds over various types of terrain.

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.

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.

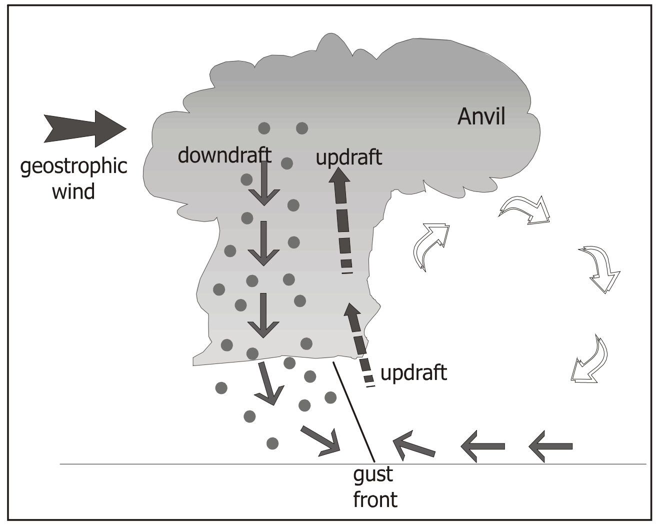
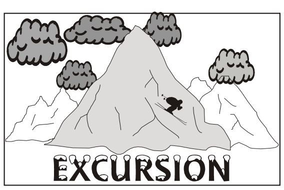


Figure 5.11- Cumulonimbus.

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.

<u>Hail</u>:

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^o to -40^o C (-22^o to -40^o 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.

<u>Lightning</u>

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^o C (32^o 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*.

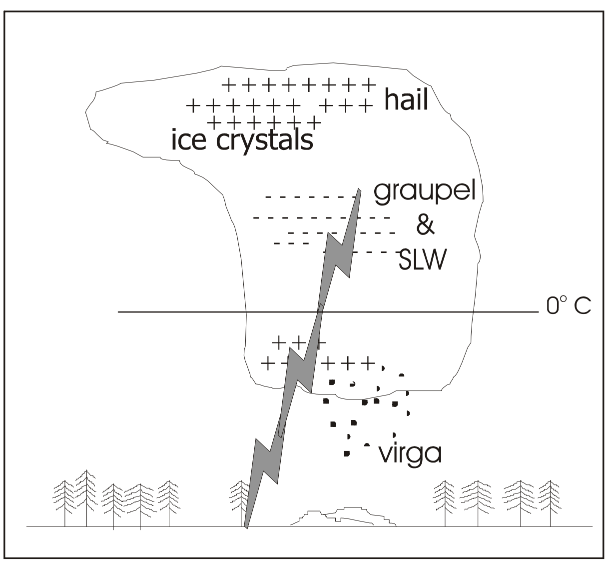


Figure 5.12- Cloud electrification. + sign indicates areas of the cloud which are positively charged, while - indicates negatively charged regions.

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.

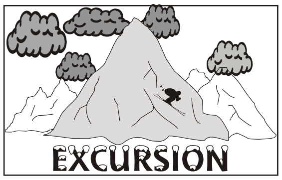
3. As large amounts of negative charge move from the cloud to the ground, a visible channel of lightning known as the return stroke is initiated. This bright discharge occurs so fast that an observer typically sees one simultaneous flash. Temperatures within the conduit can reach 33,000^{o} C.

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.

<u>Survival Tip</u>: 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.

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.*

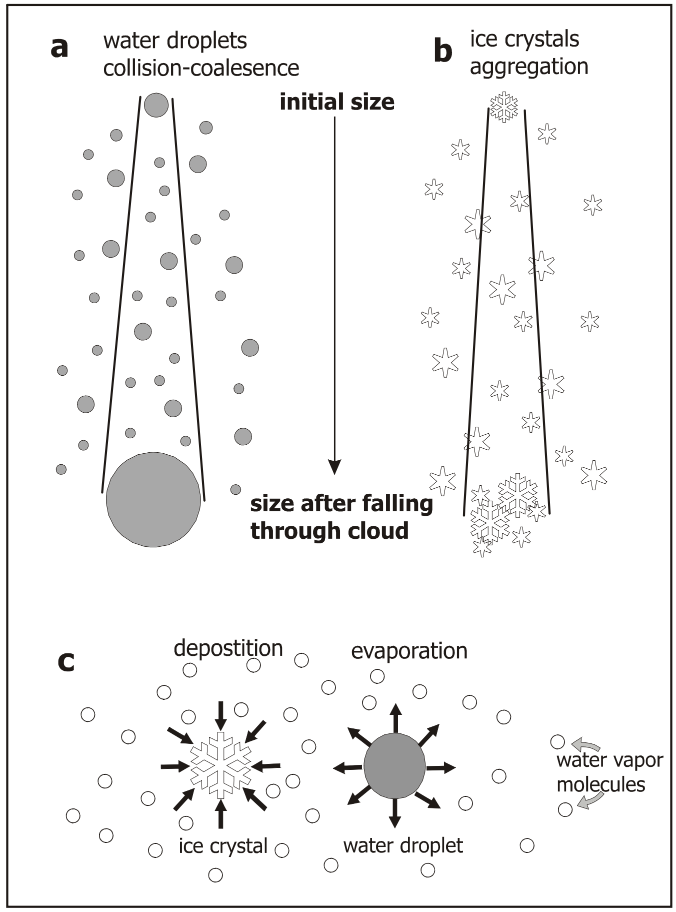


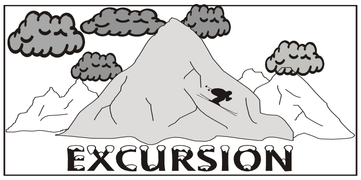
Figure 5.13- Growth of precipitation.

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

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-feeder 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.

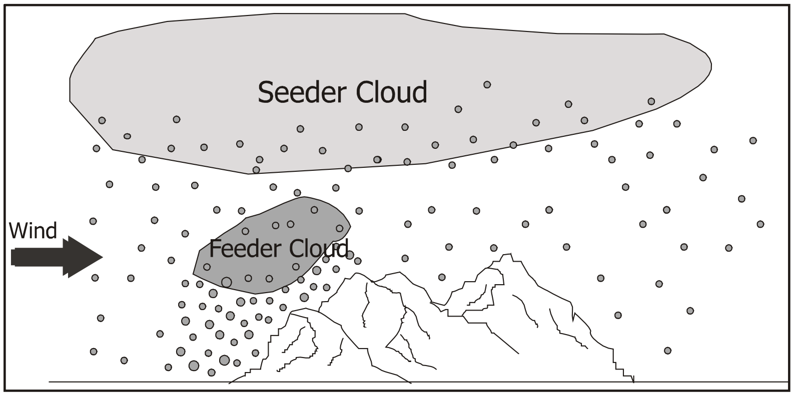


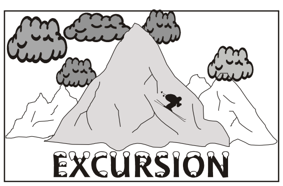
Figure 5.14- Seeder-feeder conceptual model for the enhancement of precipitation.

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.

<u>Convective orographic precipitation</u>

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 (10º N to 10º S) and tropical (10º N/S to 25º 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

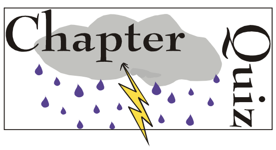
Table 5.3 is a cloud chart that summarizes cloud types and some of their attributes.

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.



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 ____________?

6

WEATHER FORECASTING: WHAT YOU NEED TO KNOW

Chapter Highlights:

- ✓ Discover weather tools- radar, computer models, upper air soundings, satellites.
- ✓ Learn how forecasts are made.
- ✓ Climate demystified.
- ✓ Web resources.

The first part of this chapter looks at the decision making process that a meteorologist uses to produce a forecast. In addition, we will give a brief overview of the various weather tools that are currently in use. The goal of this chapter is to show the strengths and weaknesses of the forecast process, in so doing it will give you some appreciation of what is involved. Secondly, it will teach you how to evaluate the forecast products that you might use. With the accessibility of weather related products on the Internet, some readers are already starting to look at model output, satellite imagery and surface observations before they venture into the mountains. Frankly this is the way it should be done, mountain travelers taking responsibility for their own comfort and safety. The caveat to this however is that just because a person regularly looks at weather maps, does not mean that they know how to forecast the weather with any skill. Here is an analogy we like to use: not everyone who carries a ice tool knows how to use it properly, let alone climb a waterfall. When it comes to weather forecasting, it takes a number of years of experience with many failures and successes along the way, in order to become proficient at it. Even then, weather forecasting is far from an exact science.

Tools of the Trade

In the following section an overview of five different forecasting tools will be presented These tools range from low-tech to high-tech, some of which are only a few mouse clicks away from your own monitor.

<u>Surface Observations</u>

Surface observations are the ‘tried and tested’ tool of forecasting.   You may be asking why would a forecaster want to spend time looking at current conditions when they should be concerned with what is going to occur in the future? The answer is pretty basic, what occurs upstream frequently moves downstream. For example, if observations indicate that it is snowing along a line from Salt Lake City to Flagstaff, then there is a reasonably good chance that this same storm will shortly move into western Colorado as well. Conversely, if it is snowing in western Colorado, it does not mean that it is going to snow in Salt Lake City, because most synoptic-scale storms move from west-to-east. We can sum this up by stating: "If you want to know what will happen in the future, you need to understand what is happening in the present." In other words, weather that is occurring at the present time <u>may</u> continue for some time to come (persistence). It's the job of the forecaster to figure that out.

Surface weather observing stations routinely measure temperature, wind speed, wind direction, dew point temperature (from which RH is calculated), horizontal visibility, cloud heights as well as sky coverage (percent of sky covered by clouds), and precipitation. Surface weather maps, which are constructed from hundreds of surface observations, are an important forecast tool because they indicate the position of highs and lows as well as frontal boundaries. Virtually all primary weather stations in the U.S. use fully automated surface observing systems (called ASOS). These weather stations take sensor readings every few seconds, however the 'official' observing time is ten minutes before the hour. This means that the values that are displayed for a particular observation, may or may not be representative of what occurred thirty or forty minutes earlier. Consider how ASOS calculates wind speeds for example; the wind sensor measures the speed of the wind every second, and then computes a 5-second average. The wind speed displayed at the official observing time is a two-minute average of the 24 previous 5-second averages. For example, if the time of observation is 10:50 AM, then the displayed wind speed is an average from 10:48 AM to 10:50 AM. A wind gust on the other hand is the single highest 5 second average that occurred during the ten minute period prior to the observation. The peak wind gust is the highest 5-second value that occurred during the last hour. If weather conditions are rapidly changing, ASOS stations are programmed to take extra observations in addition to those that are routinely taken at ten minutes before the hour.

Air temperature is calculated using the average temperature in the 5 minute period preceding the time of observation. Precipitation data is based on the amount that has fallen into the rain gage over the previous hour, unless otherwise noted. Be aware that most NWS forecast offices report a midnight-to-midnight precipitation total in their routine weather forecast products. If precipitation occurred in the form of snow, then a total depth and snow water equivalent (the amount of water resulting from melting a sample of snow) are reported as well.

There are thousands of secondary weather observation sites (cooperative observers) around the USA which use human observers exclusively. The types of data collected by the cooperative observers varies from one site to another, however they typically record: daily maximum and minimum temperatures, temperature at the time of observation, and a 24 hour precipitation total as well. Most of these sites unfortunately do not record any wind data. These cooperative observers play a key role in providing weather and climate information that supplements the automated sites. In many rural areas the cooperative observers are the sole source of weather data.

<u>Radar</u>

Radio detecting and ranging (radar) has been used in weather forecasting since the 1950's. Currently most of the contiguous USA is covered by the NWS radar umbrella, with the exception of some regions in the mountainous west and parts of Alaska. Radar is a useful tool because it indicates the location and height of clouds, the speed and direction of cloud movement, as well an <u>estimate</u> of lower and middle tropospheric wind speeds. Radars transmit radio waves that are reflected from cloud droplets and ice crystals. When the return signal is compared to the original signal, the computer processor is able to determine the relative strength of the signal.

A strong signal indicates a cloud that contains a large number of droplets and or ice crystals, while a weak signal indicates a cloud that is not as dense. The return signals are assigned values and colored for display. The limitations of these radars is that they scan a limited volume of the sky, in addition, their range is limited to a distance of about 275 km (170 mi) from the transmitter. You should note that as the original signal leaves the radar, it travels on a slant path that increases in height as it moves away from the transmitter. In fact weather radars transmit radio waves at a number of different path angles with respect to the ground, as illustrated in Figure 6.1. For example, at a distance of 100 km (61 mi) from the transmitter, the signal that is closest to the ground is already about 1240 m (4,050 ft) above the elevation of the transmitter, and at 200 km (122 mi) it is 3310 m (10,860 ft). This means that radar signals frequently over-shoot low clouds. As a result, radars are of little use during episodes of fog and in situations dominated by low stratus clouds.

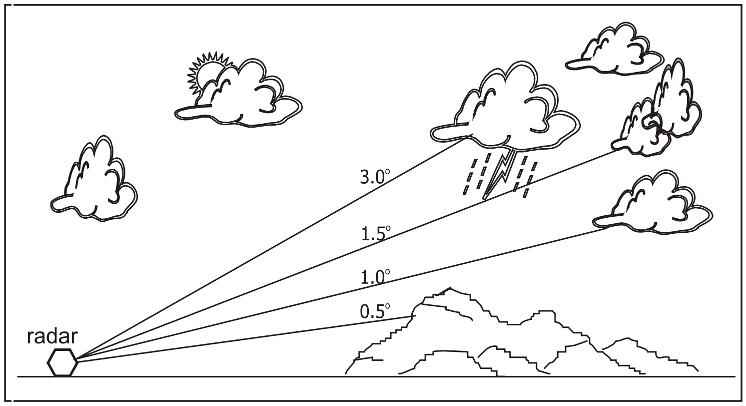


Figure 6.1- Radar elevation scans (exaggerated). The radar rotates through 360° in about six minutes. Notice how the lowest scan is blocked by the mountains.

Another radar limitation is the fact radio waves are blocked by mountains. A number of schemes have been developed in attempt to compensate for this, however the bottom line remains that radar coverage in the western USA is far from complete. In addition, radars are used to estimate rainfall rates, they do not measure precipitation directly. The return strength of the returned signal and the height of the cloud are compared with observed data to derive rainfall rates. It is important to realize that due to the aforementioned limitations, radars are primarily used for short-term forecasting (called 'nowcast'). One of the primary reasons that the weather radar network exists in the U.S. is for the 'early' detection of what is termed server weather- thunderstorms and tornados.

Satellite Imagery

Photos (i.e.-'images') taken from satellites have been used in weather forecasting since the early 1960's. There are many different types of satellite systems currently in use, however, they are usually grouped by either the type of senors that are on-board, or by the type of orbit the satellite is in. A single satellite can have a number of different sensors. These senors are like very expensive cameras that take snapshots of the atmosphere, recording electromagnetic radiation over three prominent wavelengths: infrared, visible, and microwave. The reason we use these different wavelengths is because each one is capable of viewing different phenomena in the atmosphere. For example, microwaves can penetrate clouds, while infrared wavelengths are used to observe and track clouds at night. Visible images are good for tracking low -clouds and fog. Besides tracking cloud movement, many different products are generated from satellite images, some of the more common ones are: wind speeds, areal extent of snowcover, estimates of the amount of water vapor in the atmosphere, and sea surface temperatures. Satellite images do give a glimpse of future weather, however, the interpretation of the images is the difficult part. Just because a satellite image shows a mass of clouds located 700 km (430 mi) off of the Oregon coast, does not necessarily mean that it is going to start to rain in the Cascades in 36 to 48 hours from the time the image was captured.

Satellites are placed in several different kinds of orbits: a geostationary orbit is one in which the satellite remains fixed with respect to a point on the earth's surface. Other satellite's are in a low altitude (400-1000 km) orbit in which it moves either from pole-to-pole (polar orbit), or at some oblique angle across the earth's surface. These satellites travel at a high rate of speed so they can complete an orbit around the earth in about 90 minutes.

Misc. Tools

There are several additional tools in a forecasters arsenal that are worthy of mention. First and foremost are the twice daily launches of the weather balloons which occur at the same time (00Z and 12Z) all over the world [Note that 12Z means: twelve hours Zulu, which is equivalent to Greenwich Mean Time]. The information gathered from these balloons is either called upper-air data or sounding data. The number of weather stations which release balloons globally ranges from 300 to 400. The distribution across the planet is not uniform however, Africa for example does not have the same density of stations as Europe, nevertheless there are balloons launched in some pretty remote locales. The data collected by these balloons forms the backbone of the data sets used to initialize computerized weather models. In addition, a forecaster might want to examine balloon data from a nearby station in order to look at temperature inversions or the change in wind speed and direction with height.

In a rather unique program, a number of commercial airlines have equipped some of their aircraft with special meteorological instruments that transmit data back to a central receiving station in near real-time. Also, in one of the promising new technologies that is being developed, light weight radio-controlled aircraft will be used to fly over remote parts of the world where there is no balloon data. These flying meteorological weather stations are capable of measuring winds, temperatures, relative humidity, and air pressure. This new technology will help improve collection of weather information over the data sparse oceans.

Computer Models

The use of computers to solve the fundamental equations that govern the dynamics and thermodynamics of the atmosphere is called Numerical Weather Prediction (NWP), and has been around in some form since the late 1960's. There are many different types of NWP models, they can be grouped based on: the size of the area covered by the model (know as the domain), distance between grid points (grid interval), and numerical schemes, to name a few. However, the most common classification is based on how far out into the future the model is run. We currently have: nowcasts (0-6 hours), short range (6-48 hours), extended (2-5 days), long range (5-10 days), and climate models.

These models are started, or what we call 'initialized', with data that represents current weather conditions. They are then allowed to run for a given time, during which they output data at specified model times (like every 3 or 6 hours). Below is a short overview of model characteristics, note that modifications and refinements do occur from time-to-time. In addition, as the speed of computers continues to increase, many models are run with smaller grid intervals (higher resolution). Also note that due to the fact the earth is a sphere, grid intervals vary from one point on the globe to another, therefore the grid intervals listed below are approximate values. Most NWP models are at a minimum run at 0Z and 12 Z because these are the two times per day that balloon data is collected. A good web-site that describes model characteristics can be found at: www.nco.ncep.noaa.gov/pmb/products

* Eta: ('a-tuh') The name stems from the type of vertical coordinate that is used. There are a number of different Eta grids, several of which are considered mesoscale. As of the summer of 2002, the Eta was being initialized at 00Z and 12Z, after which the model is run out to 60 hours. There are shorter run cycles at both 06Z and 18Z. This model will continue to be modified and run at higher resolutions. The domain covers all of North America and parts of the Pacific Ocean.

* NGM: The Nested Grid Model has been around for many years and will slowly be phased out over the next decade. The grid interval is about 70 km at 40º N. It is run at 00Z and at 12Z. Domain covers North America and most of the Pacific Ocean.

* RUC: Is a hybrid model that is primarily used for short-term forecast (0-12 hours). The name means Rapid Update Cycle and the grid interval is approximately 40 km. This model incorporates alarge amount of 'non-standard' data for its initialization. It has gone through considerable refinement since its inception, and will continue to in the foreseeable future. Current domain covers continental USA and southern Canada.

* AVN/MRF: The Aviation and Medium Range Forecast models differ from the previous three in that the domains covers the entire Northern Hemisphere. The AVN is displayed as the first 84 hours of the 240 hour MRF. There are some additional differences that are not worthy of mention at this point. The approximate grid interval is 75 km. The AVN is run at 00Z, 06Z, 12Z, and 18Z, while the MRF is run daily at 12Z.

* NOGAPS: This is a USA Navy model that has a grid interval of about 85 km. The domain covers the entire globe. The run times are 00Z and 12Z at which time the model is run out to 144 hours. NOGAPS stands for Navy Operational Global Atmospheric Prediction System

* MM5: This is a limited area model that is being run by a number of different research groups and agencies around the world. The name means Mesoscale Model version 5.0. The domain size and location varies from one application to another but typically consists of three or four nested grids with the smallest grid interval on the order of 4-10 km.

* ECMWF: Is a Northern Hemisphere model that has a grid interval of 60 km. It is run once per day at 12Z. The ECMWF acronym represents the European Center for Medium range Weather Forecasting, which is a consortium of 21 European and Mediterranean countries.

NWP models consist of a series of equations that are solved at individual points within the area covered by the model (Figure 6.2). These models do include mountainous terrain as well as large lakes and oceans. The typical model atmosphere consists of 30 or 40 layers, which extends from the surface of the terrain into the middle stratosphere. The model equations estimate future values of atmospheric variables such as air temperature, pressure, geopotential height, water vapor, wind speed and wind direction to name a few. The equations are solved at every model time-step, which varies from model to model but is within the 10-60 second range. Model output consist of numerical values of each atmospheric variable at each grid, which are then run through a graphics program that converts them into contoured maps of temperature, wind speed, or geopotential height. Output is often displayed on a given level (i.e.-surface, 500 mb), but vertical slices are equally important as well.

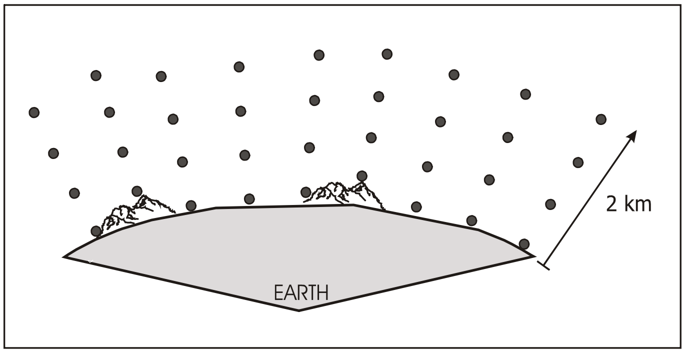


Figure 6.2- Idealized example of a NWP model grid, where the horizontal scale has been contracted. The dots represent points where the model equations are solved at every time step.

NWP models do have a number of limitations, for example, the more grid points contained within a model, the longer it takes the model to run before it can produce any output. Most NWP models have horizontal grid intervals that range from 80 km to about 20 km (49-12 mi). Grid intervals are important because they determine the resolution of the model, not only of atmospheric phenomena, but of the underlying terrain as well. With a grid interval of 40 km (24 mi), for example, the mountainous terrain of the western USA is poorly resolved. In fact to portray the terrain of the western USA realistically, the grid interval should be on the order of 5 km (3 mi). There is a trade-off between a model that covers a large area and has a grid interval of 80 km (49 mi), and a second model that covers a much smaller area but as a grid interval of 20 km (12 mi). Both of these types of models play a role in weather forecasting.

Overall, NWP models simulate synoptic-scale phenomena quite well, but do not perform consistently on the smaller-scales. Likewise, NWP models do a better job of simulating weather in the middle and upper troposphere, than they do near the ground. What this means to the forecaster is that they have to mentally adjust the model output so that it reflects smaller-scale effects, especially in mountainous regions where model terrain is often only a rough representation of the real terrain. If you spend any time looking at NWP output, you will notice that there can be considerable disagreement among the various models. In order to understand what a range of possible solutions looks like, an ensemble approach is used at times. For example, let's say we plotted on one graph the position of the 5300 m height field as it crosses North America (approximately the 500 mb level), from all the available NWP models. This graph, besides having a lot of lines on it, would represent a range of the most likely positions of the 5300 m height. If the lines on this graph are widely spaced there is little model agreement, conversely if the lines are very close or superimposed on each other, there is good agreement and the forecaster can proceed with the confidence that all of the models are predicting the same thing (at least for that particular variable).

In the U.S. the division of the National Weather Service that develops and runs NWP models is called the National Center for Environmental Prediction (NCEP---www.ncep.noaa.gov). Other countries have similar agencies. The NWS is not the only player in numerical weather prediction however, each branch of the military runs their own suite of models, as does numerous research groups. Since the mid-1990's a number of mesoscale or limited area models have been turned into operational NWP models. These models have the highest resolution of any of the current NWP models. In the USA the most common limited area model is MM5. The true beauty of limited area models is that they can run on small computer platforms and they can also be set-up to simulation atmospheric flow over any portion of the earth's surface, as long as there is data available for initialization. The caveat of these models is that higher resolution does not guarantee a better weather forecast. The limiting factor is the low resolution data used to initialize these models. The old computer adage: if you put junk in, you get junk out, certainly applies in this case.

Climate models on the other hand, which are not actually considered a part of NWP because they are not used in operational weather forecasting, are important research tools. Climate model simulations can span decades or centuries. The reason researchers would want to run a climate model is to investigate what repercussions a change in a particular climate element, like the doubling of CO_2 for example, has on the global climate.

Anatomy of a Forecast

In this section we will outline the steps that a forecaster goes through in order to construct a forecast. If you want to develop your own forecasting skills then you should consider using these steps as a guideline. One important point to keep in mind is that a professional forecaster has access to more data and more tools than you will ever have, not to mention many years more experience. Your goal should not be to try and out forecast the professionals, rather your efforts should go into understanding what kinds of mountain weather are produced by the various types of synoptic-scale weather patterns. In addition, in time you should attempt to understand what factors influenced the forecaster, how confident is the forecaster in their assessment, and what could cause the forecast to go wrong. These are not easy to detect, however, you can get a feel for a forecaster's confidence by the terminology they use: "rain likely" versus "chance of rain" for example indicates that there is some doubt in the mind of the forecasters to whether it is going to rain or not. Here are some forecasting tips.

Step One: Consider the current and past state of the atmosphere

This is important because weather is a sequence of atmospheric events that occur in succession. This involves looking at a surface weather map to see where the lows, highs, and fronts are located, and where they have moved from over the past 6-12 hours. It also includes an examination of surface winds, temperatures and precipitation for stations in and around the forecast area.

Step Two: Consider the future

Given the past and current state of the atmosphere, it is time to consider what is going to occur. There are two fundamental methodologies which are used: 1) Given the current conditions the forecaster uses their knowledge of past events that had a similar characteristics, to extrapolate into the future (pattern recognition). Here is a hypothetical example, a cold front has moved into the Wasatch Range, where moderate snowfall has been reported over the past 4 hours. A forecaster in Salt Lake City knows from past experience that the movement of cold fronts is often impeded by the steep terrain of northern Utah. Therefore, the forecast may reflect an additional 3 hours of moderate snowfall in the mountains, followed by diminishing snow showers in the following 6 hour period. The forecaster has now formed a pre-conceived model of what they think is going to happen, even before consulting the NWP models.

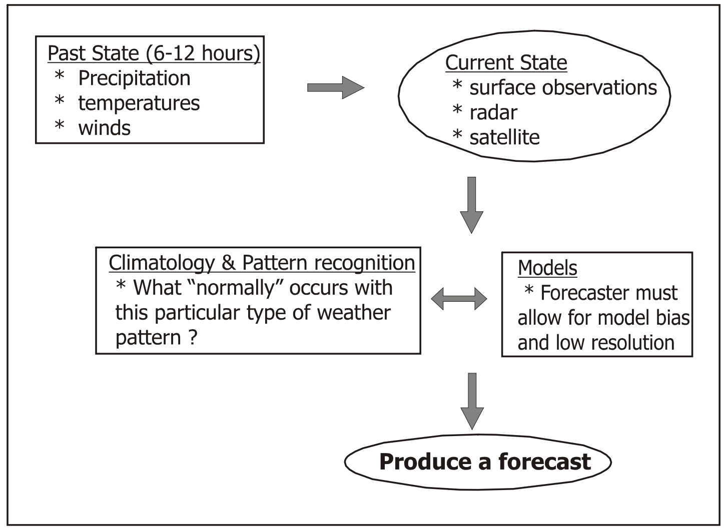


Figure 6.3- Weather forecasting decision flow chart.

2) The second method involves examining NWP model output. In a forecast office a meteorologist will look at all of the available models for a given time period, and pick the one model which seems to have initialized the closest to reality. Note that a surface pressure map is drawn every six hours by the NWS which then can be use to compare the models against. In addition, an important consideration is how well the various models have preformed over the past several days. In other words, identifying which models have a good 'track record' and which ones do not. Once a model has been selected, a forecaster will typically look at least at the following fields:

- surface: pressure and winds
- precipitable water
- 850 mb: geopotential height field, temperature, relative humidity
- 700 mb: geopotential height field
- 500 mb: geopotential height field, vorticity, temperature advection
- 300 mb: wind direction and wind speed (jet stream)

The questions that the forecaster needs to answer are: •where is the jet stream positioned? •Where are the mid-tropospheric and surface highs and lows going to move to during the forecast period? •Are any major changes in the temperature field going to occur during the forecast period? •Where is precipitation going to occur? After the forecaster has familiarized themselves with the model output, they will often refer back to their pre-conceived model and make any necessary modifications. It is at this step that the forecaster takes the model's synoptic-scale solution and adjusts to the local-scale (Figure 6.3). This includes adjustment for deficiencies in model terrain.

Step Three: Constructing a forecast

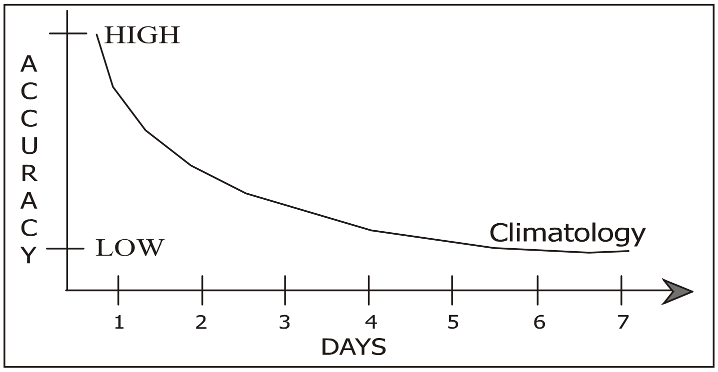


Figure 6.4- Hypothetical curve showing the decrease in forecast accuracy as time increases. Climatology refers to the longterm average weather on that particular day.

Forecasts are written for different time periods and for different areas (often called zones). The primary elements in a forecast are: high and low temperatures, wind speed and direction, cloud cover and possibility of precipitation. Other elements are added as needed. As a consumer of weather products keep in mind that from a statistical prospective, the further out in time a forecast is made for, the lower the probability that it will be accurate (Figure 6.4). There is a trend in recent years for weather forecasts to be extended further out into the future. Many commercial and NWS products include a seven day forecast. Our advice to you, is to view these products with a lot of scepticism. Anyone who tries to forecast clouds and precipitation beyond three days is trying to 'snow job' the consumer. It is fine to outline general weather trends in extended forecasts, but the specifics are a big unknown. This does not mean that during certain weather patterns, particularly when a persistent ridge of high pressure dominates the weather, that an extended forecast is going to be a total failure. In addition, temperatures are much easier to forecast than are clouds and precipitation. However, over the course of time you will find that extended forecasts frequently do not verify. In a slightly different vein, there are legitimate climate forecasts, often called seasonal forecast, where the climate guru's have run a number of their models in an attempt to predict climate trends like droughts, significant temperature anomalies, etc.

In order to help you comprehend the intended meaning of forecast products, below is a list of some of the more commonly used terminology .

• Low Clouds- typically used to denote overcast skies (stratus) with cloud bases within about 1 km (0.6 mi) of the ground.
• High Clouds- cirrus, cirrostratus, altostratus
• Marine Layer- stratus clouds that form over the ocean but subsequently move onshore. The term 'Marine Air' is often used to denote cool/moist air that is moving onshore-this air mass may or may contain clouds.
• Fog- a stratus cloud that is in contact with the ground. Can be patchy or extend over a very large area.
• Daily Mean Temperature- the average of the daily maximum and daily minimum temperatures.
• Probability of Precipitation (POP)- the probability that a precipitation gage or gages will receive <u>at least a trace amount</u> of precipitation over a given forecast period. The POP in and of itself says nothing about total <u>amount</u> or <u>intensity</u> of the precipitation that is expected.
• Drizzle- very light continuous rain.
• Showers (snow or rain)- intermittent snow or rain, does not say anything about intensity or amount of precipitation. Can be from stratiform or convective clouds. Often reserved for post-frontal precipitation, but usage may vary from forecast office to forecast office.
• Blizzard- high winds that lift snow from off of the ground producing low visibility. Often occur after

actual snowfall event.

• Freezing Rain- cloud droplets that freezing on contact with the ground, trees, powerlines, etc.
• Flurries- light snowfall but with little accumulation.
• Virga- rain that falls from a cloud (usually Tcu or Cb) that evaporates before reaching the ground.
• Thermal trough/low- A mesoscale area of low pressure that develops due to heating of the ground (and adjacent air) over a period of several consecutive days. Typically occurs during the summer.
• Thunderstorm- Cb that produces lightning, moderate to heavy rain, and quite often hail.
• Advisories, Watches, and Warnings- these three products apply to a wide range of meteorological and hydrological events: heavy snowfall, heavy rain, thunderstorms (hail, tornadoes), high winds, extreme wind chill, to name a few. Different offices of the NWS have different criteria which are used to determine which of these products should be issued, therefore we will only give a generic description.

The difference between an advisory and a warning is the intensity of the event, let us use high winds as an example. If the wind speeds are forecasted to be of moderate intensity but still strong enough to be considered a nuisance, then an advisory is issued If the forecasted winds are going to produce more than minor damage, then a high wind warning is justified. Some offices differentiate between watches and warnings based on how far in advance the product is issued. If high winds are expected to begin 18 hours from the current time, most forecasters would issue a watch, to initially alert the public. During the course of those 18 hours, if the expected high winds turned out to be a sure bet, the forecaster would then upgrade the watch to a warning. The idea behind this strategy is that warnings should only be issued when there is a high probability that it is going to happen. This reduces the number of 'false alarm' warnings that are issued.

Trip Planning and Field Forecasting

Forecasting the weather from a weather office where all of the data is available at a persons finger-tips is one thing, but it is a completely different ball game to do it while hiking over hill and dale. The following are some suggestions of information you should gather before you pull out of the driveway or board the 747 for the Himalaya. We have also included a short list of weather elements that you should pay particular attention to once you are in the mountains.

<u>Local Trips</u>

• The current and forecasted position of the jet stream(s). The mid-latitude storm track follows the jet stream, so knowing its position is valuable.

• Read forecast discussion issued by the nearest office of the NWS to your destination. How uncertain is the forecast? Keep in mind that uncertainty can be a result of model discrepancies, transition from one weather system to another, or due to the simple fact that your destination is going to be on the fringes of a storm.

• You can also hear weather reports and forecast via NOAA Weather Radio. These broadcast are made on frequencies not normal to AM/FM radios, so you have to buy a special radio. Nevertheless, these radios are pretty cheap and can be carried into the backcountry as well. Contact your local National Weather Service office for more information.

• Attempt to find out what temperatures and winds you should expect for a given altitude range. This includes the approximate height of the freezing level.

• Where are the fronts and how are they expected to move. Is there going to be any stratiform precipitation? Has it been convective for the past several days? What are the possible weather threats: lightning, high winds, extreme cold, etc.

• When you arrive in the mountains, observe the current conditions. Do these observations conform to your conceptual model of what the weather should be? If you answer NO- reevaluate. Is the weather better or worse than expected? If it is considerably worse, you may need to alter your plans, or be prepared to alter plans as the day progresses.

• Once you have returned home, it is important to compare the forecast with what occurred. If the forecast was a poor, try to find out what went wrong. Also ask yourself what you may have learned about mountain weather while on the trip.

• Several NWS offices produce summer forecasts for a number of National Parks, including those frequented by climbers: Denali and Mt. Rainier National Parks in particular. These forecasts include the usual generic weather forecast, but also contain wind and temperature forecasts at higher elevations. See the respective sections in Chapters 7-9 for more details and web-sites.

Extended Trips–Over Seas

Trying to find weather and climate information on any of the worlds's remote mountain ranges, can be quite exasperating. Here are some hints.

Pre-Trip:

• Read the appropriate chapter(s) on regional weather surveys in this book. Pay close attention to seasonal storm track information. You want to answer the following questions: When is the 'dry' season? What is the seasonal wind pattern? Are there any unique weather phenomena that I should be aware of?

• The best region-by-region description of climate, which covers the entire planet, can be found in a multi-volume work entitled: *World Climatologies*. This monumental work is somewhat dated, nevertheless it is a wealth of information. It can be found in some larger libraries.

• Get on-line with the National Climatic Data Center (NCDC), they also have some data outside of the USA because they are a designated archiving agency within the World Meteorology Organization. In addition, most countries have some type of weather service, which usually archives climate data. Some of these have useful web-sites, see the end of this chapter as well as the weather summaries given in Chapters 7-9 for website listings

• Learn from other climbers, both their mistakes as well as successes. Read books and climbing accounts as well as on-line bulletin boards that cater to climbers. You can also find out when commercial expeditions or guide services run trips to a particular mountain region. Since these services are trying to generate a profit, they generally know when the optimum weather occurs. There are few regions of the world, as you will learn in the following three chapters, where there is no optimum climbing season. In those cases mountain travelers should be prepared for serious weather delays. If waiting out extend periods of stormy weather is not on your itinerary, you should look elsewhere to satiate your climbing and hiking needs.

• Before you leave on the trip, attempt to find out what the weather has been like for the past several months. Has it been snowier, windier, colder than normal? In some regions of the world, the weather can undergo major transitions over the span of a week or two, in other regions the transition is much more gradual.

In the Mountains*:*

Keep in mind that the majority of the time you are not going to have access to a forecast. This is slowly changing as more and more large expeditions carry satellite communication gear. Even then however, most of the forecasts are quite general in nature. Here are the most important weather elements to pay attention to:

- Cloud sequence–In the big mountain ranges there is often rapid cloud development at 'lower' and mid-elevations which proceeds cloud development at higher (summit) levels.
- Changes in atmospheric pressure–Use your altimeter if you have one (see Excursion in Chapter One)
- Shift in wind direction- This may indicate the passage of a front or trough.
- Dramatic change in wind speed- By dramatic we mean a change from light winds to moderate or strong winds over a period of several hours.
- Change in visibility-- this is less reliable than other methods but at times can signify an increase in low or mid-level moisture. Meteorologists call very small non-cloud water droplets: haze. These droplets condense and re-evaporate rapidly, and they generally do not become large enough to form cloud droplets. If for example the previous four days had been very clear, and on day five the winds increase and there is a noticeable layer of haze, it could be a pre-cursor to increased cloud development.
- Increase in upper or mid-level tropospheric wind speeds before the winds at lower elevations increase. This occurs frequently on higher mountains because the summits of these mountains lie close to the jet stream.

If the weather has been good for a period of time but then you start to notice increased cloud development (more than on previous days), or the signs of an approaching front, and/or an increase in wind speeds, you should keep a close watch on the weather. A number of people fail to do this each year, and end-up 'buying the farm' (see Chapter 1). Conversely, if the weather has been stormy for a period of time, and the winds begin to diminish, and the clouds thin out, there is a reasonable chance that the storm is in its final stages. Beware of 'sucker holes' however, these are lulls in a storm that lure climbers away from the safety of a camp, only for the storm to re-intensify.

Professional Profile
An Interview with Jim Woodmencey

Jim is a meteorologist and owner of MountainWeather (www.mountainweather.com), a weather consulting business based in Jackson, Wyoming. He is the on-air meteorologist at KZJH Radio and the author of the book: Reading Weather. Jim has also worked as a climbing ranger in the Tetons and as a heli-ski guide.

Q1- Is the average backcountry skier weather cognizant?
JW- I would say the average backcountry skier is not that weather cognizant, but the more experienced skiers are. Skiers who have been through even a basic avalanche course understand the importance of weather and its affect on the snowpack and avalanche conditions.

Q2- Do monthly or seasonal changes in the weather factor into the formation and release of avalanches?
JW- In the early season (October through December in the Rockies), the snowpack that already exists is greatly affected by longer-term weather patterns. For instance, a long dry spell with clear skies and cold temperatures can weaken the existing snowpack and set up a dangerous avalanche situation later in the winter as snow eventually accumulates on top of this weaker base layer.

As warmer weather approaches in the late winter and early spring the snowpack will tend to strengthen through settlement and consolidation. However, a warm-up that occurs too rapidly can be responsible for widespread avalanche release.

Q3- How much of a threat is lightning to mountain travelers in the central and northern Rockies?
JW- Lightning is perhaps the most significant weather threat that mountain travelers have. Too little attention is paid to approaching and developing thunderstorms by folks in the mountains. A few important facts about lightning are worth remembering. Lightning is five times more likely to strike in the mountains than it is nearby valleys. In addition, lightning has been known to strike 'out-of-the-blue' from 5 to 10 miles away.

Q4- What role do you see the Internet playing in the dissemination of weather/climate information to mountain travelers?
JW- Next to watching The Weather Channel, the Internet is the world's most accessible source of weather information. The problem for the user lies in finding relevant forecasts for the higher elevation locations they are traveling to. An additional complication is the user being able to interpret satellite, radar and computer model products without the training and experience of a meteorologist. With the recent computer and communications technology advances, using computers in the mountains to receive and analyze weather information on–the-spot may become more commonplace (Jim provided forecast for a spring 2000 Everest expedition).

Although, despite all of the new technology, it is my opnion that the NOAA Weather Radio is *still the cheapest, lightest, and most portable way to get update forecasts in the mountains throughout most of the USA.*

Climate: What is it?

Climate can be defined as: the long-term average weather. This simple definition is however, in need of further clarification. When the term 'weather' is used, what we really mean are weather elements that can be measured, such as air temperature, rain, snowfall, etc. Essential we take descriptions of the weather ("July was very hot") and turn it into numerical values ("July was 4.6° C above normal"). Secondly, what do we mean by 'long-term'? Unfortunately there is no precise definition, 'long-term' can mean 10 years or 10,000 years, it should be made clear from the context in which it is used.

In most countries around the world climate statistics are, for the most part calculated using 30 years of data. For illustrative purposes let's assume that you wanted to know the mean (average) daily temperature, or what is often referred to as the 'normal' temperature for April 12th in Jackson, Wyoming. Once you have collected about 30 years worth of April 12th data, you would then calculate the mean. The National Climate Data Center (NCDC), which has been commissioned to collect climate data in the U.S., does not recalculate 30 year averages at the beginning of each new year. They use a fixed 30 year period that is adjusted once each ten years. For example, currently we are using the 1971-2000 climate averages. In January of 2011, work will begin to calculate new climate normals for the years 1981 to 2010. Other agencies and researchers may use climate statistics that have a longer than 30 year record, it just depends on the purpose of the study.

In the paragraph above we used the example of daily mean temperature at Jackson, here is an illustration on how it is calculated. If the maximum temperature on a given day at Jackson is 17° C (63° F) and the minimum is 9° C (48° F), then the daily mean is a simple average of these two values or 13° C (55° F). A second but less common way to calculate the daily mean is to average the 24 temperature observations that are taken at the top of each hour. This is not common in climate statistics because the vast majority of climate stations do not record hourly observations.

Most climate stations use special thermometers that record the maximum and the minimum temperature even when no observer is around to make the observation. Another very common climate element is the mean monthly temperature. This value is produced by averaging each daily mean over the course of the month. In a similar fashion an annual mean temperature can be calculated by averaging the 12 monthly values. Table 6.2 displays the monthly mean temperature for Jackson, using both the 1961-1990 and the 1971-2000 averages. Notice how the 1971-2000 averages are up to one degree Celsius warmer than the 1961-1990 averages, this is not surprising since the decade of the 1990's was one of the warmest on record in many parts of the world. The last row in Table 6.2 represent standard deviations for the monthly means. The standard deviation (calculated for 1971-2000) is simply a range of temperatures wherein two-thirds of all the values can be found. For example, the January standard deviation is 2.6° C (4.7° F) while the mean is -8.3° C (17° F). This can be understood to mean that two-thirds of the 30 values (one for each year) fall within the -10.9° C to -5.7° C range (calculated as -8.3° C ±2.6°). Also notice that during the warmer months of the year, the standard deviations are much smaller than in the winter. This is a result of the fact that during the winter, large temperature fluctuations occur because of the formation and dissipation of low-level temperature inversions.

Table 6.2- 30 year monthly mean temperature for Jackson, Wyoming.

years	ºC	J	F	M	A	M	J	J	A	S	O	N	D	**Ann**
1961-1990	ºC ºF	-9.3 15.3	-6.7 19.9	-2.2 28.0	3.1 37.6	7.9 46.2	12.5 54.5	16.2 61.2	15.0 59.0	10.2 50.4	4.6 40.3	-2.3 27.9	-8.9 16.0	**3.3** **37.9**
1971-2000	ºC ºF	-8.3 17.1	-6.4 20.5	-1.1 30.0	3.5 38.3	8.1 46.6	12.9 55.2	16.3 61.3	15.5 59.9	10.6 51.1	4.8 40.6	-2.3 27.9	-8.6 16.5	**3.9** **39.0**
St Dev.	ºC ºF	2.6 4.7	2.5 4.5	2.3 4.1	1.6 2.9	1.2 2.2	1.4 2.5	1.6 2.9	1.4 2.5	1.4 2.5	1.4 2.5	2.1 3.8	3.2 5.8	**0.8** **1.4**

Precipitation statistics are calculated in a similar fashion as temperature, with the notable exception that there is, of course, no maximum or minimum daily precipitation, only a daily total. Most climate precipitation statistics are based on monthly or annual values. Remember that precipitation is a combination of rainfall and the water equivalent of the snow that fell over a given period of time. For example, if 0.7 cm (0.27 in) of rain fell at Jackson during the day on March 31, but an additional 10 cm (3.9 in) of snow fell that night, before midnight. What do you think the daily precipitation would be if the water equivalent of the snow was 0.8 cm (0.31 in)? The daily total precipitation on March 31 would be 1.5 cm (0.58 in). In addition, during the winter months climate stations keep a running total of that seasons total snowfall. Let's say that the seasonal snowfall up to midnight on March 30 had been 161cm (63 in), obviously the new seasonal snowfall total would be 171 cm (67 in) as of midnight March 31.

A host of additional climate variables are compiled by different agencies, some of the more common ones are: thunderstorm days, average daily wind speed, daily peak wind gusts, snowfall, hours of sunshine. The best web-site to view climate data for the western USA is at the Western Region Climate Center (www.wrcc.dri.edu)

Weather and the Internet

With the advent of the Internet, weather information has gone from the realm of being hard to find, to more information and data than you could possibly use. Virtually every office of the National Weather Service has a website where they post forecasts as well as weather discussions, and a hosts of tools such as: satellite images, model data, sounding data, etc. You should be aware that these websites are typically not standardized, meaning one office may have considerably more tools and goodies than another office. In addition, many universities with a meteorology or atmospheric science department will have a website with numerous links. And of course there are commercial ventures that provide weather data either free or for a charge. The following is a list of places to begin your weather search.

General Weather Information:

- Weather Underground: **www.wunderground.com**
- The Weather Channel: **www.weather.com**

- National Weather Service (NWS): 136 forecast offices nation wide, in addition to a number of related research facilities that all have obscure acronyms.
  http://iwin.nws.noaa.gov
  www.nws.noaa.gov
- Environment Canada: **http://weatheroffice.ec.gc.ca/canada_e.html**

Model Output:

- University of Michigan: **www.yang.sprl.umich.edu/wxnet/model**
- NCEP: **www.ncep.noaa.gov**
- ECMWF: **www.ecmwf.int**
- MeteoSwiss (Switzereland): **www.meteoschweiz.ch**
- **http://weather.unisys.com**
- **http://twister.sbs.ohio-state.edu**
- **www.aos.wisc.edu/weather/index.html**
- Navy (use public access button): **www.fnmoc.navy.mil**

Upper Air Data:

- Forecast Systems Laboratory (FSL): **www.fsl.noaa.gov**

Satellite

- NESDIS: www.nesdis.noaa.gov
- Colorado State University: **www.cira.colostate.edu**

Climate Data:

- National Climatic Data Center (NCDC)- Archives data for NWS. It also has limited climate data from around the world. Some of it is free and some cost $.
  www.ncdc.noaa.gov
- Western Regional Climate Center: **www.wrcc.dri.edu**
- Climate Prediction Center (CPC): **www.cpc.noaa.gov**
- Climate Diagnostics Center (CDC): **www.cdc.noaa.gov**

Snow Data:

- National Resource Conservation Service (NRCS)- This is the organization in the USA that is mandated by the Federal government to collect snow data. They operate snow courses and snotel sites (automated snow senors). Water Resources and Climate Center: **www.wcc.nrcs.usda.gov**
- California Snow Cooperative: **www.cdec.water.ca.gov/snow**
- National Snow and Ice Data Center: **www-nsidc.colorado.edu**
- Swiss Federal Institute for snow and avalanche research: **www.wsl.ch/slf**
- Avalanche Forecast Center: **www.avalanche.org**
- Northwest Weather and Avalanche Center: **www.nwac.noaa.gov**

Lightning

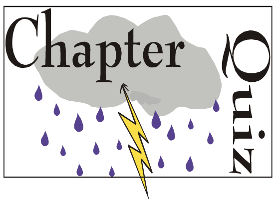
- Pueblo Forecast Office: **www.crh.noaa.gov/pub/ltg.shtml**

Research

- National Center for Atmospheric Research (NCAR): **www.ncar.ucar.edu**

Misc.

- Mt Washington Observatory: **www.mountwashington.org**
- World Meteorological Society (WMO): **www.wmo.ch**



1. True/False: Weather balloons are launched at the same time no matter the location?
2. Name at least two NWP models.
3. True/False: Radar can be used for long-term weather forecasting?
4. What is freezing rain?
5. True/False: Climate data is of little value when you are planning a major overseas expedition?
6. True/False: Lightning is not a hazard in the Rockies?
7. True/False: The mid-latitude storm track is closely associated with the sub-tropical jet stream?
8. What range of wavelengths (visible, infrared, microwave) are used in satellite imagery to view/track clouds at night?

7

REGIONAL WEATHER SURVEY, PART I
Mountains of Alaska, Cascades and Sierra Nevada

Chapter Highlights:

- ✓ Detailed information on the weather and climate of the Alaska Range, St. Elias Mountains, Coast Range, Cascades, and Sierra's.

Alaska

Weather in Alaska is a direct result of two interrelated phenomena: the position and strength of both the polar jet stream and Aleutian low pressure system. In Chapter 4 you learned that over the course of a year the polar jet migrates through a very broad range of latitude. Along the 150^o W meridian (Pacific Ocean) for example, the polar jet can be found anywhere between the Alaska Range (64^o N) and the North Pacific (30^o N). For reasons which are not germane to our present discussion, there are times when the polar jet stream is very strong and other times when it is weak. Occasionally the polar jet stream splits into two parts, one section at high latitudes and the other in the mid-latitudes. During most of the year the polar jet is positioned just to the south of the Aleutian Islands and the Gulf of Alaska. However, when the jet moves north over the state, widespread clouds and precipitation can be expected. This is a very simplified view of the movement polar jet since it is rarely parallel to latitude circles; more often than not the jet contains curves and bends which redirect the flow of air north and south as well. For example, when the polar jet is over the North Pacific it is common for it to cover as much as 30^o of latitude.

The Aleutian low is a seasonal (October-April) center of low pressure that resides near the Aleutian Islands, and is a region of frequent storm generation. The majority of large storms that move into Alaska as well as the west coast of North America, originate in the Aleutian Low. Subsequently, the primary direction from which synoptic-scale storms move into Alaska is from west to south, as illustrated in Figure 7.1. These are the storms that transport large amounts of moisture from the either the Bering Sea or Gulf of Alaska. Storms that originate from the east are much drier, less frequent and less energetic than their western counterparts.

The parameters which influence weather in any specific region of Alaska are latitude, the distance from an ocean, and elevation. Latitude influences weather and climate by regulating temperatures, which in turn restricts the amount of moisture that an air mass can hold. Generally, at higher latitudes the annual temperature is cooler and therefore the annual precipitation is reduced, when compared to mid-latitude sites. This does not mean that daily or monthly averages of temperature and precipitation cannot be higher than locations further to the south. For example, the area around Fairbanks (this includes all of the interior of Alaska) is much warmer in June-August, than it is in Anchorage 450 km (280 mi) to the south. In fact it is considerably warmer in the summer in Fairbanks than it is in any major city on the West Coast. However, since winters are so much colder north of the Alaska Range, the annual temperature for towns like Fairbanks is well below those of Anchorage or Seattle.

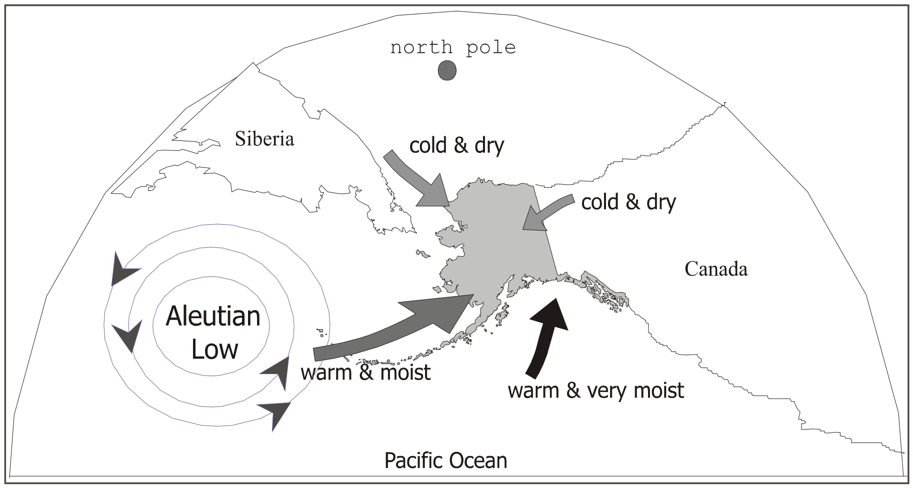


Figure 7.1- Winter storm track for Alaska.

An additional influencing factor is the distance that a particular region is from either the Gulf of Alaska or the Bering Sea. The term that is used by meteorologists to describe this effect is: <u>continentality</u>**.** Since the ocean contains a large volume of water that heats-up and cools-down very slowly, towns located near the ocean, like Anchorage or Valdez, have modest seasonal temperature ranges. In winter for example, the Interior (the area north of the Alaska Range and south of the Brooks Range) is considerably colder (15º to 25º C or 27º to 45º F) than the area along the Gulf of Alaska because of the warming influence of the ocean in the latter region. In the summer, the Interior is warmer and the coastal areas are cooler. Due to the almost year round cover of sea ice, the thermal influence of the Arctic Ocean on the area north of the Brooks Range is limited. In addition, since the storm track is rarely from the north, the Arctic Ocean is not a very big supplier of moisture. Incidently, most of the moisture that reaches the Brooks Range is transported from the Bering Sea or Gulf of Alaska.

Elevation influences weather because in general: both temperature and moisture decrease with increasing elevation. In Chapter 5 it was pointed out that precipitation does not continue to increase with increasing elevation for most large mountains. In the Alaska Range for example, there are very few precipitation gages higher than 1000 m (3,280 ft), therefore our knowledge of precipitation (snow and rain) at higher elevations is based in large part on an educated guess. The following list is an <u>estimate</u> of annual precipitation in Alaska's mountain ranges:

Brooks Range: 0.5-0.8 m (20-31 in)	St. Elias: 3.0 to 6.0 m (110-220 in)
Alaska Range: 0.5 to 2.0 m (19-78 in)	Alaska Coast Range: 2.0 to 5.0 m (78-196 in)

The wide range in annual precipitation in a particular mountain range occurs not only because of the variation in elevation, but also due to the fact that windward slopes intercept the bulk of the precipitation. As a consequence, two sites with the same elevation can have dramatically different precipitation regimes if one site is located on the windward side of the range and the other is on the leeward side. One of the wetter regions in the state is around Mt. Fairweather, where on the windward slopes annual precipitation is estimated to be 7.6 m (300 in). Also keep in mind that precipitation gradients tend to be much steeper on the leeward side of these ranges. For example, on the St. Elias Mountains the annual precipitation ranges from 6.0 m (235 in) on the windward slopes to around 0.5 m (19 in) at the base of the range in the Yukon Territory. This occurs over a horizontal distance of about 150 km (92 mi). As a consequence the western half of the Yukon Territory is in the rain shadow of the St. Elias Range.

Alaska Range--Denali National Park

The Alaska Range is a 1040 km (650 mi) arc of mountains that divides the Interior from Southcentral Alaska. The western section of the Alaska Range is primarily composed of volcanic mountains, while the northern section is of a fault-block origin. The average height varies between 2700-3500 m (8,800-11,500 ft), with a handful of peaks higher than 4000 m (13,100 ft). Most of the Alaska Range can be considered to lie in a continental climate regime, the exception being the southwest corner near Cook Inlet, which is semi-maritime (Figure 7.2, data for April-August). The Alaska Range as does most of Alaska, receives the majority of its precipitation from mid-August through October. Winters are cold with a considerably more cloud-free to partly cloudy days than overcast days. This pattern is broken by the occasional multi-day snowstorm. The northern side of the range is colder than the southern slopes. Climatologically, the months of April, May, and June are the driest, with only about 20% of the annual precipitation occurring during this period. This should not be understood to mean that it rarely rains or snows in the Alaska Range in April through June. In fact most precipitation that does occur in April and May is associated with synoptic-scale disturbances that move over the area.

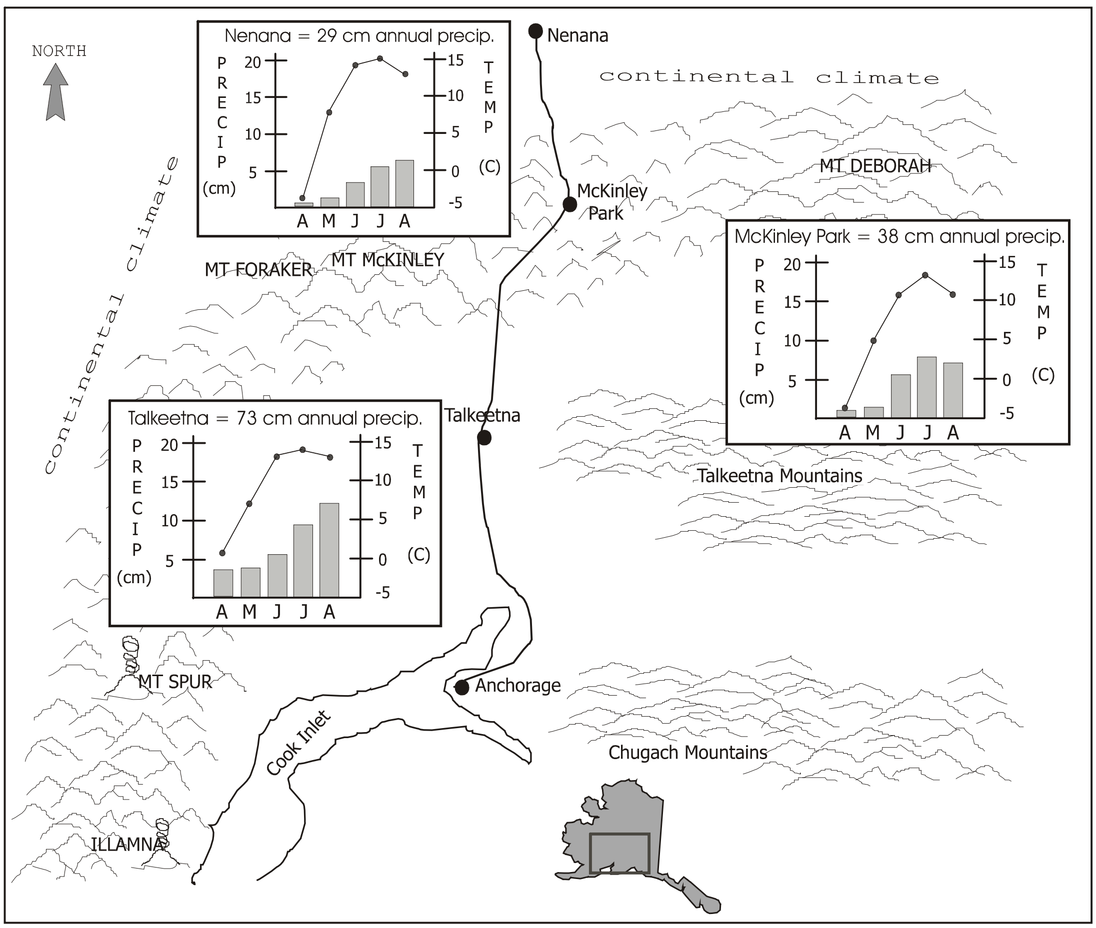


Figure 7.2- Topography and climate of the Alaska Range. In climate boxes, temperatures are shown as a line with the scale on the right axis. Precipitation is represented by the bars with the scale on the left axis.

Climate data from towns such as Talkeetna, Healy, and McKinley Park, show a pronounced increase in precipitation in June when compared with the previous two months. However, this has to be taken in perspective since most of the summertime precipitation falls in the afternoon and evening from convective rain showers. These showers can produce several centimeters (1 in) of rain at lower elevations, and 10-20 cm (4-8 in) of snow at higher elevations in the Alaska Range.

Since the majority of climbers in the Alaska Range are attracted to Denali National Park, most of the remaining material in this section will focus on Mt. McKinley weather. With 4000+ m (13,100 ft) of relief, Mt McKinley and Mt. Foraker are major orographic barriers to air moving through the region. What makes these mountains unique when compared to other mountains around the planet (for completeness we should also include Mt. Logan and Mt. St. Elias), is their position at high latitudes (63^{o} N). This translates into cooler temperatures at a given elevation compared to the same elevation on a mid-latitude mountain. However, their true uniqueness is due to the proximity in which the summits of these mountains lie in relation to the polar and arctic jet streams. Since the polar jet stream is constantly migrating between the Pacific Ocean and Alaska, there are times when it is positioned directly over Denali National Park (Figure 7.3). When this occurs, the summits of McKinley and Foraker are quite close to the level of maximum jet stream winds.

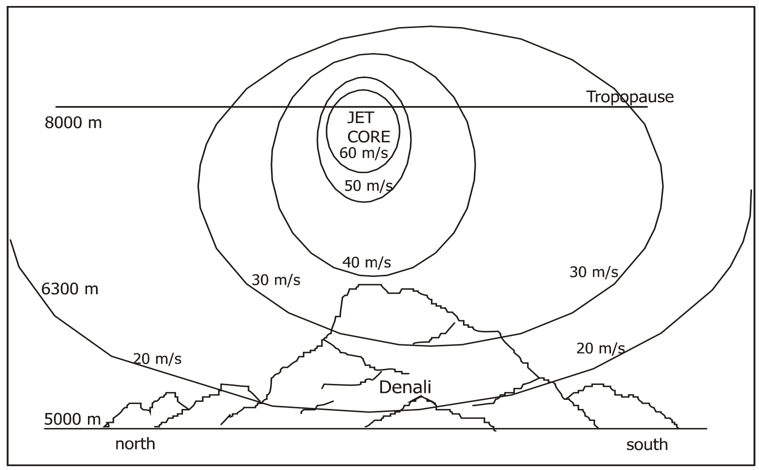


Figure 7.3- Idealization of jet stream winds over the Alaska Range. Due to a lower tropopause at high latitudes, strong winds extend down toward the summit in this example.

From a statistical perspective the Aleutian low is much stronger and persistent in the winter than in the summer. However, occasionally during the April-June period, powerful lows do develop in the central and southern Bering Sea (Figure 7.4a). These storms have a tendency to transport large amounts of moisture into the Alaska Range from the southwest. Most of these storms produce high winds and fresh snow for a period of 36-48 hours, however on occasion they can linger for much longer. With a low positioned in the eastern Gulf of Alaska (Figure 7.4b) generally two scenarios are plausible: if the storm is fairly weak, climbing weather on McKinley will remain good. If the storm is powerful then expect moderate to strong easterly winds with considerable amounts of new snow.
This particular pattern tends to generate widespread convection, hence snow is intermittent (showers) rather than continuous.

Westerly flow often brings moderate moisture to the slopes of McKinley from the Bering Sea (Figure 7.4c,d). When the state of Alaska is under the influence of a ridge of high pressure (Figure 7.4e), a broad range of weather can result: if the pressure gradient (or height gradient) is weak and the jet stream winds are weak or well to the south of McKinley, the climbing weather should be good. If the jet stream is over the mountain and moderate to strong and the flow is from the northwest, expect substantial wind speeds high on the mountain, although there is a good chance that precipitation will be light or not a factor at all. If however the ridge moves east, and the winds become southwesterly, expect poor climbing weather. In fact many of the most significant storms that occur

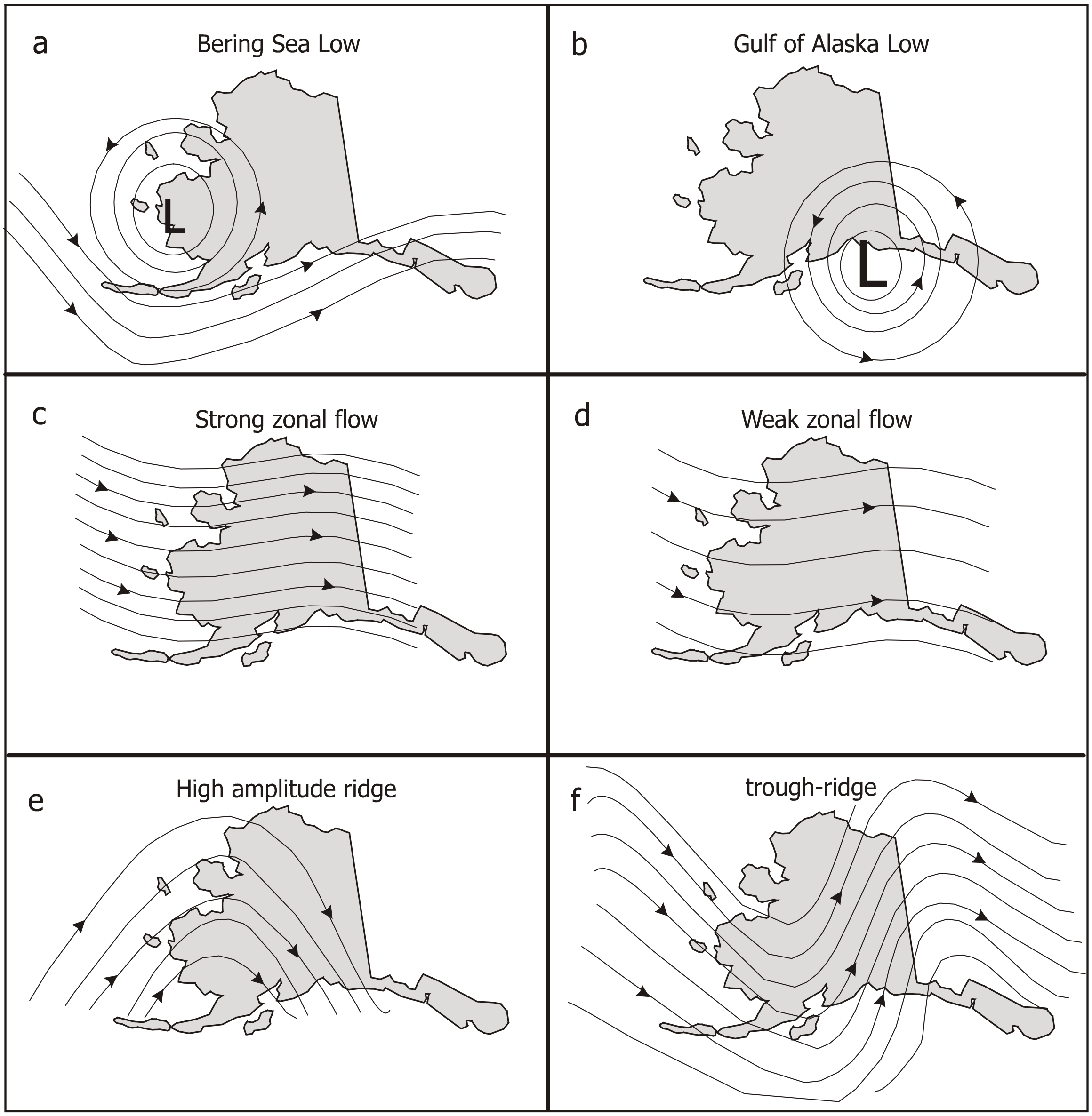


Figure 7.4- Synoptic-scale weather patterns over Alaska as depicted by 500 mb heights.

during April-June are associated with a high amplitude ridge positioned over the eastern Gulf of Alaska, with a trough to the west (Figure 7.4f). These storms are the most dangerous for climbers who are on Mt. McKinley and Mt. Foraker because the ridge is semi-stationary, often producing high inds and fresh snow for many consecutive days.

At lower elevations around Denali (and the Alaska range in general), the annual temperature range is extremely large. During the winter, when Alaska is under the influence of a ridge of high pressure, the coldest temperatures occur at lower elevations where very strong inversions develop due to intense radiational cooling. Temperatures in the -35^o C to -50^o C (-31^o to -58^o F) can be expected during these periods, which often last for one to two weeks. ‘Normal’ winter temperatures at low elevations (below 2000 m) range from -15^o C to -25^o C (5^o to -13^o F) . At higher elevations winter temperatures (based on 500 mb free atmosphere values) are typically around -30^o to -40^o C (-22^o to -40^o F). In the winter when synoptic storms move over the region, temperatures warm considerably at all elevations. Some of the largest increases in temperature however, occur at lower elevations. As a warmer air mass moves into the Alaska Range, cold surface air which was trapped by an inversion, is replaced by the warm air. When this occurs surface temperatures can increase by 20º C (36º F) in a 24 hour period.

During the summer at elevations below 1500 m (4,900 ft), temperatures range from lows around +5º C to highs near +25º C (40º to 77º F). At the Kahiltna Base Camp (2135 m) mid-April through June temperatures typically range from -10º C to +5º C (14º to 40º F) , with a gradual warming from late April to June. At the rangers camp (4270 m) during the summer temperatures range from to -10º C to -25º C (+14º to -13º F) , while near the summit -25º C to -35ºC (-13º to -31º F) . These numbers represent 'typical' values, temperatures can of course deviate substantially from these. The seasonal temperature trend at 700 mb and 500 mb in the free atmosphere is illustrated in Figure 7.5. The data used to construct this plot represents a 30 year average of each day from April 15 to June 30. The steady increase in temperatures is evident at both levels. On a daily basis the coldest summer temperatures occur when a low or trough moves into the region (opposite of winter case). Also note that even though it is summer time and the sun is above the horizon for 17-21 hours, there is a pronounced diurnal temperature cycle in the Alaska Range in general.

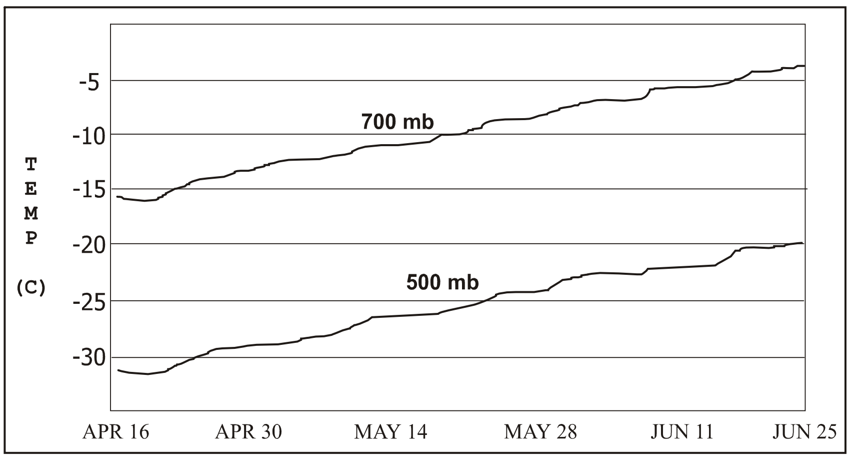


figure 7.5- Thirty-year average of free atmosphere air temperatures over Mt. McKinley. 700 mb corresponds to an elevation of about 3300 m (10,800 ft), and 500 mb is roughly 5400 m (17,700 ft).

The most important weather element that climbers on Mt. McKinley should be prepared for are high winds. This is based on the following considerations: 1) periods of high winds on the upper half of the mountain are frequent during the climbing season, 2) high winds often have a rapid onset, and; 3) A typical wind storm on Mt. McKinley lasts for 2 to 3 days, on occasion however they have been known to last for a week. Wind storms occur in association with many different weather patterns, low pressure systems (and troughs) may or may not produce high winds (figure 7.4a). In fact, many of the classic wind events that have taken place on the mountain, occurred when a ridge of high pressure was positioned over the eastern Gulf of Alaska (Figure 7.4f). Closer analysis of these wind events shows that they all occur when the upper level polar jet stream was positioned over or in close proximity to Denali National Park. The strongest winds occur between 300 mb and
200 mb (8.5-11.5 km), in response to large temperature changes in the lower stratosphere and upper troposphere. The second half of May 1992 for example, was a period when the upper half of McKinley and Foraker experienced very strong winds. During this interval however, there were periods of 12-24 hours, when wind speeds decreased and climbers were able to move around without getting blown off the mountain. Wind storms can occur in association with snowstorms or under cloud free conditions. Most multi-day wind storms tend to produce considerable amounts of fresh snow as well. In such cases as you might imagine, white-out conditions are the rule.

In order to determine the frequency and duration of wind storms on Mt. McKinley, 40 years of mid-tropospheric wind data was analyzed, restricting the time frame from mid-April to the end of June. The results are displayed in Tables 7.1 and 7.2. The data in Table 7.1 shows the frequency of free atmospheric winds above a given threshold. For example, 500 mb wind speeds are on average greater-than or equal-to 15 m/s (33 mph) about 19% of the time. What it does not indicate is how many days in a row the wind was above this threshold. The frequencies in Table 7.1 should be considered as minimum values since they were constructed using an average which tends to smooth the data. More importantly these values do not include local wind acceleration. Many climbers would like to know what is a typical critical wind speed above which movement on the upper slopes of the mountain is restricted. There is no deterministic value that works for everyone. There are a number factors that must be considered: the ability of the climbers and their fitness, the route, visibility restrictions, too name a few. As a rough estimate, I would speculate that 20 m/s (44 mph) is a threshold above which most climbers would be advised not attempt to move higher up the mountain.

TABLE 7.1- Wind frequency during climbing season

700 mb winds (approx. 3100 m)				
	≥ 10 m/s	≥ 15 m/s	≥ 20 m/s	≥ 25 m/s
Mean	21%	4.5%	1%	----
500 mb winds (approx. 5300 m)				
Mean	47%	19%	8%	2%

In Table 7.2 we have categorized wind storms based on both speed and duration. There will be times when a rapidly moving weather system such as a shortwave trough or low, brings high winds and snow to the mountain, but the storm is short lived (less than 48 hours). These are an inconvenience but are not that disruptive to climbers. From time-to-time however, a weather pattern is created which produces very high winds over the mountain for a sustained period of time. These

TABLE 7.2- Number of wind events per category between 1958-1997 at 500 mb

	≥ 20 m/s 24-48 hours	≥ 20 m/s 48-72 hours	≥ 20 m/s > 72 hours
April*	10	5	9
May	13	4	4
June	5	2	2
* from April 15-30			

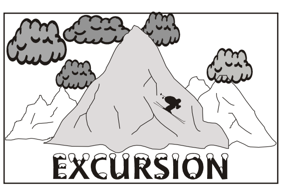
are the storms that can create havoc and loss of life. Fortunately, these types of storms can be forecasted, we strongly suggest that if you have access to the Denali weather forecast, to make use of it (see paragraph below).

Several conclusions can be reached from the data displayed in Table 7.2. From April to June there is a definite downward trend in the frequency of wind storms, keep in mind that the data for April only includes the second-half of the month. The frequency of the really large wind storms that pin climbers down for days on end are: once per every 3 seasons for April, once every 5 seasons for May, and once every 10 seasons for June. The sharp decrease in the number of high wind events between May and June is due to the weakening of the polar jet stream as the temperature gradient across East Asia and the North Pacific weakens in early summer.

There are several additional factors regarding winds on Mt. McKinley that should be taken into consideration: 1) Winds at the higher thresholds are usually very gusty. Make sure you factor this into your decision to climb on exposed routes during periods of high winds. 2) There are locations on the mountain where funneling of the wind through gaps and passes is common. Two locations on the West Buttress route where this commonly occurs are Windy Corner and Denali Pass. The amount of amplification through these gaps is unknown, but based on written accounts it must be substantial. In addtion, I highly recommend that you spend the time reading some of the accounts of major storms on Denali, this would include: *Minus 148* (Davidson), *Kingdom of the Mountain Gods* (Synder 1970), *White Winds* (Wilcox 1973), and *Facing the Extreme* (Koncor 1998). If nothing else these accounts will give you some appreciation of what extreme conditions on the mountain are like.

The Fairbanks office of the National Weather Service issues a daily weather forecast for Denali National Park (http://pafg.arh.noaa.gov). This forecast includes estimates of wind speeds, wind direction, cloud coverage and possible precipitation; for base, middle and summit elevations. These forecasts are available at both base and ranger camps on the West Buttress route. The detailed portion of the forecast is valid for 48 hours, while a general weather description extends the forecast for an additional three days. Keep in mind that the forecasted wind values are for the free atmosphere. It is very likely, due to mountain amplification and funneling, that actual wind speeds will be much higher at times.

Many readers are probably wondering if there are any periods during the climbing season when the weather is better for a summit attempt than at other times? In our opinion there is no magical period between mid-April and the end of June, when the weather is statistically better than at other times. In light of this however, there is certainly a decrease in the frequency of strong winds and a gradual warming as the season progresses. But this comes at a cost: in June there is an increase in cloud cover (mainly convective), softer snow, greater chances of avalanching and weaker snow bridges on the lower half of the mountain. The bottom line for weather considerations on Mt McKinley is: be prepared for extreme conditions. Along the same line of thought, while your on the upper half of the mountain keep a close watch on significant changes in wind speed and direction, cloud cover and visibility.



6000 meters- Alaska Range versus Himalaya

On a number of occasions we have heard the following statement "climbing at 6000 meters on Mt. McKinley is equivalent to climbing at 7000 meters in the Himalaya." True or false? Let's examine this statement in more detail. What are the *main differences in weather and climate between the two ranges? The first consideration are temperatures. It will certainly be colder in the Alaska Range; how much colder depends on the current weather affecting each region at any given time. Fortunately we do know the <u>average free atmosphere</u> temperatures at 30° N and 60° N (close enough to each region for this argument), at various times of the year. This data set was collected over many years by the U.S. Air Force and is called the Standard Atmosphere. On July 1 the middle troposphere (between 3-8 km or 9,800-26,200 ft) is on average 12° C (22° F) colder over the Alaska Range. On January 1 for comparison, the temperature difference increases to about 21° C (38° F).*

The second factor we must consider are the winds. This really depends on the position of the arctic and polar jet stream over the Alaska Range and the subtropical jet stream over the Himalaya. In general, during the summer monsoon in the Himalaya (June-September), the sub-tropical jet disappears, however in the pre-monsoon and post-monsoon seasons, upper-level winds are generally quite strong. In addition, winds on a given mountain in the Himalaya are highly dependent on the type of terrain in the surrounding area. If you are at 6000 m in an area with many peaks above 7000 m, than it is possible that the higher peaks act as a barrier to the winds to some degree. In all fairness, this type of wind comparison is highly questionable since actual wind speeds vary on a daily basis. So we will assume that no meaningful wind comparison can be made between the two ranges.

Thirdly what about the amount of available oxygen at each site? For reasons discussed in Chapter 2, oxygen decreases vertically due to a decrease in air density. The amount of oxygen available at 500 mb remains constant. However the height of the 500 mb level (we could use any level) above the ground changes from day-to-day. This means that for a climber who is at a fixed elevation, the amount of available oxygen depends on the pressure at that elevation. In July for example, the difference in the geopotential height field at 6000 m (19,700 ft) over the Alaska Range and the Himalaya is about 270 m (890 ft). This translates into a pressure difference of about 18 mb, or a difference in the oxygen content of about 4%. Therefore during the summer, a climber at 6000 m on Mt. McKinley for example, has about 4% less oxygen to work with than a climber at the same altitude on any peak in the Himalaya. In January, a climber would have about 7% less oxygen on Mt. McKinley due to considerably colder conditions at higher latitudes.

In summary, <u>on average</u>, climbing at 6000 m in the Alaska or St Elias Ranges is roughly equivalent to climbing at an elevation of about 6500-7000 m in the Himalaya as per temperatures and available oxygen.

<u>Wrangell Mountains and St. Elias Range</u>

The Wrangell Mountains located in the Copper River Basin, form a link between the Alaska Range to the northwest and the St. Elias Range to the southeast. Due to their closer proximity to the Gulf of Alaska, the Wrangell Mountains tend to receive a higher amount of annual precipitation when compared to the eastern half of the Alaska Range, but considerably less than the St. Elias Range. In other respects this range has a similar climate regime as the Alaska Range. Most synoptic storms that move into the region do so from the southwest or south. In addition, from May-August the whole area experiences vigorous convection.

Anyone who has ever flown over or even just looked at the St. Elias Range on a map, quickly notices the sheer volume of glacial ice that can be found in this region, a testament to the large amounts of precipitation this range receives each year. A major factor is that the southern half of the St. Elias Range is considered the highest coastal range on earth. The St. Elias Range has a stepped structure that gradually rises to the highest peaks which are located some 40-100 km (25-61 mi) from the coastline. This means that large amounts of moisture are transported into the higher mountains. Most of the synoptic-scale storms which produce so much precipitation, move in from the southwest to south. With a large amount of terrain above 3000 m (9,800 ft), the St. Elias Range produces considerable upstream blocking and therefore enhanced precipitation due to very moist Pacific air being forced up the windward side of the barrier.

The St. Elias Range is virtually uninhabited, and as a consequence there is very little

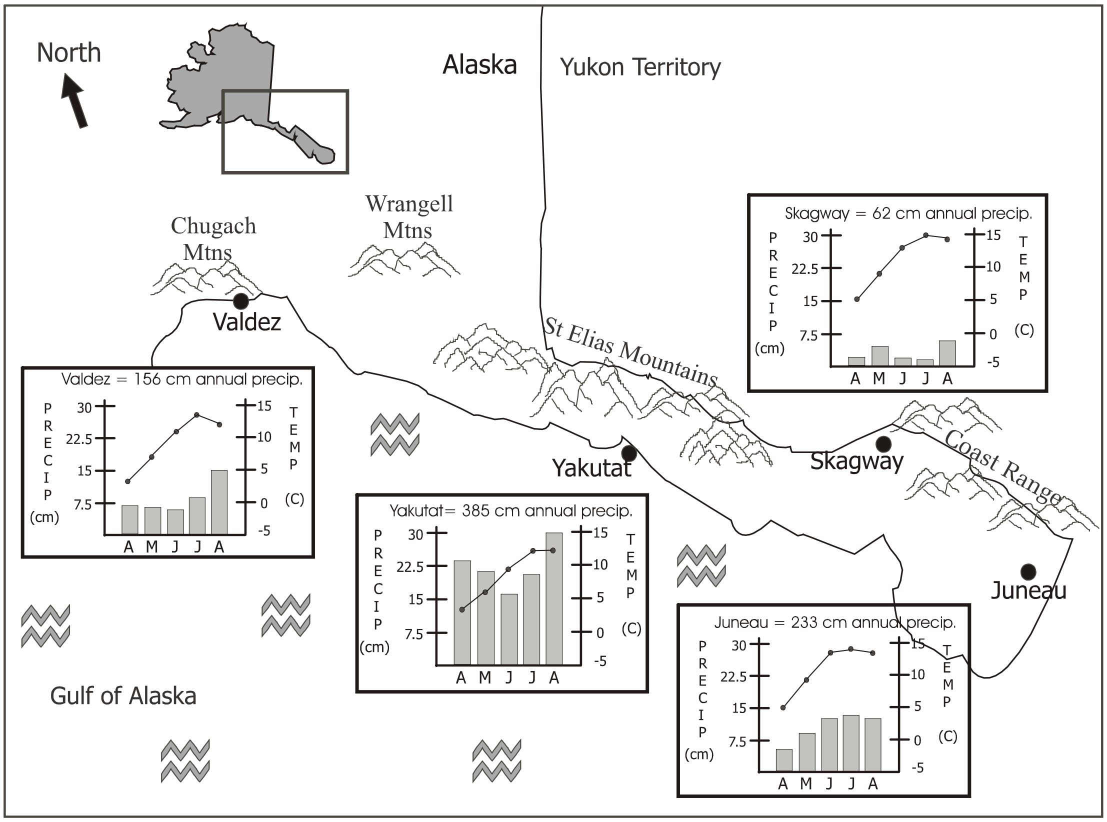


Figure 7.6- Summer precipitation and temperature for selected sea-level stations along the Gulf of Alaska. Notice how these stations have similar temperature trends, but differ widely in precipitation.

meteorological data. The nearest long-term weather station is at Yakutat, although situated on the coast, does provide some estimate of the regional weather (Figure 7.6). Yakutat's annual precipitation is about 3.85 m (151 in). The driest period is June and July, when only about 10% of the annual precipitation occurs. The wettest two month period is September and October when some 28% of the annual precipitation can be expected. Since Yakutat lies some 60 km (37 mi) upstream of the higher mountains, its precipitation is considerably less than what occurs at higher elevations. It has already been noted that the mountains around Mt. Fairweather probably receive the largest amounts of precipitation of any location in Alaska. It is not hard to figure out why; the Fairweather Range rises from sea-level to over 4000 m (13,100 ft) in about 40 km (25 mi).

In order to provide you with at least rough estimates of freezing levels in the St. Elias range during the climbing season, 13 years worth (1985-1997) of sounding data from Yakutat was analyzed. A typical freezing level above Yakutat in May is about 1200 m (3,900 ft), however on any given day it ranges from 500 to over 3000 m (1,600-9,800 ft), depending on the weather pattern. In June the mean rises to about 1900 m (6,200 ft) with a range of 1000 to 3500 m (3,200-11,400 ft). The highest freezing levels occur when a ridge of high pressure is located over Southeast Alaska (i.e.-the panhandle) or western Canada.

Alaska Coast Range

The coast range starts in the vicinity of Skagway where the St. Elias Range terminates, and extends down the eastern third of Southeast Alaska. The highest mountains are around 2150 m (7,050 ft), but the area is really known for its large icefields, especially the Juneau Icefield. Like the rest of the mountains lying adjacent to the Gulf of Alaska, the Coast Range has some large precipitation gradients in which the role of steep terrain cannot be over emphasized. The predominate moisture producing storm track for this region is from the south and southwest. During many of these storms, the low-level winds are often from the southeast as air moves from British Columbia into a low positioned in the eastern Gulf of Alaska. At the same time there is considerable funneling of these low-level winds in the fjord topography (also known as channels, passages, canals,) of Southeast Alaska.

Juneau's annual precipitation is on the order of 1.4 m (56 in) at the airport, to about 2.3 m (90 in) in town. Precipitation between May and July amounts to about 17% of the annual precipitation, while the August-October period receives about 37% of the annual precipitation. In both the St. Elias and Coast Ranges, due to large amounts of snowfall throughout the September through May period, avalanches are a real concern during the warmer months of the year. Freezing levels above Juneau in May are on average about 1500 m (4,900 ft) and in June 2000 m (6,500 ft), but vary considerably from week-to-week.

Alaska Weather Summary

* Driest period- April through June
* Wettest period- August through October
* Primary storm track direction- Southwest or south.
* Upper level winds- Strongest from the west-through-south, can occur anytime of the year. Beware of periods of very strong winds when jet stream(s) is over Alaska.
* Convective activity occurs in the Alaska Range, Wrangell Mountains, north side (interior) of St. Elias and Coast Ranges. Cloud-to-ground lightning possible from late May through August, but generally not a major concern.
* Gorge winds: Frequent in all mountain ranges in Alaska in the winter, including the islands of southeast Alaska.
* Barrier jets: north side of Brooks Range, all mountains bordering Gulf of Alaska.
* Downslope windstorms: all mountain ranges primarily during the cooler months of the year. Main areas affected: Western side of Chugach Mountains, north side of Alaska Range, and west side of Coast Range.

WEB: National Weather Service

Fairbanks **http://pafg.arh.noaa.gov**

Anchorage **http://pafc.arh.noaa.gov**

Juneau **http://pajk.arh.noaa.gov**

Mt. McKinley weather observations (~19,200 ft) **www.denali.gi.alaska.edu**

Cascades

The Cascade Range extends some 1000 km (600 mi) from Mt. Girabaldi in southern British Columbia to Mt. Lassen in northern California. The average height of the Cascades is about 1700 m (5,500 ft), with the occasional stratovolcanoe rising above 3000 m (9,800 ft). You cannot begin to understand the weather and climate of the Cascades without consideration of the influence of the Coast Ranges as well. With an average height of about 1000 m (3,280 ft), and lying on average about 100 km (61 mi) upstream of the Cascades (with respect to the primary wind direction), the Coast and Olympic Ranges heavily influence the distribution of precipitation in northern California , western Oregon and western Washington. In fact there are sections of the Coast Ranges that receive as much annual precipitation as any location in the Cascades. In Washington for example, the relatively compact Olympic Range, is one of the wettest regions in the continental USA. The Coast Ranges of Oregon and northern California, although not as wet as the Olympics, do substantially block the inflow of low-level moisture as well.

There is a marked increase in annual precipitation between the southern and northern halves of the Cascades. Schermerhorn (1967) estimated the latitude dependency of annual precipitation at 8% per degree of latitude. Therefore, the mountains in the vicinity of the Oregon-California border receive about half the annual precipitation as the mountains near the Washington-British Columbia border. This particular distribution of precipitation is a result of the seasonal position of the storm track (Figure 7.7). During the winter in the Pacific Northwest, the jet stream and associated storm track, are frequently found between 45^o N and 55^o N, resulting in higher storm frequency in the northern half of the Cascades. From July through September the polar jet stream weakens and migrates northward. This shift typically produces extended periods of warm, dry weather in both the northern and southern parts of the range.

Since the Cascade Range is oriented north-to-south and due to the fact that the winds predominately blow from west-to-east, there is considerably more variation in temperature and precipitation across the width of the range, than there is over a comparable north-to-south distance. Due to the relatively low height of the Cascades, considerably more moisture is transported to the eastern slopes, when compared to the Sierras for example. In order to understand the west-to-east distribution of snow and rain, two precipitation transects, one from central Washington and other from northern California are displayed in Figure 7.8.

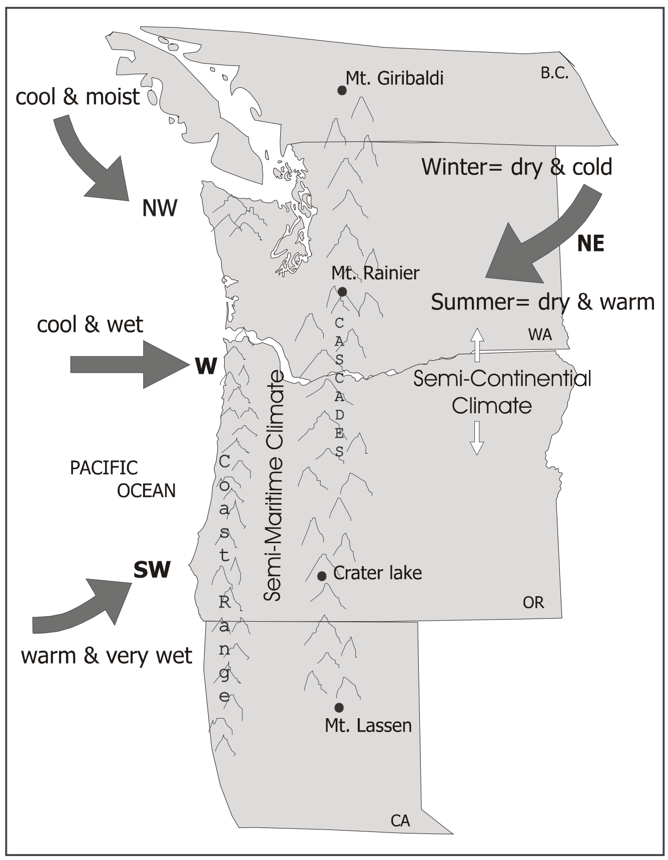


Figure 7.7- Storm tracks of the Pacific Northwest.

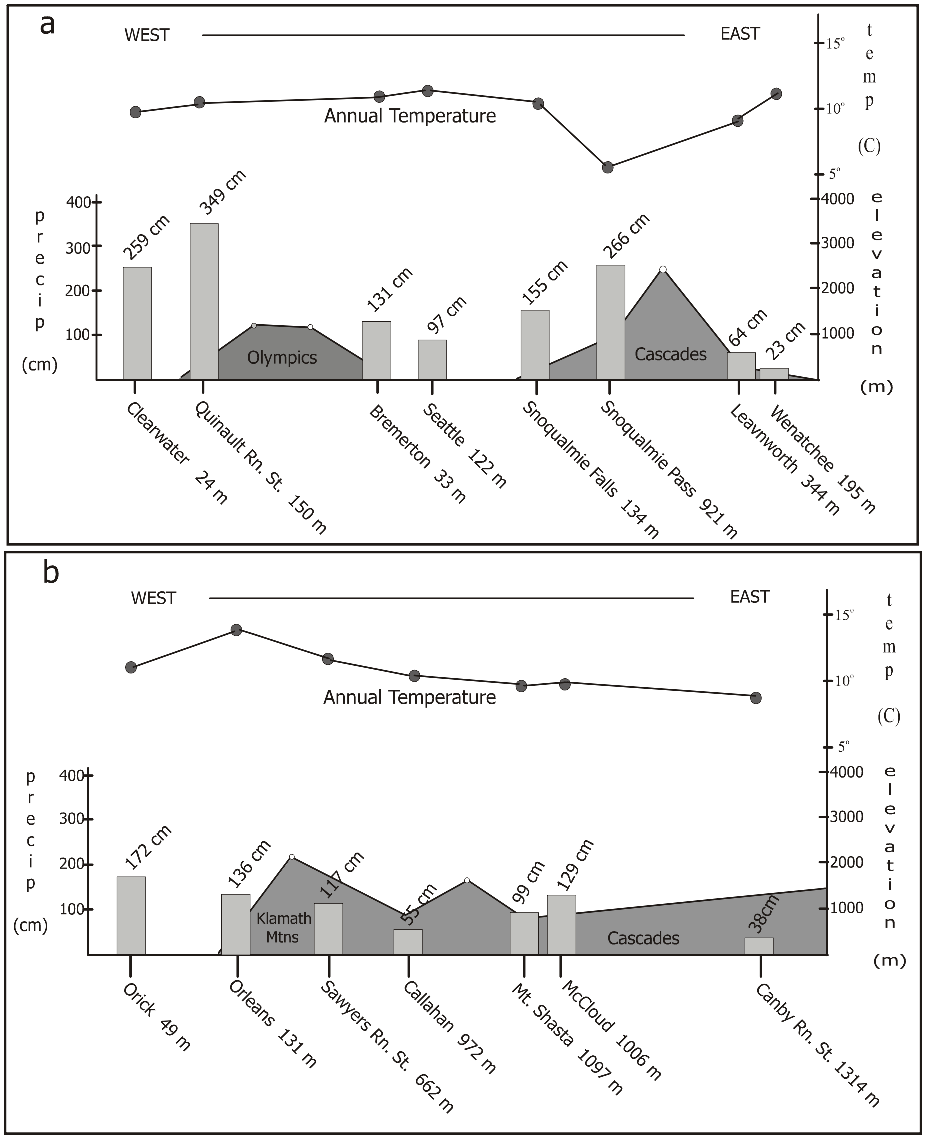


Figure 7.8- Two west-to-east transects across the Coast and Cascade Ranges. Vertical bars show annual precipitation in centimeters. Shading is a rough representation of the topography (m). (a) is through central Washington, and (b) is through northern California.

In the upper panel notice the pronounced drying to the lee of the Olympics and then a sharp increase in precipitation over the windward slopes (western) of the Cascades. This is followed by a rain shadow on the eastern slopes of the range. In northern California, precipitation reaches a maximum along the coast and western slopes of the Klamath Mountains, with a secondary maximum in the vicinity of Mt. Shasta.

Figure 7.9 displays annual values of precipitation, temperature, and snowfall for selected stations in the Cascades, with some stations along the coast added for comparison. Large-scale precipitation, as was noted earlier, is controlled by latitude (position of storm track), elevation, and distance from coastline. On the local-scale however, the actual amount of precipitation that a station receives is a function of the topography surrounding the site. For example, a station located in a valley will usually receive considerably less precipitation than the higher ground surrounding the valley. It should be point out that most of the climate stations in the Cascades are at relatively low elevations. You should also note from Figure 7.9 that along the coast, there is a definite increase in the number of rainy days from central Oregon northward. The coastal stations of northwest California, for example, typically have only about half (55%) the number of rainy days as stations to the north. However, when it does rain along the coast of southern Oregon and northern California, it has a higher intensity (higher rainfall rate) when compared to coastal stations in Washington and northern Oregon

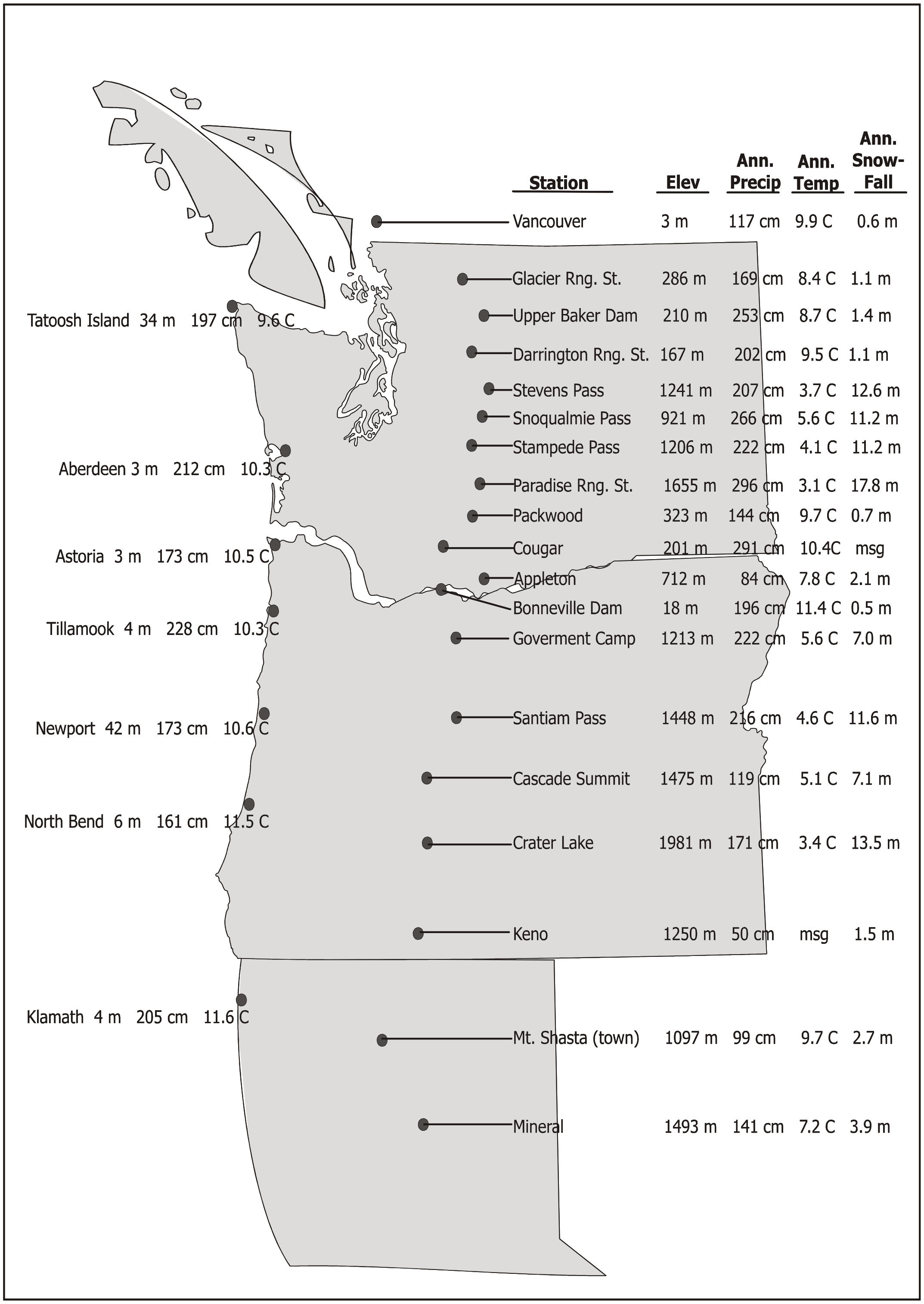


Figure 7.9- Climate data of the Pacific Northwest.

Another useful climate statistic is the number of days in which snow and rain occur over a given region. In Table 7.3 a precipitation frequency chart has been constructed that starts on the Washington coast and runs east through Mt. Rainier National Park (MRNP) to Yakima.

Table 7.3- Number of days with precipitation for west-to-east transect across central Washington.

STATION NAME	Number of days per year with > 0.2 mm (light)	Number of days per year with > 13 mm (moderate)	Number of days per year with > 25 mm (heavy)	Annual precipitation (cm)
Aberdeen	191	58	23	212
Olympia	164	33	9	129
La Grande	183	21	3	98
Longmire	184	59	22	213
Paradise	194	81	37	296
Ohanapecosh	169	51	21	196
Tieton Dam	113	15	5	68
Yakima	70	2	2	21

Notice the sharp decrease in the number of days with greater than 0.2 millimeters (.01 inch or what we call a “light” precipitation) to the east of the Cascades. What is of real interest is the frequency of days with moderate and heavy precipitation (greater than 25 millimeters). There is a pronounced decrease from the coast into the southern Puget Sound area, with a large increase for stations in the Cascades, followed by the dramatic decrease to the lee of the Cascades. The higher frequency of days with moderate and heavy precipitation in the Cascades can be attributed to two causes: the first is the stalling or blocking of fronts as they pass over the mountains. This produces precipitation events that have a longer duration. Secondly; the higher terrain of the Cascades creates very strong updrafts, which in turn produces areas of significant moisture convergence that subsequently produce heavy precipitation. What is also apparent from last column Table 7.3 is the increase in annual precipitation between southern Puget Sound (La Grande) and the Cascades, this clearly shows how elevated terrain enhances precipitation.

In general the Cascades of northern California receive about half as many rainy days as the Washington Cascades. However, the percentage of days with heavy precipitation is relatively unchanged from north-to-south. Unlike other mountainous regions of the western USA, convective activity in the Cascades is relatively mild. Summer hailstorms and thunderstorms are not much of a concern to backcountry travelers. The reason for the subdued nature of convection in the Cascades is primarily due to the presence of warm stable air in the middle-troposphere.

<u>Mt. Rainier National Park</u>

Since more than 10,000 people per year attempt to reach the summit of Mt. Rainier, not to mention the tens of thousands of hikers and skiers that congregate in the park at various times of year, a detailed look at the weather and climate of this area is presented. Figure 7.10 is a map of the park with distances between key geographic points given in kilometers. There are four locations within the park where year-round climate data is either presently collected or has been in the past

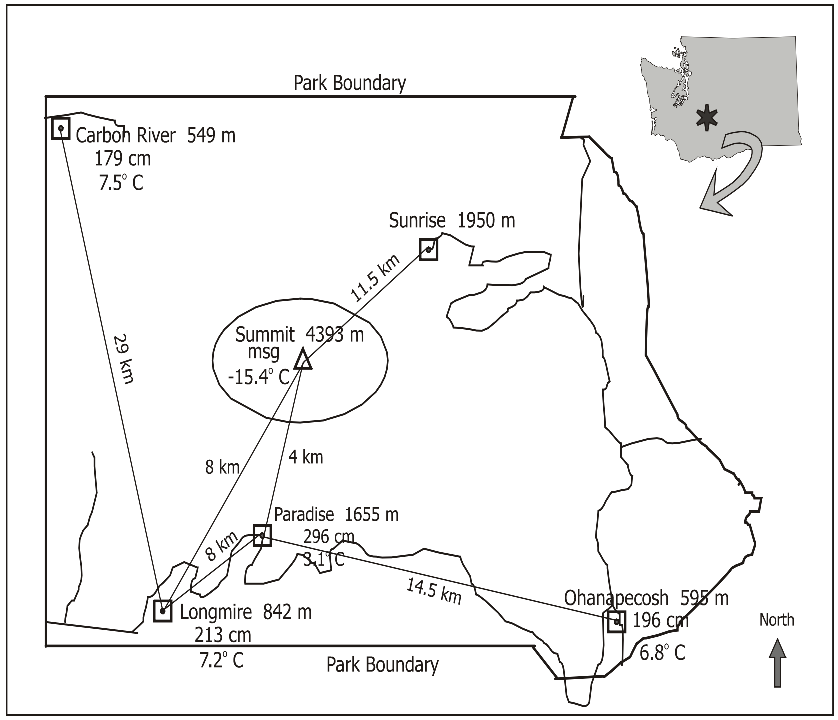


Figure 7.10- Mt. Rainier National Park with elevation (m), annual precipitation (cm) and annual temperature (c) displayed below station name. Dot-dashed lines represent distance (km) between key geographical locations. Dashed lines are roads.

These stations and their elevations are displayed on the first line of Figure 7.10, while annual precipitation (cm) and annual temperature (°C) are displayed on the lines below.

During the period of June 1989 through August 1990, the National Park Service installed a temperature sensor on a small tower which was located near the summit (4393 m or 14,410 ft) of Mt. Rainier. Monthly average temperatures for the 14 months of data that was collected is displayed in Figure 7.11 Keep in mind this is only a rough approximation of the true long-term average. Notice how July through September temperatures are fairly constant. This period is followed by rapid cooling, with the coldest temperatures occurring in the January and February period, followed by a large spring warming trend.

Examination of the entire temperature record shows that the lowest temperature measured was -37° C (-34° F) which occurred on several occasions in January and February. The highest temperature was +2° C (34° F) which occurred in July. Since temperatures were recorded every 3 hours, it is instructive to analyze the diurnal cycle. The typical diurnal temperature range during the summer is on the order of 5° C (9° F) . The largest change in temperature (±10° C or 18° F) occurs when there is a exchange of air masses over the Pacific Northwest. In order to provide the reader with some idea what temperatures they can expect at various elevations, Figure 7.12 was constructed. If you are planning a summit climb in mid-June for example, you can expect temperatures to range between

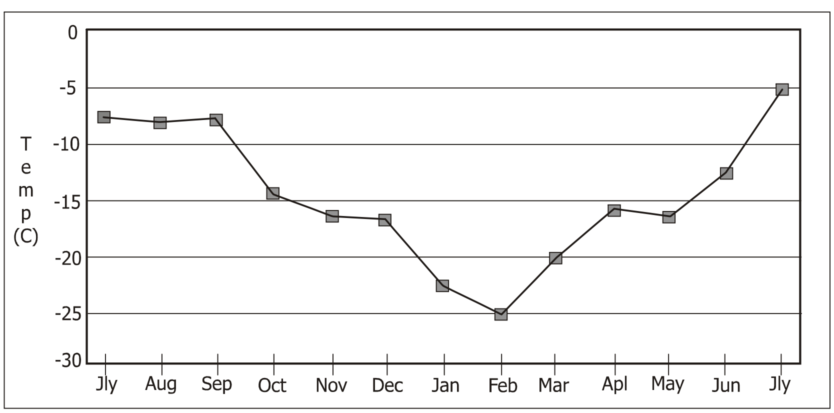


Figure 7.11- Mean monthly summit temperature (C) measure between June 1989 and August 1990.

-1º C to -11º C (12º to 30º F) at the elevation of Camp Muir (3050 m or 10,000 ft), and between -7º C and -16º C (3º to 19º F) on the summit. Note that these ranges do not include extreme values.

In Table 7.4 monthly statistics for precipitation, temperature and snowfall for the four climate stations in MRNP are given. Annual values are given on the bottom row. The wettest months are December and

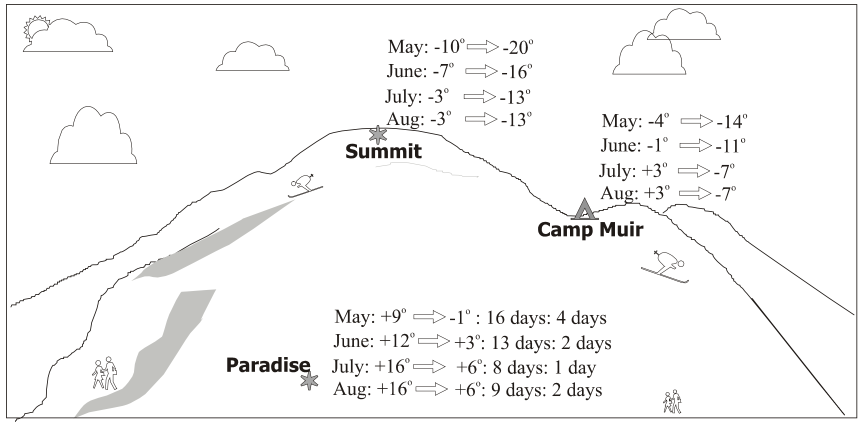


Figure 7.12- Summary of May-August temperature ranges. Paradise data also lists the average number of days per months with at least a trace of precipitation (third value), followed by the number of days per month with heavy precipitation (fourth value).

Table 7.4- Climate data for Mt. Rainier National Park

	Mean Monthly Precip. (cm)				Mean Monthly Temp. (C)				Mean Monthly Snowfall (m)			
	Carb	Long	Para	Ohn	Carb	Long	Para	Ohn	Carb	Long	Para	Ohn
Jan	21.4	31.9	45.1	33.5	0.0	-1.0	-3.3	-1.0	0.37	1.33	3.31	1.27
Feb	20.3	22.6	34.6	22.7	2.3	1.0	-2.3	1/0	0.26	0.94	2.56	0.73
Mar	18.4	22.5	31.0	17.9	2.4	2.3	-1.7	2.4	0.25	0.84	2.72	0.48
Apr	12.5	13.3	21.5	12.3	6.3	5.3	0.5	5.1	0.05	0.27	1.68	0.11
May	11.0	10.2	12.7	7.4	10.2	9.3	4.2	10.7	0.0	0.03	0.57	0.0
Jun	12.0	9.1	10.2	6.1	12.7	12.6	7.3	Msg	0.0	0.0	0.13	0.0
Jly	5.0	3.5	5.1	2.6	15.1	16.1	11.1	16.8	0.0	0.0	0.0	0.0
Aug	6.2	5.5	6.9	3.9	14.9	15.8	11.4	16.8	0.0	0.0	0.0	0.0
Sep	9.8	10.4	13.1	7.7	12.3	13.2	8.9	11.8	0.0	0.0	0.07	0.0
Oct	17.2	20.0	24.6	17.1	7.7	8.5	4.4	9.3	0.0	0.04	0.73	0.0
Nov	21.4	28.9	44.0	30.5	4.2	3.1	-1.0	2.4	0.03	0.35	2.33	0.26
Dec	24.2	34.8	47.2	34.0	1.9	0.5	-2.8	0.0	0.30	0.98	3.07	1.00
ANN	**179.4**	**212.7**	**296.0**	**195.7**	**7.5**	**7.2**	**3.1**	**6.8**	**1.3**	**4.8**	**17.2**	**3.9**

(Carbon River Ranger Station= Carb; Longmire Ranger Station= Long; Paradise Ranger Station=

Para;   Ohanapecosh Ranger Station= Ohn, Msg=missing data)

January , followed by February and November. There is an approximate 28% increase in annual precipitation at Paradise compared to Longmire. This increase in precipitation occurs over a distance of 8 km (5 mi) and a 800 m (2,600 ft) elevation difference

The annual precipitation measured at Paradise varies considerably from year-to-year. The long term average is 296 cm (116 in), the wettest year on record was 1975 with 386 cm (152 in) of precipitation, and the driest year was 1969 with 179 cm (70 in). In terms of snowfall, the seasonal mean for September through August, at Paradise is 17.8 m (58 ft) and 4.9 m (16 ft) at Longmire. The biggest snowfall season at Paradise was the winter of 1971-72 when 28.6 m (93 ft) of the white stuff fell. The lowest seasonal total at Paradise was 1976-77 with 10.5 m (34 ft). The single largest observed monthly snowfall was January 1972 when they received 7.1 m (23 ft).

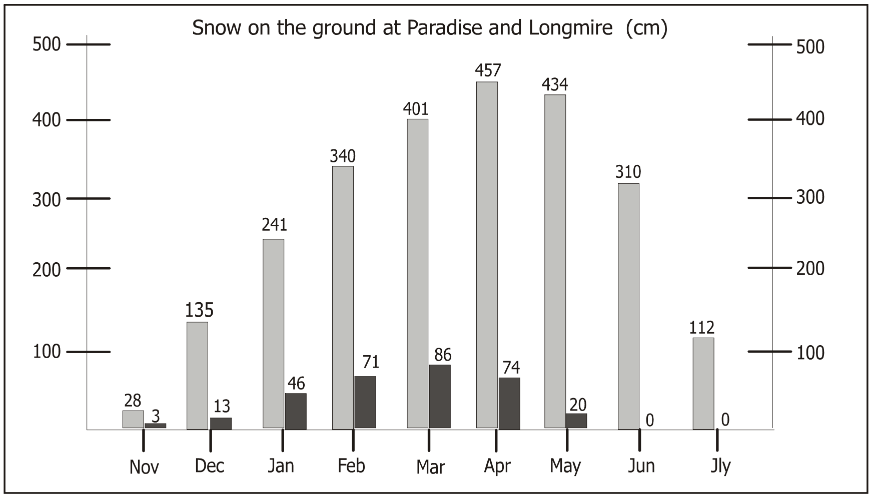


Figure 7.13- Comparison of snow on the ground (cm) at the beginning of selected months for Paradise (light shading) and Longmire (dark shading). Numbers on top of bars are the actual values. This graph is based on 30+ years of data.

Seasonal snow-on-the-ground at both Paradise and Longmire is shown in Figure 7.13. On average the deepest snowpack at Paradise (4.57 m or 15 ft) occurs during the first two weeks of April. At Longmire the peak is during the first two weeks of March with a typical depth on the order of 0.9 m (35 in). The long-term snow-on-the-ground records indicate that on average, snow disappears from Paradise during mid-August and from Longmire during the second week in May. There can of course be considerable variation from year-to-year.

<u>Olympic National Park</u>

The west side of the Olympic Peninsula in general and Mt. Olympus specifically, is the wettest region in the 48 contiguous states. Unfortunately little meteorological data has been collected at higher elevations within the Park. What climate data that does exists comes from a handful of low elevation sites on the edge of the Park, which are listed in Table 7.5. The wettest area is the southwest and western slopes. Not all of the stations listed in Table 7.5 are still recording data, nevertheless their period of record was long enough to establish climate statistics. The barrier aspect refers to what side of the Olympic Range the station is located on with respect to the storm track.

Notice the dramatic decrease in the number of days with heavy precipitation on the east and north side of the range. Table 7.5 illustrates two apparent precipitation anomalies: the relatively dry area between Port Angeles and Port Townsend, and the relatively wet zone on the east side of the Olympic Mountains centered around Quilcene. The dry area on the northeast corner of the Olympic Peninsula lies to the lee of the mountains for moist flow from the west and southwest. Northwest flow does bring rain to the area, but flow from the northwest is considerably less frequent and drier then flow from the west and southwest. The relatively wet region on the east side is probably due to the fact that the Olympics lie on the western edge of the Puget Sound Convergence Zone. We will not describe this meteorological feature in any detail, just summarize by saying that it forms when winds flowing around the north and south side of the Olympics, converge over Puget Sound, forming an area of enhanced rising motion and precipitation.

Table 7.5- Climate data for Olympic Peninsula

Station	Elev (m)	Barrier Aspect	Annual Precip (cm)	Number of days with precipitation >0.2 mm	>25 mm	Annual Temp. (C)
Sappho	232	NW	242	146	29	9.4
Forks	107	W	301	212	38	9.8
Spruce	122	W	318	188	41	msg
Quinault Rn. St.	67	SW	349	163	46	10.4
Cushman Dam	232	SE	255	168	35	10.4
Quilcene	37	E	140	162	15	10.1
Port Angeles	30	NE	65	141	3	9.6
Elwha Rn. St.	92	N	142	148	15	9.2

On the west and southwest side of the Olympic Range from November through March, on average about 20-23 days out of the month will have measurable precipitation, with 4-8 of those days receiving in excess of 2.5 cm (1 in) of rain per day. From June through September, measurable rainfall occurs on 4-8 days out of the month, with 0-2 days receiving heavy precipitation.

During the August 1957-July 1958 period, a meteorological station was temporarily established at the base of Mt Olympus. During this period that station recorded 378 cm (149 in) of precipitation (snow and rain). During the same period the stations at Forks and Spruce, which had below normal precipitation, measured 240 cm (94 in) and 296 cm (117 in), respectfully. This would make the area around Mt Olympus the wettest in the region with an estimated annual precipitation on the order of 410-430 cm (160-170 in).

<u>Crater Lake National Park</u>

Crater Lake is unique because it represents one of the highest stations in the Cascades that has a long term record of observations (60+ years). The climate station is located at an elevation of 1981 m (6500 ft) and receives an annual average of 171 cm (67 in) of precipitation. This is considerably less precipitation than what is observed at Paradise on Mt. Rainier, but a larger percentage falls as snow rather than rain. As is true for most of the Cascades, the wettest months are December and January followed by November and February. However, the month that receives the most snow is January with 2.7 m (106 in) followed by December with 2.4 m (94 in). The average annual snowfall is 13.5 m (44 ft), the largest snowfall in a season was 1950-51 with 21.2 m (69 ft). The lowest was 1976-77 with 6.5 m (21 ft). The deepest snowpack occurs in late March and early April with an average depth of about 3 m (9.8 ft). Snow on the ground typically disappears somewhere around the first of July.

Mount's Shasta and Lassen

As noted previously in this chapter, the Cascades of northern California receive considerably less precipitation than the rest of the range. As seen in Figure 7.9, the town of Mt. Shasta located just off of Interstate 5, at the relative low elevation of 1097 m (3,600 ft) receives on the order of 99 cm (39 in) of annual precipitation. There are no long term climate observations higher on Mt. Shasta, however there is a snow course located at 2410 m (7,900 ft) on the west side of the mountain. Long term records from the snow course show that the average snow depth as of April 1 is about 3.23 m (128 in). The same data set also suggest that the annual precipitation at the snow course is on the order of 1.6 m (63 in), a 60% increase over what is measured in the town of Mt. Shasta.

Mt. Lassen which lies approximately 120 km (75 mi) to the southeast of Mt. Shasta is in an area that receives significantly more precipitation than Mt. Shasta This is due to the fact that when storms move into the region from the west to southwest, there is considerably less mountainous terrain upstream of Mt. Lassen. The Lower Lassen Peak snow course, located at 2515 m (8,200 ft) has a long term average April 1 snow depth of 4.44 m (175 in). The amount of water equivalent in the average April 1 snowpack suggests that this site receives about 77% more precipitation than is observed at the nearest climate station at Mineral, some 16 km (10 mi) to the southwest.

Winter snowstorms occur less frequently here than in any other part of the Cascades, nevertheless, when these storms do occur, snowfall can be on the order of a meter (3+ ft). The predominate snow producing pattern is when a 500 mb low or trough is positioned off the coast of Oregon. This produces strong and moist west-to-southwest flow into northern California. However, since the polar jet is frequently well to the north of this region, sunny winter days are common here. Convection during the summer is in turn more substantial here than in the rest of the Cascades due to the drier conditions. Therefore cloud-to-ground lightning is a bigger threat here than in the rest of the range.

Cascade Weather Regimes

As you have already figured out, the weather on any given day for a specific region of the Cascade's depends on many factors. The number one concern, as in any mountainous region, is knowing the location of the polar jet stream. Figure 7.14 shows six of the major weather patterns that control Cascade weather. For example, a cut-off low pressure system is depicted in 7.14a. This type of storm typically produces plenty of precipitation for periods ranging from 2 to 4 days. When the center of the storm is located near Vancouver Island or off the Washington coast, expect heaviest precipitation in the northern half of the range. The Coast Range will of course get hit hard as well with this type of pattern. When the storm center is near the mouth of the Colombia River or the central Oregon coast, the southern half of the range receives the heaviest precipitation.

These storms are far more intense and common from October through April, than at other times of the year. You can expect strong westerly winds at pass level and above. During the cooler months of the year, these storms can also produce fair amounts of rime ice at higher elevations in the

Cascades. The exact timing of precipitation depends on considerably on the passage of fronts. Note that with these types of events, precipitation at lower elevations is much less than what occurs at higher elevations. Therefore the amount of precipitation that may be currently falling in Seattle or Portland, is not much of an indicator what is occurring at 1,200 m (4,000 ft) for example. The good news is that these storms are pretty easy to forecast several days in advance. Their longevity

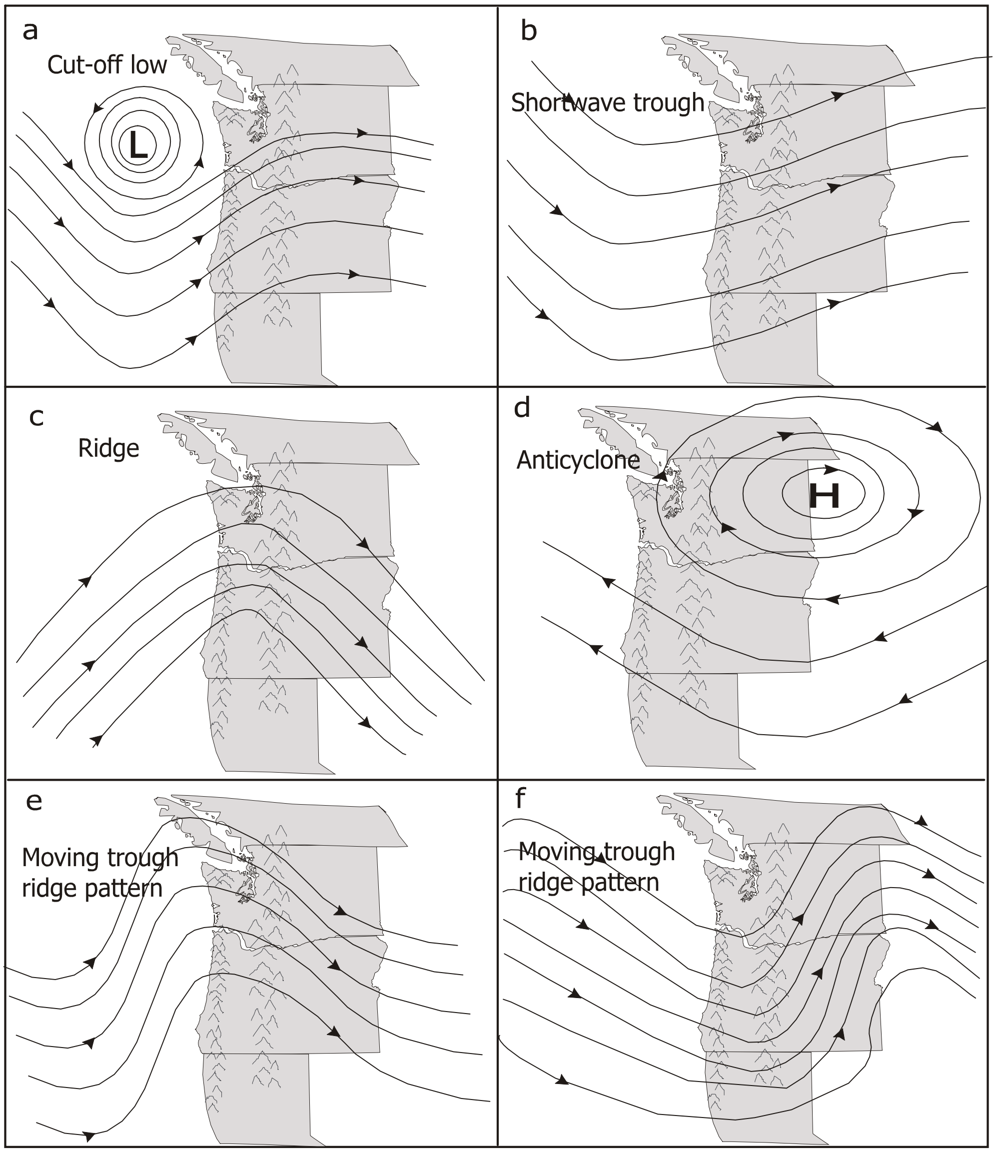


Figure 7.14- 500 mb flow patterns for the Pacific Northwest.

however is a bit harder to predict.

The pattern displayed as 7.14b is called a shortwave trough and can occur in winter as well in summer. It is a weaker version of the pattern in 7.14a. The winds are usually not as strong and precipitation is not as great either. When this pattern occurs in summer expect a day, possibly two days of clouds and light precipitation. In a case like this you also need to look at what is occurring at the surface, if there is a fairly well developed surface low off the coast, then you should expect a more powerful storm (more precipitation and wind). If 7.14b occurs with no surface support, then you should expect mid-level and high clouds, with precipitation in the mountains being absent or light.

Figure 7.14c shows a high amplitude ridge (high pressure) over the Cascades. This pattern produces dry relatively cloud free conditions as long as the ridge axis remains to the west of the range. This occurs because the air in this part of the ridge is descending from higher elevations within the troposphere, hence it warms and dries in the process. Once the ridge axis moves to the east however, the flow into the Cascades is from the southwest, which in turn may produce rising motion. When this regime occurs during the summer you should expect some very high freezing levels (3500-4500 m or 12,000-15,000 ft). Note however, when the freezing level is mentioned in a weather forecast or discussion, it is for the free atmosphere only. At night, due to the ground emitting large amounts of longwave radiation, the air near the ground at high elevations can actually reach the freezing point. For example, if you are camped at 2130 m (7,000 ft) on an exposed ridge in mid-August, and if the free atmosphere freezing level is 3660 m (12,000 ft), its quite possible that your location could experience freezing temperatures by the early morning hours, due to the aforementioned cooling.

The pattern shown as 7.14d is called an anticyclone. In the winter this regime will produce some of the coldest temperatures of the year in the central and western regions of Washington and Oregon, especially if the anticyclone center is located north of the Canadian border. This pattern will produce moderate to strong easterly winds along the eastern slopes of the Cascades. After several days a pool of cold air often starts to accumulate and deepen in central and eastern Washington and Oregon. When this occurs very strong gap and gorge winds will occur in the western Cascades. On occasions these winds are strong enough to produce considerable property damage. If this pattern or something similar should occur during the summer then you should expect hot and very dry conditions (high fire danger as well), as air flows into the range from the Great Basin.

The flow patterns depicted in 7.14e,f are related. If you happen to look at the forecast models and see a pattern like what is shown, here is what you can expect: 7.14e is of course very much like what we already discussed in 7.14c, with one major difference, notice the trough out to the west. Expect several days of very nice weather as the ridge approaches and is over the Cascades. Increasing clouds and cooler temperatures will occur as the trough approaches (7.14f), the extent of cloud cover and precipitation of course depends on the strength of the trough. This pattern represents are eastward moving traveling wave which are more pronounced in winter than in summer.

The vast majority of precipitation that falls in the Cascades and Coast Ranges occurs in the October through April period. Needless to say, winters in the region are typically cloudy and wet. From the central Oregon area northward, May and June can be considered a 'transitional' weather period. What we mean is that the weather during these two months typically alternates from being relatively cloud free and warm, to cool and wet. The cycle roughly last a week or so, and is linked to the north-south migration of the polar jet. The dry season typically starts in early July and extends into late September, and in some years into late October.

Note that during the summer months, weak west-to-southwest flow often leads to the intrusion of marine air (low stratus clouds) into western Washington and western Oregon. This marine layer is a 'fly in the ointment' to hikers and climbers. The eastern extent of these clouds can vary considerably. The good news is that the top of the cloud layer is frequently located between 1500 m and 2100 m (5,000-7,000 ft). One of the best marine layer forecasting rules of thumb is to look at the pressure difference between North Bend (Oregon) and Seattle. When the geostrophic flow is from the southwest and the pressure at North Bend is 4 mb or more than at Seattle, marine stratus clouds will usually form. Also note that you should use visible satellite imagery instead of infrared when you want to look for low stratus and fog. This is the case because infrared images tend to 'emphasize' higher clouds rather than low clouds.

Cascade Weather Summary

* Wet Season: October through April
  Transition season: May and June
  Dry season: July through September
* Large snowfall accumulation above 1500 m (5000 ft)
* A significant decrease in frequency of rainy days from north-to-south
* Decrease in annual precipitation from north-to-south (not necessarily true for coastal stations)
* Storms move into region from northwest-to-southwest
* Summertime convective activity is limited, but more frequent in the southern part of the range.
* Gorge winds: Columbia Gorge, Snoqualmie Pass, Fraser Gap, Howe Sound
* Barrier jets: Coast Range, western slopes of Olympic Mountains
* Glacier winds: very localized around glaciated mountains
* Downslope windstorms: more frequent on east side but occasionally occur on the west side during the cooler months of the year, when the geostrophic winds blow across the Cascades from east-to-west.

WEB- National Weather Service

Seattle **www.wrh.noaa.gov/seattle**
Portland **www.wrh.noaa.gov/portland**
Medford **www.wrh.noaa.gov/medford**
Sacramento **www.wrh.noaa.gov/sacramento**
Northwest Avalanche Center- **www.nwac.noaa.gov**

Sierra Nevada

The Sierras begin south of Mt. Lassen and stretch some 600 km (375 mi) further to the south, terminating northeast of Bakersfield. The Sierras differ from the Cascades in a number of ways: the Sierras are primarily fault block mountains formed by the differential lifting/sinking of the earth's crust. As a result the Sierras form a continuous elevated barrier, especially the region from Yosemite to Sequoia National Park. This section of the Sierra crest, often referred to as the High Sierra, has an average height of 3500-4000 m (11,400-13,100 ft).

The predominate storm track for heavy precipitation events in the Sierras is depicted in Figure 7.15. This type of flow pattern forms when either a trough or low forms off of the coast of Oregon or even Washington. This produces strong southwesterly winds at virtually all levels within the troposphere over the Sierras, as the axis of the polar jet stream becomes positioned directly over central California. Since most of the Coast Range along the central California coast is not very high (less than 1000 m or 3,280 ft), considerable amounts of Pacific moisture is transported the 250 km (155 mi) into the Sierras. Heggli and Rauber (1988) note that powerful storms as shown in Figure 7.15 are responsible for the majority of floods which occur in the Sacremento and San Joaquin valleys. It should be pointed out that precipitation in the Sierras tends to be much more episodic in nature than precipitation in the Cascades. What do we mean by episodic precipitation? Basically there are fewer days with precipitation in the Sierras compared to the Cascades, however when precipitation does occur, it is usually of high intensity. For example, the average number of days with a trace or more of

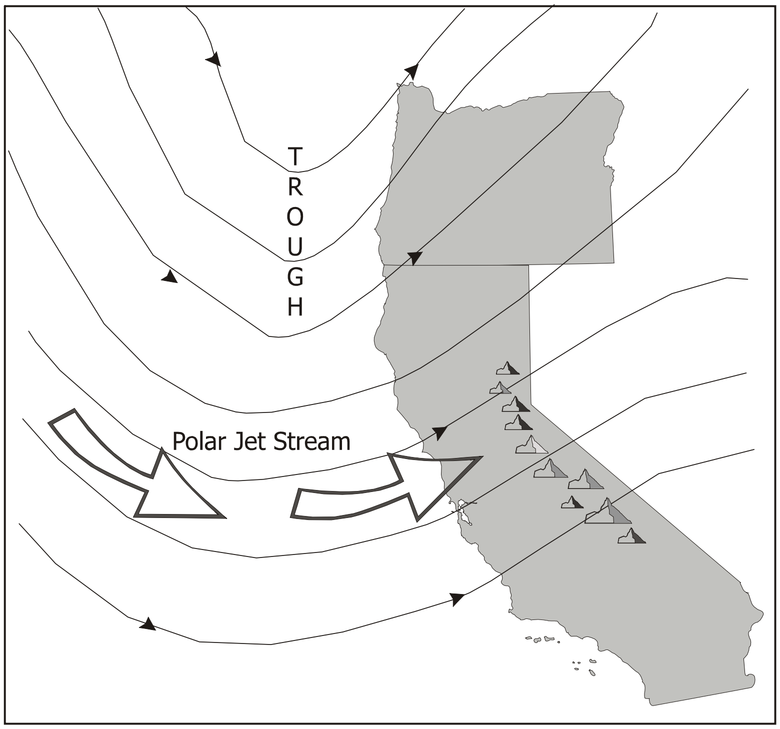


Figure 7.15- Typical 500 mb flow pattern for a major wintertime precipitation event in the Sierras.

precipitation in the western Sierras is on the order of 75 days, less than half of the number of days that occur in the northern Cascades. What this should mean to you, the backcountry traveler, is that when a synoptic-scale storm moves into the Sierras, rainfall or snowfall totals can be quite impressive.

Climate statistics for select stations in the Sierras are displayed in Figure 7.16. Keep in mind that these climate stations are located at mid-elevations (between 1000-2000 m or 3,300 and 6,600 ft), so they may not be representative of higher elevations. What is readily apparent from this figure is the increase in precipitation in the Sierras relative to stations in the San Joaquin and Sacremento valleys. The amount of annual precipitation that a specific station receives, is of course, a function of the height of the terrain that is upstream of the station, as well as the height of the climate station itself. Notice for example that Blue Canyon along Interstate 80 receives double the annual precipitation as any of the stations in the vicinity of Lake Tahoe. This holds true because by the time parcels of air reach Lake Tahoe, much of the moisture has already been depleted over the western slopes of the Sierras.

In Figure 7.16 also notice how dry it is directly in the lee of the Sierras, as represented by Lee Vining (38 cm or 15 in) and Bishop (14 cm or 5.5 in). By comparison, the climate station at 3800 m (12,400 ft) in the White Mountains, to the east of Bishop, has an annual precipitation of 48 cm (19 in), more than three times the Bishop average. This shows that more moisture is flowing over the lee of the Sierras at crest level than one would expect by simply using the precipitation data from low elevation sites at the base of the eastern Sierras.

The distribution of annual precipitation in the Sierras can be summarized as follows: there is a pronounced winter wet season in the Sierras, and in the state of California in general. On the order of 75-80% of the annual precipitation falls from the first of November through the end of March. January is the wettest month without question. What is evident from the climate record is that there is a noticeable increase in the percentage of precipitation that is observed in the second half of winter, between the northern and southern Sierras. In other words, in the southern Sierras (south of Yosemite), more snow occurs during the months of February and March, than in November and December. The opposite is true in the northern Sierras.

In Table 7.6 we have listed average snow depths for a number of snow courses scattered throughout the Sierras. These sites are generally at higher elevations than the climate stations, and a number of them are in remote locations.

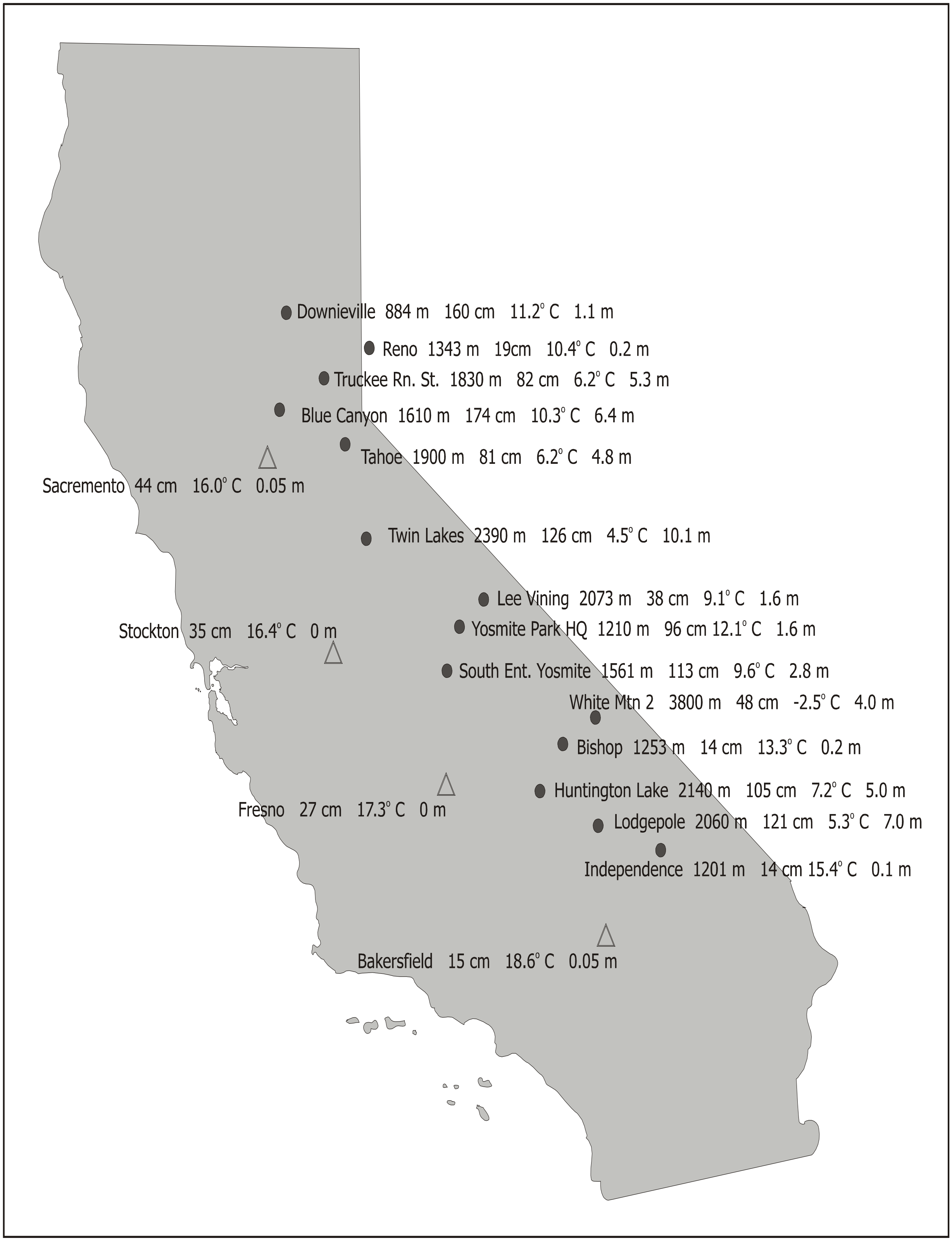
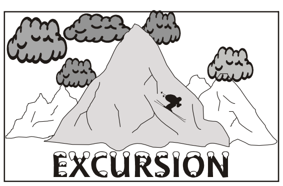


Figure 7.16- Climate data for the Sierra Nevada. Order of the data is: station name, elevation (m), annual precipitation (cm), annual temperature (C), annual snowfall (m). Note valley data for comparison.

Table 7.6- Sierra snowcourses and snowdepth

STATION	BASIN	elev (m)	FEB.	MAR.	APRIL	MAY
Upper Carson Pass	American	2590	155	201	208	147
Lake Lucille	Lake Tahoe	2500	---	358	378	---
Rubicon Peak 2	Lake Tahoe	2290	---	203	198	137
Ward Creek 3	Lake Tahoe	2060	165	211	234	188
Highland Meadow	Mokelumne	2682	201	267	287	256
Deadman Creek	Stanislaus	2820	---	---	221	196
Dana Meadows	Tuolumne	2990	155	198	218	150
Tioga Pass	Mono Lake	2990	142	162	193	---
Gem Pass	Mono Lake	3277	147	196	233	---
Mammoth Pass	Owens	2900	196	256	287	269
Bishop Lake	Owens	3445	104	147	173	155
Snow Flat	Merced	2650	191	246	264	229
Mono Pass	San Joaquin	3490	135	188	218	---
Bishop Pass	Kings	3415	150	216	234	218
Bighorn Plateau	Kern	3460	99	142	163	140



Sierra Corn Snow

The higher elevations of the Sierras are well known for their spring and early summer skiing. Corn snow is responsible for this worthy distinction. Corn snow is the name given to a type of large well-rounded snow particle that forms when the surface of the snow is subject to a consistent diurnal melt-freeze cycle. During the day, the direct solar radiation incident on the surface of the snow, as well as increasing daytime temperatures, produce a net positive energy balance within the upper layers of the snowpack that in turn produces ever increasing amounts of melt water. As snow particles near the surface melt, the water slowly percolates through the pore space of adjacent particles and into the layers below. In this water rich environment, larger snow particles tend to grow larger while smaller snow particles melt. By late afternoon, the upper layer of the snowpack is drained of most of its melt water. The water that does remain acts to bond the individual snow particles via surface tension. As the energy budget becomes negative (at night), the water-snow particle matrix freezes. In time (over a number of days) relatively large clusters of rounded snow particles form in the upper layers of the snowpack, these particles are what we call corn snow. Corn snow has a rounded shape because a sphere has the least surface area for any given volume of matter.

These rounded snow particles provide some very good skiing conditions. By the afternoon however, as melt water is produced within the snowpack, the upper layers of the snowpack become nearly saturated, and the snow becomes mushy and the skiing is over for the day. This also means that the avalanche danger increases throughout the day.

Why are the Sierras a great place for the development of corn snow? The Sierras have good corn snow development because of sunny days and cold clear nights which frequently occur at higher elevations in the spring and early summer. Sunny days provides abundant melt water for the ice-water mixture, and clear nights allows the longwave radiation given off by the upper layers of the snowpack to be admitted to space, driving the melt-freeze cycle.

The Sierras are a mountain range that does experience considerable amounts of summer convective activity. Backpackers and climbers should note that heavy rain, hail, and cloud-to-ground lightning are all common in the afternoon and early evening, from late June through early September. The area of greatest thunderstorm activity lies from the central part of the range to the crest of the Sierras. The reason that thunderstorms are much more frequent and vigorous in the Sierras than in the northern half of the Cascades is a result of a combination of factors. First, the middle and upper troposphere over the Sierras is a little drier, and secondly, there is considerably more elevated terrain in the Sierras's that acts as an elevated heat source once the ground becomes free of snow.

Sierra Nevada Weather Summary

* Wet season: November through March.
* Storm track is from the west or southwest at which time the polar jet will be located directly over central California.
* Summer convective activity (hail and lightning) is substantial, especially near the crest of the Sierra.
* Precipitation is infrequent compared to the Cascades or Coast Range, however, when it does occur, rainfall or snowfall are substantial.
* Barrier jets: well developed during winter when geostrophic flow is from west or southwest.
* Downslope windstorms: Common during winter east of Sierra crest. These events can produce very strong winds in the Owens Valley.

WEB: National Weather Service

Sacramento **www.wrh.noaa.gov/sacramento**

Hartford **www.wrh.noaa.gov/hartford**

Reno **www.wrh.noaa.gov/reno**

California Snow Cooperative **www.cdec.water.ca.gov/snow**

8

REGIONAL WEATHER SURVEY, PART II
Rocky Mountains and the Mountains of New England

Chapter Highlights:

- ✓ Detailed information on the weather and climate of the Rocky Mountains as well as the mountains of New England.

The Rocky Mountains (Rockies) are composed of a vast number of smaller mountain ranges that extend from central New Mexico and central Arizona, into British Columbia and Alberta. The Rockies are approximately 2500 km (1500 mi) in length and vary in width from 250 km (155 mi) to about 650 km (400 mi). The height of these mountains vary from range to range, the tallest peaks being located in Colorado and northeast Utah. In addition to the mountains, large portions of the Rockies consist of intermontane valleys and in some regions, elevated plateaus. In fact, most of the region which we call "the Rockies" consists of a plateau that has an average height of 1500-2000 m (4,900-6,500 ft). On top of this plateau are interspersed mountain ranges that provide an additional 1000-2500 m (4,900-8,200 ft) of relief. In order to facilitate this discussion on the weather and climate of the Rockies, the material has been divided into two sections. The first of which covers Utah, Colorado, New Mexico and Arizona in what we will call the southern Rockies, while the second covers all of the range to the north.

Since the Rockies cover such a large area, there are considerable regional weather and climate differences that should be taken into account in this type of survey. For example, when a major snow producing weather system moves into the Wasatch, it does not necessarily mean that the San Juans of Colorado are going to receive heavy snow as well. Likewise, when the Front Range of Colorado receives heavy snow, quite often the mountains of central Colorado only receive light amounts of snow. This chapter will consider a number of these weather regimes in some detail.

The Southern Rockies

To start this survey let us consider position of the polar jet stream over the southern Rockies. In general, due to the cooling of the interiors of the continents during the winter, the polar jet stream often reaches its southern most trajectory over the middle of the continents (North America, Europe, Asia). For example, when the polar jet moves on shore over the west coast in the winter, quite often its trajectory will continue southest into northern Arizona and New Mexico before curving back towards the northeast. When the polar jet stream extends into the southern tier of states it produces the coldest temperatures of the year, as cold Arctic air moves down from high latitudes. The day-to-day or even week-to-week position of the polar jet, as was discussed in Chapter 4, varies considerably. On any given day the polar jet could be located over Arizona or Alberta, or any point in between. As a rule of thumb, when the polar jet is located in the northern Rockies, you should not expect any major weather systems in the southern Rockies. This does not disqualify mesoscale or local-scale weather systems from developing. Skiers should note that it does not take a large weather system to produce 10-15 cm (4-6 in) of dry powder snow over the higher mountains of the southern Rockies. In fact, a considerable amount of the annual snowfall at most ski resorts in the region is produced by weak weather systems or from purely orographic lifting of moderately moist air.

Another important factor concerning winter snowfall in the southern Rockies is that the moisture is for the most part is transported from the Pacific Ocean. This means that it has had to travel either over the Cascades or Sierras as well as over the Great Basin. As a result, the lowest kilometer of the troposphere in the Rockies is very dry, the bulk of the moisture is in the middle-troposphere (700-500 mb). There is one important exception to this last point as we shall discuss in the section on the San Juan Mountains. The fact that the lowest kilometer (3,280 ft) or so is dry and the bulk of the moisture is carried in the middle-troposphere means that the lifting of air as it flows over large mountains is the primary producer of precipitation, more so in the Rockies than in the Cascades.

Figure 8.1 shows the principal storm tracks for the southern Rockies, including the summer monsoon flow from the Gulf of California. During the winter the predominate storm track is from the northwest-to-southwest. Other flow regimes include north to northeasterly flow which produces cold, relatively dry conditions. Residents of the eastside of the Rockies are well acquainted with low-level easterly winds which produce upslope flow. We will study this regime in some detail later in this section, nevertheless, when low-level easterly flow occurs in conjunction with mid-tropospheric flow from the southeast-to-southwest, the Front Range of Colorado generally receives its largest snowfalls of the season.

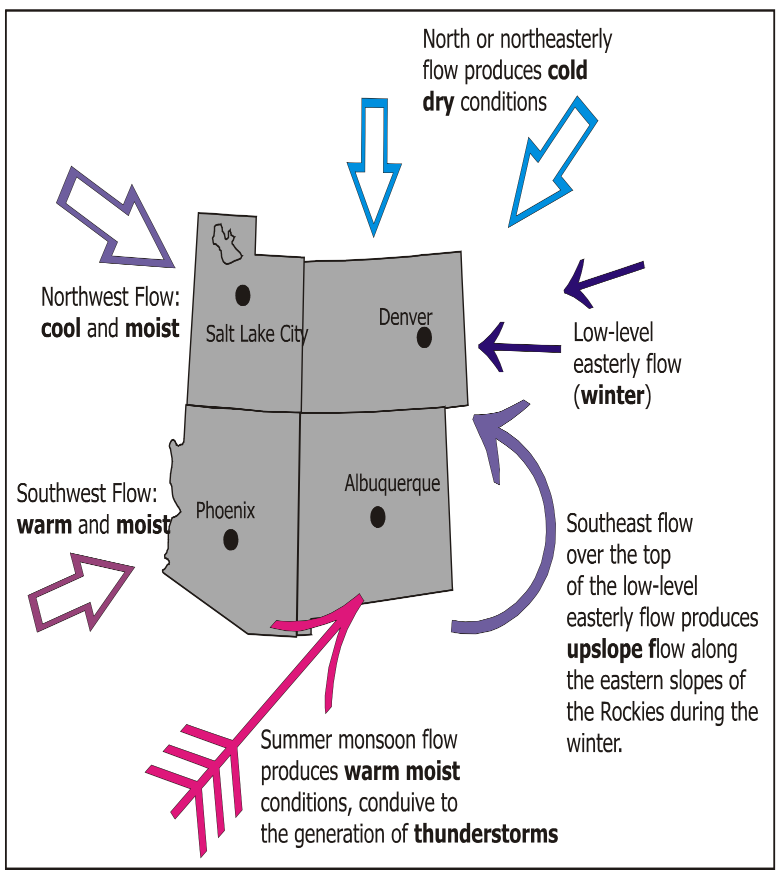


Figure 8.1- Southern Rockies flow regimes.

Figure 8.2 displays several 500 mb flow patterns for major precipitation producing storms. The actual precipitation received in any given mountain range within the Rockies, of course, depends on the path of the jet stream and direction of moisture bearing winds (Changnon *et al* 1993). In Figure 8.2a for example, this type of low amplitude ridge can produce moderate amounts of precipitation at higher elevations, with almost no precipitation occurring at lower elevations (Hjmerstad 1970).

Climate data for selected valley and mountain stations is displayed in Figures 8.3 and 8.4. The wettest region of Arizona for example is along the Mogollon Rim and the area around Pinetop. The mid-elevation climate stations from Sedona north all have similar amounts of annual precipitation. The temporal distribution of precipitation in this area is bi-modal (has two peaks). The first occurs in winter and the second occurs in the July-September period as a result of increased thunderstorm development. In addition, the number of days per year with a trace or more of precipitation, is noticeably higher in the region around Pinetop than it is in the area north of Flagstaff. This is a direct result of the increase in moisture in eastern Arizona, New Mexico and southern Colorado due to the transport of moisture from the Gulf of California during the second half of summer. In Figure 8.3 notice the difference in annual precipitation and temperature between Bright Angel Ranger station (2561 m or 8,400 ft) and the bottom of the Grand Canyon represented by Phantom Ranch (784 m or 2,570 ft). Air within the Grand Canyon tends to be dry, causing what precipitation that does fall into the top of the canyon to either evaporate or sublimate on its long descent to the bottom of the canyon.

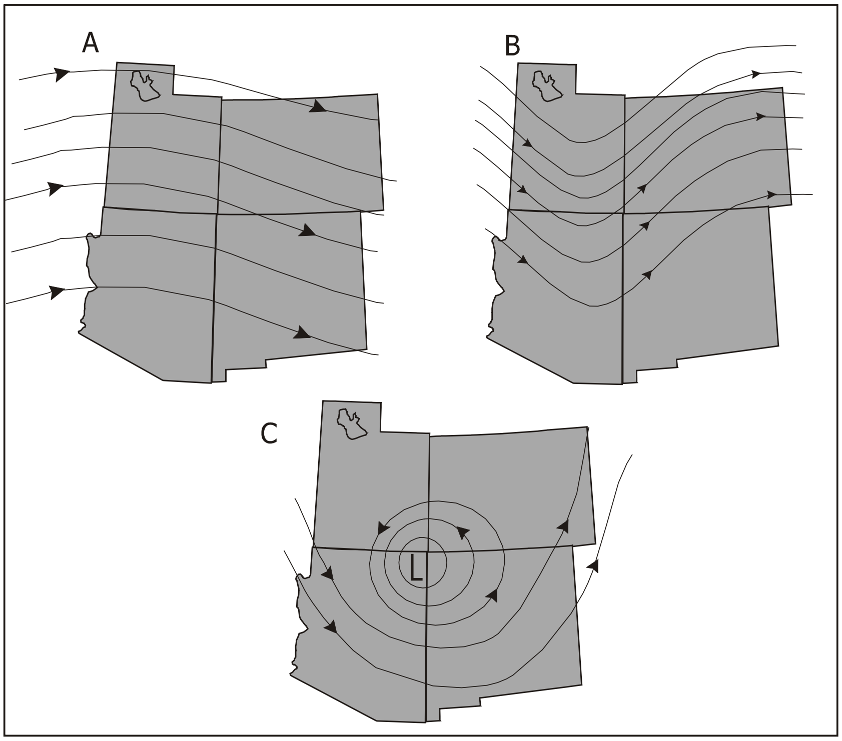


Figure 8.2- 500 mb flow patterns for major precipitation events in the southern Rocky Mountains. A) low amplitude ridge or zonal flow, B) shortwave trough, C) fourcorners cut-off low.

The White Mountains and Mogollon Rim are areas of high thunderstorm activity and cloud-to-ground lightning from July through early September. One of the interesting aspects of lightning strikes across Arizona is the time of day that the peak number of strikes occur, it varies widely across the state (Watson et al 1994). On the Mogollon Rim and White Mountains for example, the highest frequency of strikes occurs around 2 pm, in the central mountains around and to the southeast of Prescott, the peak occurs around 4 pm. In the southeast corner of the state a broad peak exists between 4 pm and 10 pm.

The wettest region in Utah is the Wasatch Mountains where from 100-150 cm (39-59 in) of annual precipitation is measured. This is should come as no surprise since the Wasatch is the first large mountain range that westerly flow encounters east of the Cascades or Sierras. The mountains of southern Utah are considerably drier than the Wasatch, in large part due to the fact that they are much smaller and hence do not disrupt westerly flow as much as their counterparts in the north. The seasonal distribution of precipitation in northern Utah is a function of elevation. Salt Lake City for example, has a slight spring precipitation maximum, although the month-to-month variation is not that large. Alta on the other hand has a very distinct winter precipitation maximum (December through March), with a typical winter month receiving 300-400% more precipitation than the typical summer month. The number of days with measurable precipitation in the Wasatch is about double that of the mountains and plateaus of southern Utah.

In northern New Mexico annual precipitation varies between 35 and 45 cm (14-18 in), for the 2000-2500 m (6,500-8,200 ft) elevation range as shown in Figure 8.4. The annual average increases as one moves toward the Colorado border. There is a prominent summer (July-September) maximum in precipitation in this region due to thunderstorm development. At many mid-elevation stations, winter is the driest of the four seasons. Average monthly winter precipitation at Santa Fe for example, is on the order of 25-35% of a typical summer month. Hikers and climbers should note that some of the highest frequency of cloud-to-ground lighting in the western US occurs in the mountains of New Mexico, especially in the Apache and Gila National forests as well as the northern mountains around Taos.

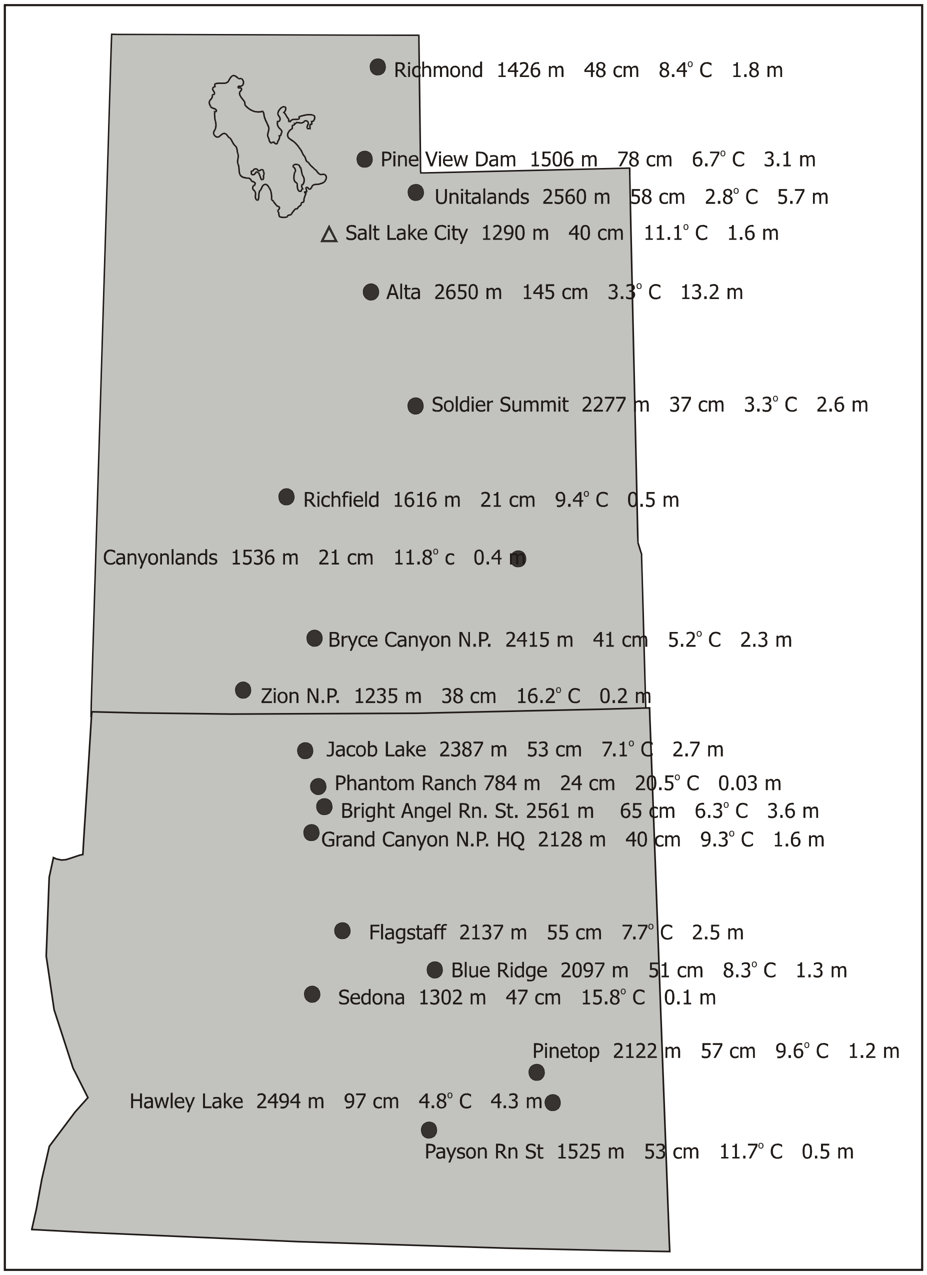


Figure 8.3- Climate data for northern Arizona and Utah. Data sequence is: elevation (m), annual precipitation (cm), annual temperature (C), and annual snowfall (m).

In Colorado it is difficult to identify any one particular range as being wetter than another.

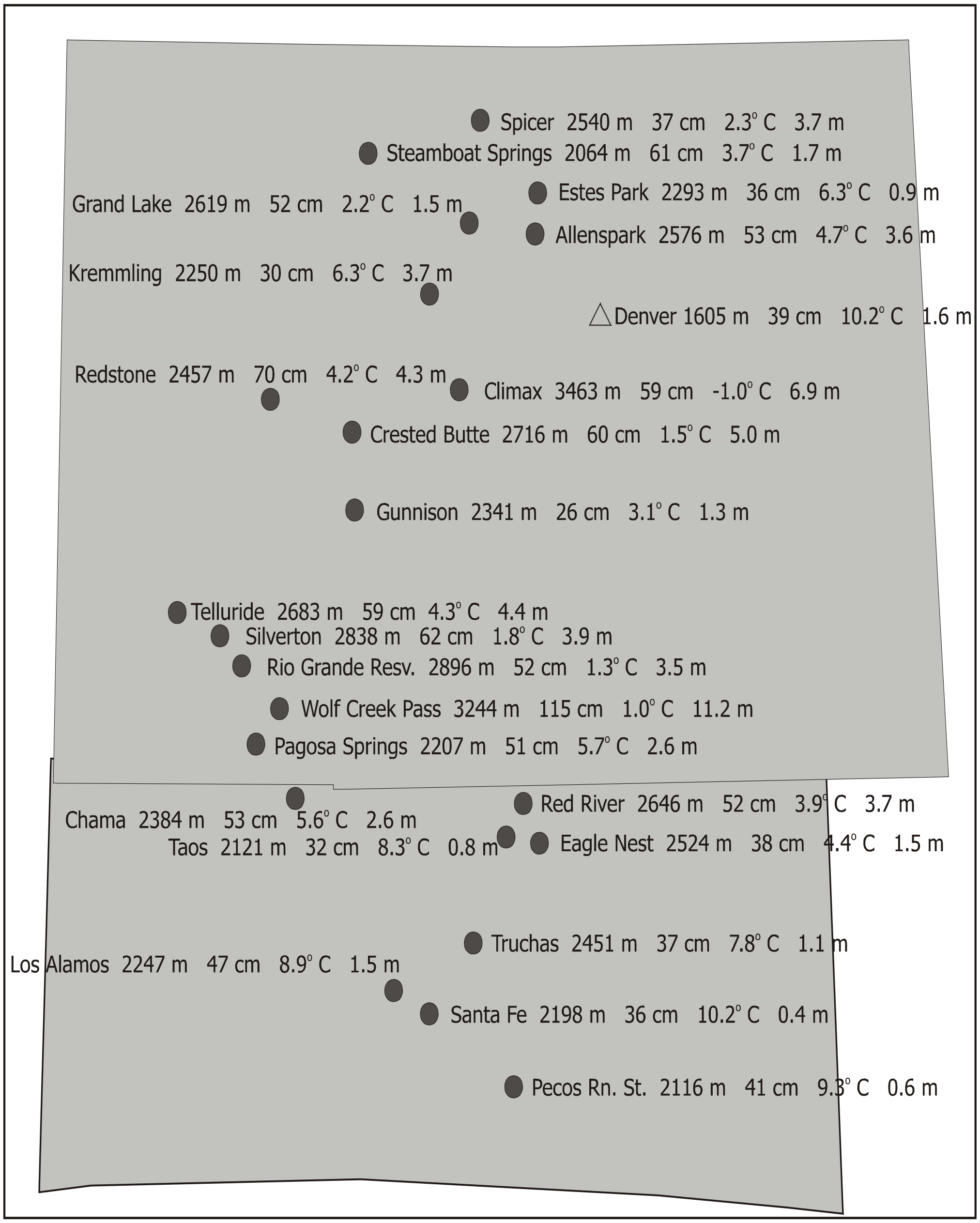


Figure 8.4- Climate data for northern New Mexico and Colorado. Data sequence is: elevaton (m), annual precipitaton (cm), annual temperature (C), and annual snowfall (m).

Annual precipitation in the Colorado Rockies is controlled by elevation and by the height of upstream terrain. When a storm moves across the state from a given direction, certain mountain ranges tend to be favored for precipitation while others are not. For example, when a low is located over the Four Corners region, moisture moves into Colorado from the southwest. This produces considerable amounts of precipitation in the San Juans, moderate amounts in the central Colorado Rockies and either light amounts or no precipitation in the northern ranges. With a low positioned in northern Utah, the northern half of the Colorado Rockies receives the bulk of the precipitation. In cases with moist northwest flow, the northern and central ranges tend to receive the heaviest precipitation.

Since the mountain ranges that comprise the Colorado Rockies vary in size and height, the amount of mountainous terrain upstream of a particular observation point should also be factored into the precipitation equation. This is particularly true in the San Juans and central mountains, where the leeward side of the range (with respect to prevailing storm track) can be quite dry while the windward slopes receive large amounts of precipitation.

Wasatch and Uintas

The Wasatch mountains receive considerably more precipitation than the Uintas because the Wasatch are upstream of the Unitas in relation to the westerly storm track. Annual precipitation in the Wasatch is 65-150 cm (25-59 in), and 65-75 cm (25-29 in) in the Uintas. Although the Wasatch mountains are quite narrow, they are the first substantial mountain barrier that moist westerly winds have to flow over east of the Cascades and Sierras. The height and especially the steepness of the Wasatch produces some of the wettest conditions in all of the southern Rockies. As a result, air moving into the Unitas contains considerably less moisture than it did when it first entered the Wastach.

In Figure 8.5 precipitation at six climate stations that form a west-to-east transect across the Wasatch Mountains is displayed. Notice the sharp increase in precipitation in the mountains, and the sharp decline in the lee of the mountains. The ratio of annual precipitation at the foot of the Wasatch (Midvale, Salt Lake City) compared to the mountains is on the order of 1:3.5. However, this varies from storm-to-storm, so be careful applying this ratio to every storm. Williams and Peck (1962) noted that the ratio is largest for storms that have large orographic lifting. In other words, synoptic-scale troughs and lows produce widespread weak uplift along frontal boundaries, which is independent of orographic lifting. When frontal lifting occurs the ratio of precipitation that falls at high and low elevations is reduced. Conversely, when frontal lifting is weak but orographic lifting is strong, the ratio increases dramatically. Figure 8.5 also displays the number of days per year where the precipitation meets or exceeds the 'moderate' threshold (>13 mm or 0.5 in). As you can see there is about a fourfold increase in the number of days meeting or exceeding this criteria from the lowlands to the mountains. This example shows just how much influence steep terrain has on the formation of precipitation.

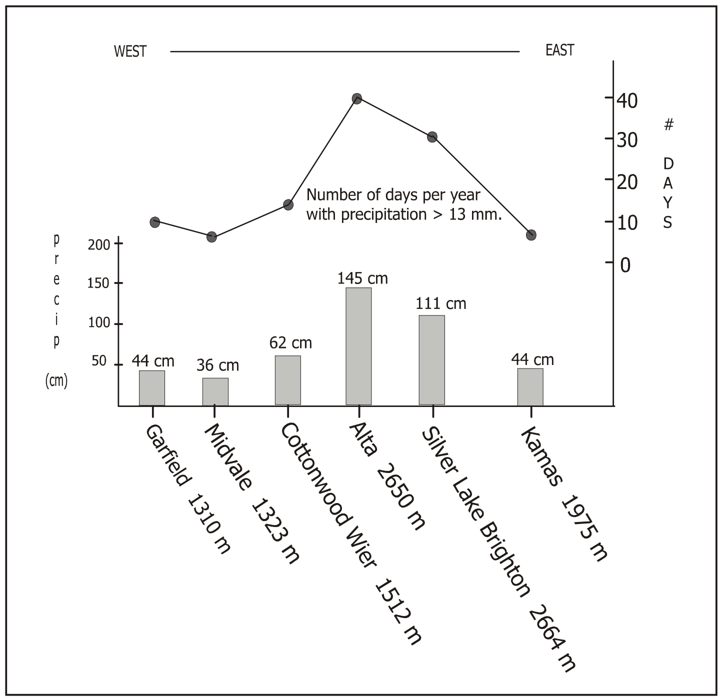


Figure 8.5- West-to-east transect across the Wasatch Mountains. Vertical bars represent annual precipitaton (cm) and the dots are the number of days per year with precipitation exceeding 13 mm (0.5 in).

Snowcourse and snotel data for the Wasatch and Uintas indicates that the amount of snow on the ground in the spring and early summer is a function of not only the amount of winter snow accumulation, but the amount of melting and snowpack consolidation that has occurred during the spring. The amount of spring melt is tied into the mean daily temperature at the site, which is controlled by elevation and cloud cover. The areas with the deepest April 1st snowpack are the central Wasatch in the vicinity of Alta and Snowbird, where the long-term average is about 250 cm (98 in). In addition, the snowcourse at Ben Lomond Peak has a long term average of 245 cm (96 in). The snotel sites in the Uintas on the other hand, which are located at a mean elevation of about

2800 m (9,200 ft), average 75-100 cm (30-40 in) of snow on the ground as of April 1st, with several of the higher sites (3300+ m) averaging 150 cm (59 in). We examined the daily precipitation and snowfall records for climate stations such as Alta and Silver Lake Brighton to see if any winter trends could be detected. What we found is that during the December trough March period, there is no multi-day period that has a better chance of receiving new snow than any other period. Therefore, when you book that week long ski vacation five months in advance, feel at ease that the time slot you selected has a equal chance as any other period when it comes to receiving fresh powder on the slopes.

In the summer months, northeast Utah is a region of considerable convective activity in the lowlands as well as in the mountains. Cloud-to-ground lightning, as well as moderate to large hail, are all real threats to mountain travelers. As we noted in Chapter 5, convective activity generally starts in the early to mid-afternoon and diminishes in late evening or at night. About the only thunderstorm data collected by the NWS occurs at first order weather stations, like Salt Lake City (SLC) airport. The peak month for thunderstorm activity at SLC is August, followed by July, with June and September being in a close tie for third.

San Juan Mountains

The San Juan Mountains of southwest Colorado extend some 200 km (120 mi) west-to-east and about 125 km (75 mi) north-to-south. This range is large enough that there are distinctive areas that receive heavy or light amounts of precipitation from a given storm, depending on the position of the jet stream. For example, when Wolf Creek Pass (3244 m or 10,600 ft) receives a large amount of new snow, Telluride usually receives considerably lighter amounts. When a 500 mb trough or low is positioned over Arizona, as depicted in Figure 8.2a,b the stage is set for the southern San Juans to receive large amounts of snow. This is a favorable pattern for the generation of heavy snow because moisture from the Gulf of California and Pacific Ocean off of Mexico, is added to the moisture that was transported with the original storm.

It is instructive to compare climate station data taken from Pagosa Springs (2207 m or 7,240 ft) with data from Wolf Creek Pass. From Figure 8.4 note that Pagosa Springs receives about 44% of the annual precipitation that is measured at Wolf Creek Pass. The average number of days per year with a trace or more of precipitation is 97 at Pagosa Springs and 117 at Wolf Creek Pass, which are fairly similar. However, the number of days with moderate or heavy precipitation at Wolf Creek Pass is three times (29 days) the number of days at Pagosa Springs (10 days). This comparison shows the ability of steep terrain to greatly enhance the background precipitation on the local-scale. It is also interesting to note that Telluride (2683 m or 8,800 ft) and Rico (2698 m or 8,850 ft) have about the same number of days with light amounts of precipitation as Wolf Creek Pass, but less than half of the number of days with moderate or heavy precipitation.

The western and northwestern regions of the San Juan's can receive moderate snowfalls from a southwesterly flow, but in order to do so, a trough or low has to be positioned over southwestern Utah-northwestern Arizona. Depending on the strength of the storm, moisture from the Gulf of California and the Pacific Ocean off of Baja California may or may not be transported into the region. Flow from the west and northwest tends to favor heavier snow in the western and northern San Juan's, at which time stations like Wolf Creek Pass tend to receive light amounts of snow.

The monthly distribution of precipitation between various stations in the San Juan's is interesting. For example, at Silverton (2838 m or 9,300 ft) and Telluride the wettest month is August, compare this with Wolf Creek Pass where the wettest month of the year is March.

In the following paragraph we will present a short case study of a storm that occurred between March 28-30, 1998, and which produced widespread heavy snowfall across the San Juan's. The 500 mb low center was located over central California and southern Nevada during the time of heaviest snowfall (Figure 8.6). The polar jet axis was positioned over central New Mexico, nevertheless the 500 mb winds in the southern San Juan's were from 30-40 ms^{-1} (65-90 mph).

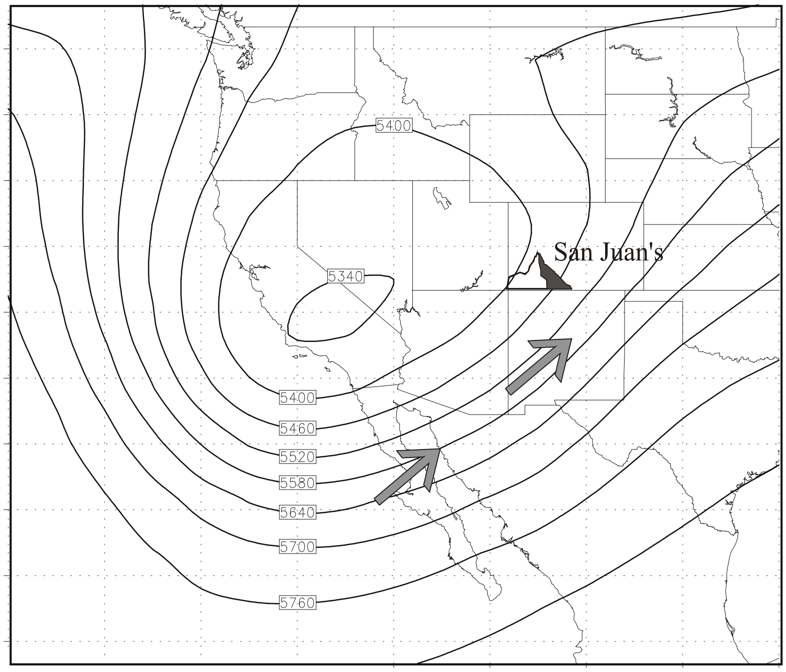


Figure 8.6- March 29, 1998 500 mb height field. Arrows indicate core of jet stream winds.

Figure 8.7 shows precipitation amounts for select snotel sites around the San Juans. Notice how the southern and southwestern stations received the largest amounts of snow (measured as water equivalents), while the northwestern and northeastern stations received much lighter amounts. The different precipitation totals measured at Red Mountain Pass, Lizard Head Pass, and Molas Lake is probably due to local orographic effects, rather than due to any significant differences in the synoptic-scale flow or the transportation of moisture from afar.

Basically, there are two precipitation maxima in the San Juan's, the first occurs in March and is a result of the southern trajectory of the polar jet stream. The second maximum is in August and corresponds with the southwest monsoon. The southwest monsoon (which is sometimes called the Mexican Monsoon) starts to develop sometime around late June or early July. When it occurs, southwest winds transport moisture into the region from the eastern Pacific, and the Gulfs of California and Mexico (Douglas *et al* 1993). There are times when this moisture is transported as far as northern Utah and southern Wyoming, but it primary impact is over Arizona, New Mexico and

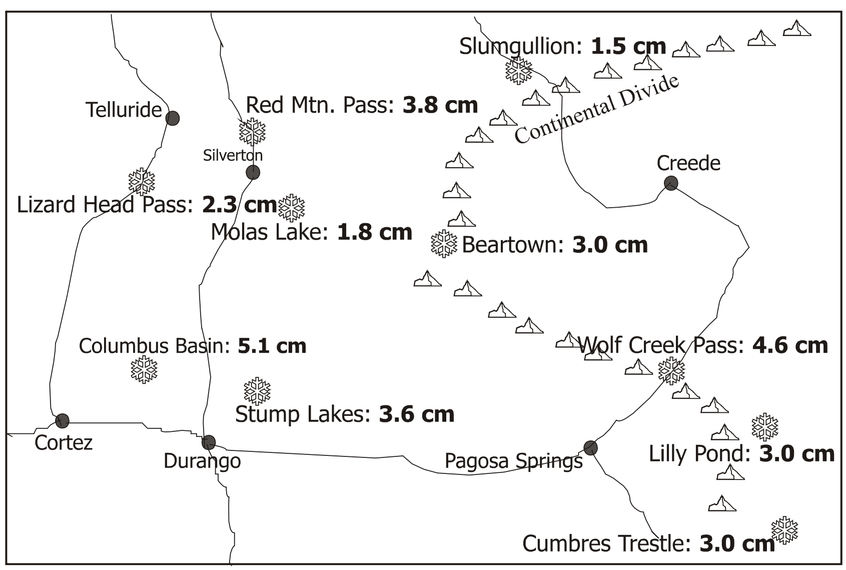


Figure 8.7- Precipitation (water equivalent) measured at select snotel sites in the San Juan's during the March 29, 1998 snowstorm. One centimeter of water equivalent equals anywhere from 10-20 cm (4-8 in) of snow.

southern Colorado. During mid-summer the subtropical jet stream moves northward to a position around 25°- 30° N. With cooler dry air to the north of the subtropical jet, the stage is set for the outbreak of vigorous convection and thunderstorms over the southern Rockies. Most thunderstorms develop on days when the mid-tropospheric winds are light to moderate. The driest months in the San Juans are May and June. This corresponds with a period when the polar jet stream has moved north and before the onset of the summer monsoon.

Temperatures in the San Juans are a function of elevation to a large degree. However keep in mind that during the winter months, the air in the bottom of valleys may be as cold or colder than at elevations 500 m or even 1000 m higher (1,640-3,280 ft), due to persistent valley inversions. A comparison of monthly temperatures between Wolf Creek Pass and Telluride indicates that the monthly mean temperatures for each station are similar, as one would expect based on the fact that both stations are located within the same climate zone. However, when comparing the monthly temperature range (average monthly high and average monthly low), Telluride has a noticeably larger range. In fact, monthly lows at Telluride are equivalent to those at Wolf Creek Pass which lies some 560 m (1,800 ft) higher, due to persistent temperature inversions at Telluride. During the summer months, Telluride is warmer due to its lower elevation and due to the fact that the heating of the valley keeps the valley atmosphere several degrees warmer than the free atmosphere.

Central Colorado Rockies

This group of mountains lies between Grand Junction to the west and Denver to the east, with Gunnison to the south and Interstate 70 to the north. The region is split in two by the Continental Divide, which has a height of about 4100 m (13,500 ft). This is the area that Hjermstad (1970) did his snow distribution study that we have referred to a number of times previously. In that study, the author found that the distribution of snow across this region is in large part controlled by the mid-level (700-500 mb) wind direction. For example, along the Western Slope at elevations below 2500 m (8,200 ft), southwest flow produces the biggest snowstorms of the season. Above 2500 m (8,200 ft) however, the biggest snowfalls of the year generally occur when the flow is from the northwest. This does not mean that the mountains do not receive any snow when the snow is from the southwest, it simply suggests that the observed amounts are less than when the flow is from the southwest. East of the Continental Divide, heavy snowfall events occur when the mid-tropospheric flow is from the southeast or south. Overall, annual precipitation decreases from west-to-east in the central mountains.

Why is there such a difference in the spatial distribution of winter precipitation between southwest and northwest flow? We have mentioned this difference in earlier sections, and will now give a brief explanation. Figure 8.8 depicts southwest and northwest flow patterns with the associated vertical motion fields (ascending and descending regions). With a trough (or low), there is ascending motion to the east of the trough axis and descending motion to the west. These vertical motion patterns are a result of the a number of atmospheric processes, which we will not attempt to explain at this time. The region of ascending motion is linked to the development and movement fronts. From Chapter 5 you should recall that precipitation forms in areas of ascending air. Therefore, in large synoptic-scale storms like the one depicted in Figure 8.8a, precipitation forms by frontal lifting, which is independent of orographic lifting. This should not be understood to mean that precipitation formed in areas of synoptic lifting, cannot be further enhanced by orographic lifting. It does however explain why areas upstream and downstream of mountains (non-orographic), receive precipitation during this type of flow pattern.

In Figure 8.8b, a small amplitude ridge (this feature can also be called a shortwave ridge), and its associated vertical motion field is displayed. Notice that when the flow is from the northwest, the overall motion is downward to the east of the ridge axis. This type of synoptic pattern in and of itself does not produce very much precipitation. However, when this type of flow moves over a mountain range, low and mid-tropospheric air is forced over the mountains, the ascending motion can be about 10 times as great as the synoptic-scale descending motion. The net result are localized areas of strong orographic lift and the subsequent development of precipitation over the mountains. It is for this reason that northwest flow produces little snowfall at low elevations, but can produce substantial snowfall at higher elevations. This also illustrates the important point that you have to be careful correlating weather data from the low elevation stations (Glenwood Springs for example), to the mountains (Vail Pass); sometimes there is a high correlation, but most of the time the correlation is pretty low. In addition, since many storms contain both a trough and a ridge as depicted in Figure 8.8, there may be periods during the storm when frontal lifting may dominate, and other times when orographic lifting overwhelms all other types of vertical motion. Also, the path which a particular storm takes will determine what mountain ranges receive heavy precipitation and which ranges get bypassed.

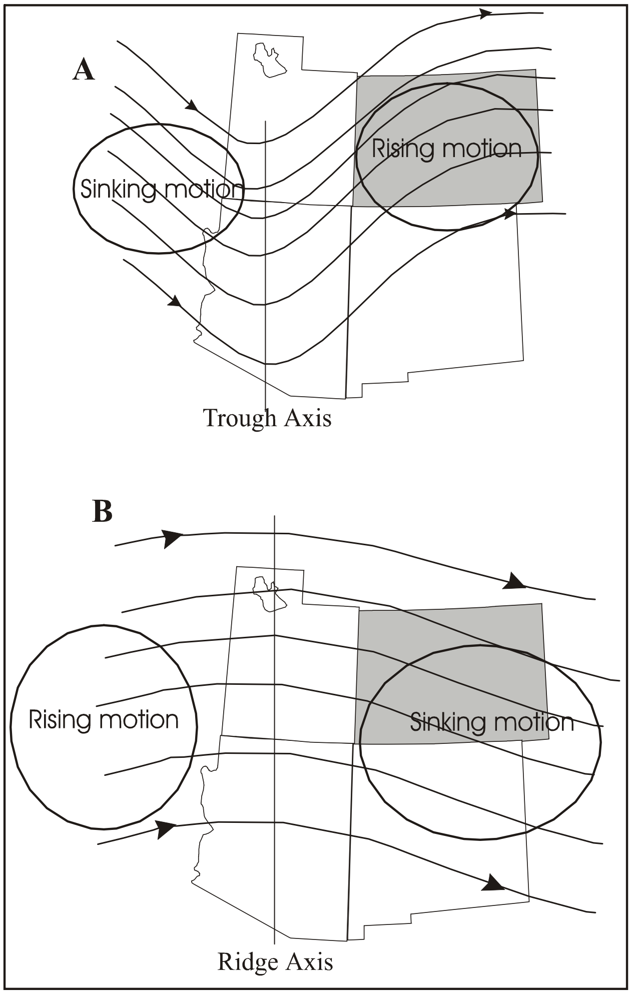


Figure 8.8- Vertical motion fields associated with synoptic-scale disturbances. A) trough pattern. B) ridge pattern.

The one high elevation climate station in the central mountains is at Climax (3463 m or 11,360 ft), which should be quite representative of most of higher elevations within this region (Figure 8.9) The wettest month on average is April, although from the November through May monthly totals are quite similar. By June there is a definite drying trend, this is short lived however, as the July and August convective season produces an increase in rainfall. The driest period is September and October. With regard to mean monthly temperatures, Climax runs about 2-3^{o} C (3.5-5.5^{o}) cooler than Wolf Creek Pass, primarily due to the fact that the station at Climax is 220 m (720 ft) higher and several hundred kilometers north.

Note that ranges such as the Sangre De Cristo Mountains are very dry due to their location downstream of the San Juans and central mountains, with respect to westerly winds. Likewise many of the valleys that cris-cross large mountain ranges tend to be very dry as well. A prime example is the along Highway 24 between Leadville and Salida. This is a deep narrow valley, where most of the moisture falls on the higher terrain bordering the valley. Because of very dry air within these valleys, what little precipitation that does fall over the valley often evaporates or sublimates before it reaches the ground.

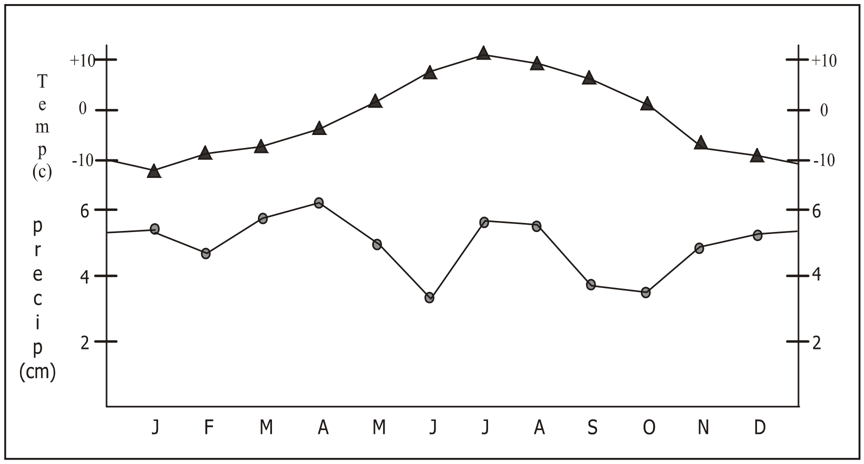


Figure 8.9- Mean monthly precipitation (circles) and temperature (triangles) at Climax, Colorado.

Thunderstorm formation over Colorado has been studied by Banta and Schaff (1987), they suggest that thunderstorms are best developed on days when the 500 mb wind speed is less than 12 m/s (26 mph). This same study also indicates that thunderstorm development also depends to some degree on the wind direction.

<u>Front Range</u>

The eastern slopes of the Rockies are distinct from the rest of the range when it comes to winter precipitation. Local residents know that easterly flow is required in order for the Front Range and foothills to receive any significant snowfall. There are two basic flow patterns that can produce snow in the region, both involve low-level flow from the east. In reality the surface winds can range from north-to-southeast, with northeast and east being the most common. This occurs when a surface high is located over the central or northern Midwest. As this cold air moves towards the Front Range, it slows down and forms a blocked layer that extends into eastern Colorado. Since this shallow cold air is also quite dry, moisture has to be transported into the region from another source. It turns out that moisture is transported into the region by a upper level low or a trough located in northern Arizona or northwestern New Mexico. This pattern often produces southeast flow (700-500 mb) over the top of the cold air located at the surface (Figure 8.10a). When this occurs it is referred to as *overrunning*.

As the moist southeast flow is lifted up and over the cold air dome, precipitation forms in areas of uplift. Typically, clouds that form in this process have a cigar or elongated shape, some 100 km (62 mi) in length and 50 km (31 mi) in width (snow bands). To an observer on the ground snow bands may not be distinguished from the overcast skies, nevertheless they do show up very well on weather radar. These snow bands move around as the storm evolves, causing snowfall distribution across the Front Range to be very inhomogeneous. For example, it is common for an area to receive moderate to heavy snow, while an area 20 km (12 mi) away receives light amounts. In addition, with this type of pattern it is not uncommon for the region between Interstate-25 and the foothills to receive considerably more snow than the Front Range itself.

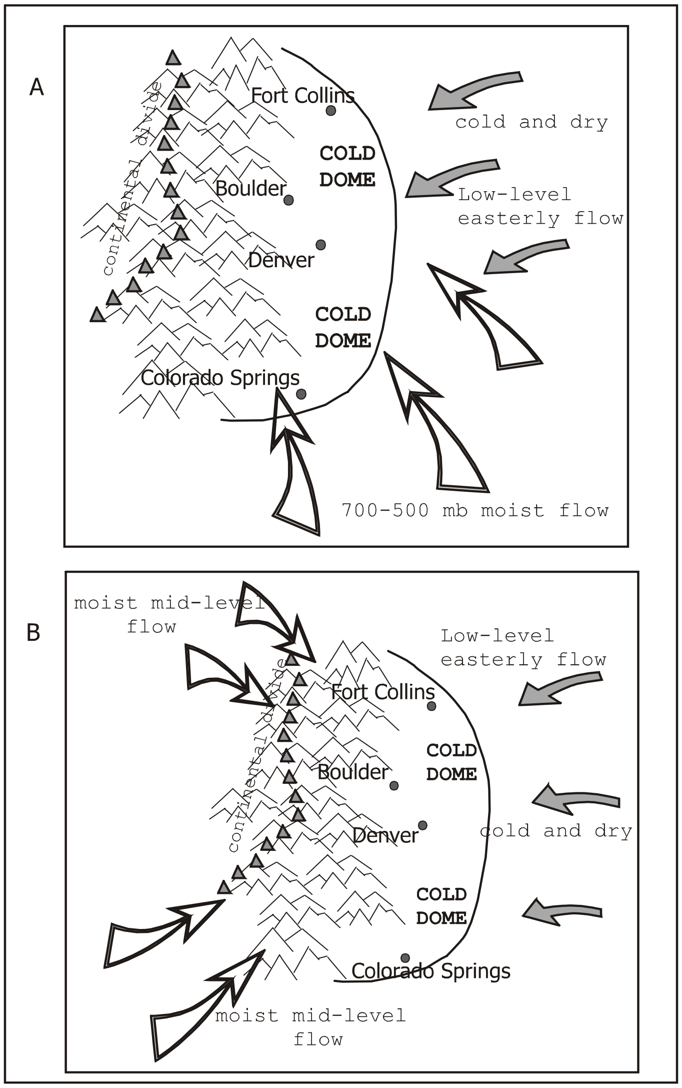


Figure 8.10- Two Front Range snow producing flow regimes. A) low-level easterly flow (upslope) with 700-500 mb flow from the southeast suppling the moisture. B) like in (A) but moisture is either from the southwest or northwest.

The second type of Front Range storm is a slight deviation from what was just described. In this new scenario cold air damming occurs over eastern Colorado, but now the middle and upper-level flow is from either the northwest or southwest (Figure 8.10b). Without the presence of a dome of cold air which is typically 500 m to 1000 m deep (1600-3200 ft), upper-level winds would descend down along the eastern slopes of the Front Range, creating warm dry conditions at the surface.

Of course there are multiple variations of the two storm patterns that were just described. In addition, smaller-scale terrain features such as Palmer Divide and Cheyenne Ridge, help modify storms by either producing upslope or downslope flow. When heavy snowfall does occur in the Front Range, it is usually confined to the east side of the Continental Divide. Generally, a storm that produces moderate or heavy snow in the vicinity of the foothills, would only be expected to produce light snow over Winter Park.

It should also be noted that since the distance from the Continental Divide to the base of the Front Range is on the order of 50 km (31 mi), there are locations in the middle of this region that do not receive very much precipitation during the winter at all. In addition, the bottom of deep canyons are also very dry, this becomes evident as one drives up the Poudre and Big Thompson canyons, and the Interstate 70 corridor near Idaho Springs.

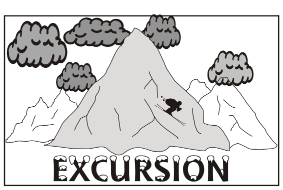
The seasonal distribution of precipitation in the Front Range is highly correlated to elevation. A number of studies have indicated that near the Continental Divide, the maximum in annual precipitation occurs during the winter months. At elevations below 3000 m (10,200 ft) maximum precipitation occurs in April and May as a result of convective rainfall. Along the base of the Front Range maximum precipitation shifts to July and August, this time in response to large convective systems that form to the lee of the mountains, as moisture is transported into the region from the Gulf of Mexico.

Temperature patterns in the Front Range are consistent with the rest of the southern Rockies. In mid-winter expect air temperatures at higher elevations to range between -10^{o} C and -16^{o} C (12^{o} and 3^{o} F). During mid-summer temperatures often range from 3^{o} C to 13^{o} C (38-55^{o} F) at these same elevations.

The Front Range is also well known for its strong downslope windstorms that occur during the cooler months of the year. The requirements for a downslope windstorm are strong westerly winds in the 700-500 mb layer, and a stable layer or inversion just above the Continental Divide. If a dome of cold air is entrenched to the east of the Front Range, then as was noted above, downsloping winds will have a difficult time reaching the surface. The strongest surface winds occur in areas where the descending winds are funneled through gaps and passes in the mountains. Over the course of a typical winter, 5-8 downslope storms will occur in this region, with one or two of these storms able to produce damaging winds. During these more powerful events, windspeeds of 40-50 ms^{-1} (90-125 mph) in the foothills are possible. Most of these events last 6-12 hours. Obviously these wind events play havoc with the snowpack in the Front Range, creating large drifts in places and scouring the snow in others. The distribution of snow in the Front Range on the local-scale is a function of the wind more than any other factor. Down at the base of the Front Range some of these high wind events turn out to be chinooks, where there is a dramatic rise in air temperature, and a significant melting of the snowpack.

Cold air damming often leads to the formation of barrier jets along the east side of the Front Range. Since the low-level flow is from the east, unlike other ranges where barrier jets are common, this jet flows from north-to-south. This features are not that common, however when they do occur they can produce moderately strong northerly winds (10-15 ms^{-1} or 22-44 mph) between the Front Range and Interstate 25.

Like all areas in the southern Rockies, the Front Range has frequent thunderstorms during the summer months. From mid-April through early June, small cumuli form over the Front Range by mid morning. As the day progresses these clouds grow in size and move eastward over the plains. Rainfall during the summer months in the Front Range typically occurs in the late afternoon and early evening hours. Over eastern Colorado in contrast, rain typically occurs in the late evening and at night. Since thunderstorms are frequent in the Front Range, there is naturally also a high incidence of cloud-to-ground lighting strikes. Most bolts make their appearance in the late afternoon and in the early evening hours. An old Chinese proverb we once heard sums it up pretty well: “hike early, or die later.”



Wind Drift Glaciers of the Eastern Rockies

South of the Canadian border, the Rocky Mountains are pretty much devoid of glaciers except in a few locations such as Rocky Mountain National Park (RMNP), the Wind River Range, and Glacier National Park. In order for a small alpine glacier to exists, snow accumulation must exceed losses due to melting. A glacier does not disappear the first few years that melting exceeds snowfall, nevertheless, it does not take long for a small glacier to begin to retreat or thin. If the southern Rockies are not glacier-friendly at the present time, then why do small cirque glaciers exist in select locations? The answer can be found in one word: wind. Note that the three mountain regions that do have small cirque glaciers, straddle the Continental Divide. For example, as the Continental Divide cuts across RMNP, it tends to be quite broad, several kilometers in width as a matter of fact. This high elevation plateau is a good place for snow to be deposited. The problem is that being a high elevation site, it also has a high frequency of strong winds. These westerly winds blow freshly *deposited snow eastward off of the divide. As air flows over the steep east slopes of the Divide, the wind loses its ability to carry the snow. As a result, the snow is deposited on the lee slopes, especially in areas where the terrain forms an amphitheater or a cirque (Latin for "bowl shape."). By late spring or early summer the amount of snow accumulation on the Taylor, Andrews, or Tyndall glaciers in RMNP can exceed the surrounding non-glacier, non-wind blown areas by 700% (Outcalt 1965). Just as the strong winds nourish these small glaciers, excessive amounts of wind blown snow on the leeward slopes of the Continental Divide also creates high avalanche danger that can last well into early summer.*

In the present climate regime, glacier development in the southern Rockies is not possible because conditions are too dry. It is interesting to speculate what increase in snowfall would be required before glaciers could develop. It would probably require at least a doubling of the present winter precipitation, and a substantial decrease in summer temperatures as well.

Northern Colorado

The northern part of the Park range, specifically the higher elevations to the east of Steamboat Springs, is one of the wettest locations in Colorado. The area around Buffalo Pass receives somewhere on the order of 120 cm (47 in) of precipitation during the winter, with an annual precipitation near 165 cm (65 in). It is not hard to see why when you look at a map. The Park Range is the first major topographic feature that northwesterly flow encounters in hundreds of kilometers. It is interesting to compare this area with Wolf Creek Pass. The Park Range has more days with light snowfall, but Wolf Creek Pass has more days with heavy snowfall. This fits in well with the conceptual model that the southern San Juans receives its major snowstorms from moist southwest flow, while the Park Range which primarily derives its snow from drier westerly and northwesterly flow. Since westerly and northwesterly flow is drier than southwesterly flow, it takes more days of light snowfall to make up the difference. As it happens, westerly and northwesterly flow is more common over northern Colorado than southwesterly flow is in the San Juans. The net result is that the area around Buffalo Pass is the snowiest location in the state.

Southern Rockies Weather Summary

* Synoptic-scale snowfall distribution primarily a function of mid-level wind direction and elevation. Areas with little upstream terrain generally receive the most snow.
* Mesoscale and local-scale snowfall distribution is a function of orographic enhancement and wind speed. The amount of snow on the ground at any given time is highly dependent on wind redistribution.
* When a trough or low is positioned over the Four Corners region, considerable amounts of sub-tropical moisture is added to the flow, producing heavy snowfall events in the San Juans.
* Most of the region has two precipitation maxima, one in the winter and the other in late summer.
* Since many of the higher peaks in the southern Rockies lie in the middle-troposphere, strong winds are common, especially during the winter.
* From July through September, as the subtropical jet stream migrates northward, moisture from the Pacific Ocean and Gulf of California is transported over the Southern Rockies. This monsoonal flow produces frequent and intense thunderstorms over the mountains.
* Cloud-to-ground lighting is <u>frequent</u> throughout the region during the summer.
* Downslope windstorms and chinooks are most notable in the Front Range during the cooler months of the year.

WEB National Weather Service

Flagstaff	**www.wrh.noaa.gov/flagstaff**
Salt Lake City	**www.wrh.noaa.gov/saltlake**
Grand Junction	**www.crh.noaa.gov/gjt**
Albuquerque	**www.srh.noaa.gov/abq**
Denver	**www.crh.noaa.gov/den**
Pueblo	**www.crh.noaa.gov/pub** (see their lightning page)

Niwot Ridge, Colorado-	**http://culter.colorado.edu:1030**
Colorado Climate Center-	**http://climate.atmos.colostate.edu**
Colorado Avalanche Center-	**http://geosurvey.state.co.us/avalanche**
Utah Avalanche Center-	**www.avalanche.org/~uac**
No. Utah Avalanche info.-	**www.usu.edu/braic**
So. Utah Avalanche info.-	**www.avalanche.org/~lsafc**

Forest Service National Avalanche Center- **www.avalanche.org/~nac**

The Northern Rockies

The weather and climate of the northern Rockies is in many respects distinct from the climate of the southern Rockies. In fact the climate of northern Idaho and northwest Montana, is more closely related with the climate of the Cascades, than with the rest of the Rockies to the south. This is not too surprising since the mountains of northern Idaho lie within 500 km (300 mi) of the Pacific Ocean. In addition to differences between the northern and southern Rockies, there are some major west-to-east differences in precipitation and temperature regimes within the northern Rockies.

Select climate stations that are located in or adjacent to the mountains are shown in Figures 8.11,12. The decrease in annual precipitation from the northwest to the southeast is evident in these figures. The winter storm track in the northern Rockies ranges from the southwest-to-northwest, the same as it is throughout western North America. The area surrounding Yellowstone National Park (referred to as Yellowstone), for example, receives its largest winter snowstorms when the flow is from the southwest. In contrast, large snowstorms occur in Glacier National Park (Glacier) when the storm track is more from the west. Mountains further to the east are favored for winter precipitation when the flow is either from the north-northwest or south-southwest. In addition, the eastern slopes of the northern Rockies receive snow from upslope events (easterly low-level flow), similar to what we described in the section on weather patterns of Colorado's Front Range.

The two wettest regions within the northern Rockies are northern Idaho and northwest Montana, in the vicinity of Glacier. These two regions receive roughly 60% of their annual precipitation between November and March. In contrast, the annual precipitation in Yellowstone is evenly distributed throughout the year (Figure 8.13). The eastern slopes of the northern Rockies on the other hand located within a continental climate zone, receive their heaviest precipitation in May and June.

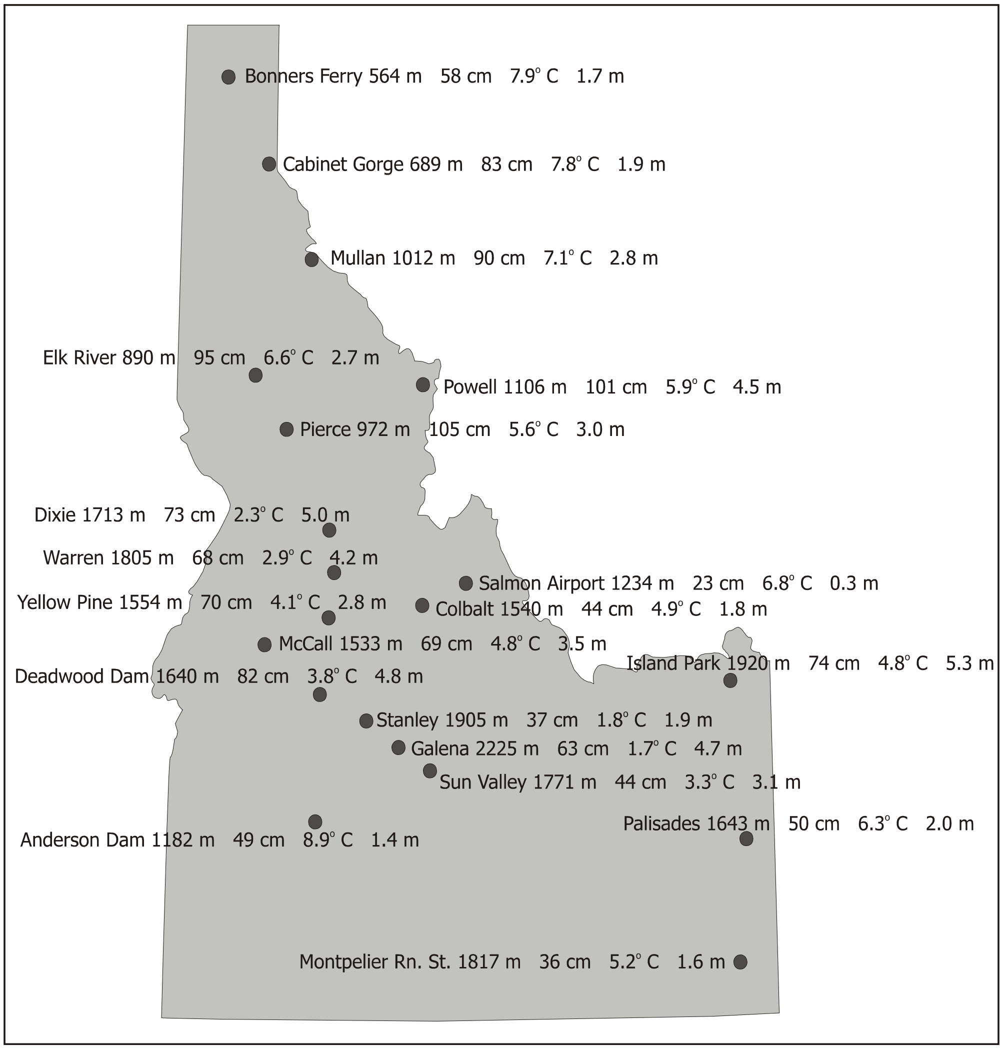


Figure 8.11- Climate data for Idaho. Data format is: station name, elevation (m), annual precipitation (cm), annual temperature (C), and annual snowfall (m).

The winter precipitation maximum in the western mountains occurs as a result of the polar jet stream transporting moisture inland from the Pacific Ocean. The polar jet stream migrates north and south over the northern Rockies between the months of October and April. Since the Rockies are some 300-600 km (200-400 mi) wide, the amount of Pacific moisture that is carried to the eastern mountains is greatly reduced. The May and June precipitation maximum in the eastern mountains can be explained in the following manner. In late spring and early summer there is rapid heating of the lower atmosphere over the northern Great Plains. In the mid-troposphere however, the air remains cool and dry, this makes the troposphere over the eastern mountains conducive to generation of cumulus clouds. Evaporation and sublimation of the mountain snowpack in May and June provides the moisture for the generation of towering cumulus and cumulonimbus clouds. By July the moisture supply in the mountains is diminished, which in turn causes a reduction in convective rainfall during the remainder of the summer. You may be wondering if the southwest monsoon, which begins in July in Arizona, has any effect on precipitation in the northern Rockies? Overall its impact is minimal, although from time to time large amounts of water vapor is transported into the northern Mid-West and Great Lakes regions.

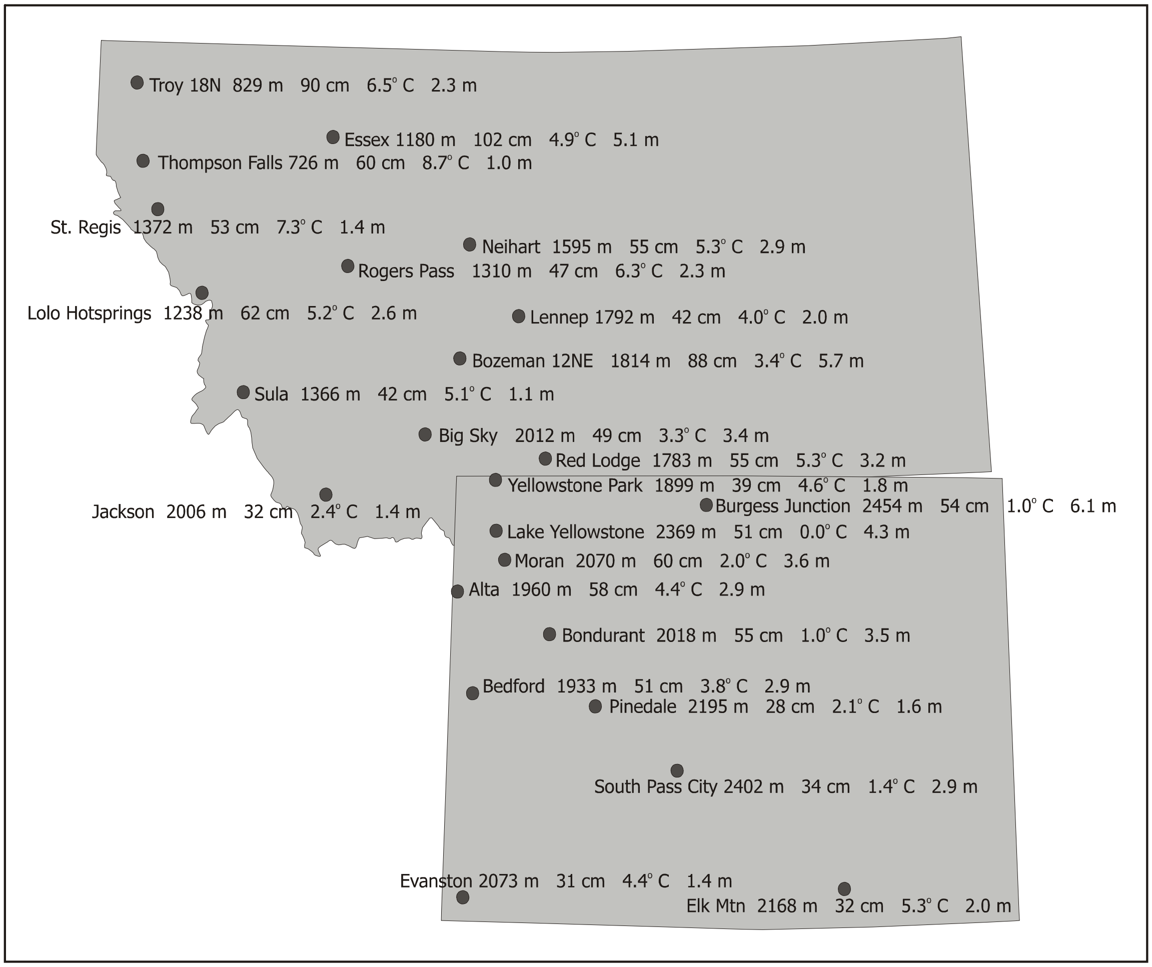


Figure 8.12- Climate data for Wyoming and Montana. Data format is: station name, elevation (m), annual precipitation (cm), annual temperature (C), and annual snowfall (m).

Temperature patterns in the northern Rockies are primarily controlled by elevation and secondarily by latitude and longitude. Northern Idaho and northwest Montana are influenced to some degree by the Pacific Ocean, this results in slightly warmer winters and slightly cooler summers when compared to the ranges that lie further to the east.

Yellowstone and Teton National Parks

The greater Yellowstone area consists of an elevated plateau which has an average height of about 2300 m (7,500 ft). Surrounding the plateau on the east and south are mountains that range in height from 3000 to 4000 m (10,000-13,000 ft). Yellowstone, being the oldest of the national parks, has a long history of meteorological observations, much of which is displayed in Figure 8.14. On average most of the Yellowstone/Teton plateau receives between 50 and 70 cm (20-30 in) of annual precipitation. Even though there are no climate stations in the mountains, limited data from snotel stations indicate that precipitation in the mountains is significantly higher than on the plateau. The time at which snow on the ground obtains its greatest seasonal depth is a function of elevation; in Yellowstone it occurs in late March-early April, and in the surrounding mountains generally between the middle and the end of April. Precipitation is spread fairly evenly over the year, there being a slight

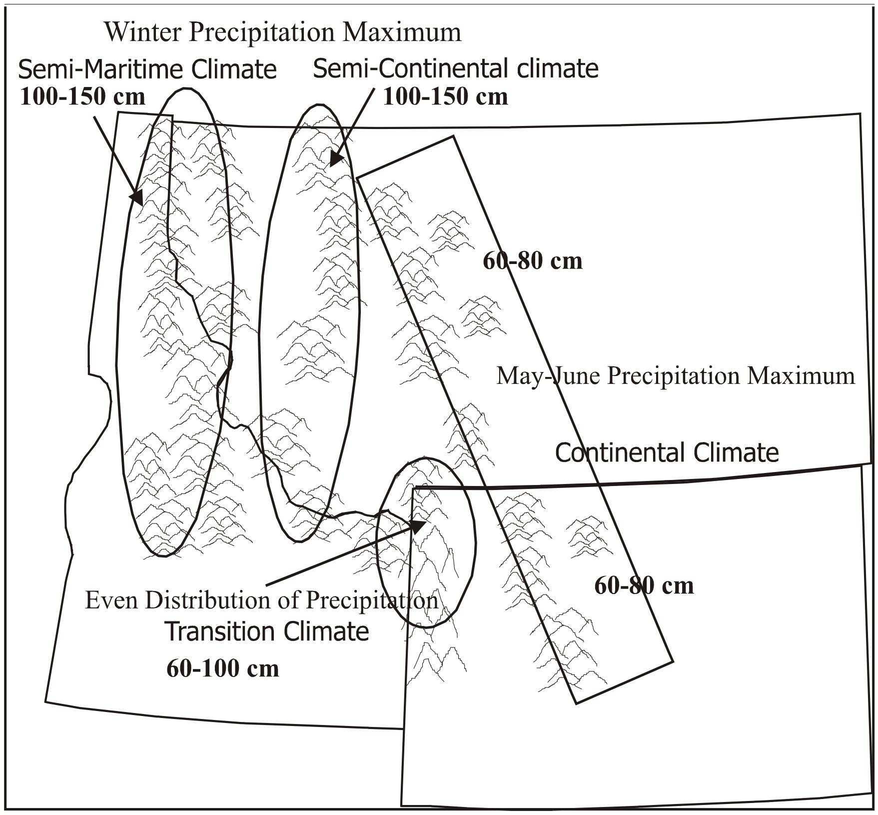


Figure 8.13- Climate regimes of the northern Rockies. Numerical values are estimates of annual precipitation at mid to high elevations within each region.

maxima in December-January and again in May-June. In Figure 8.14 notice how a station like Jackson, which is located in a valley, receives considerably less precipitation than more 'open' sites to the north. Of course the mountains around Jackson receive considerably more snow than what is measured in the valley itself.

Data from the climate station at Moose (1972 m), located on the east side of the Tetons, indicates that during midwinter, measurable precipitation occurs one out of every two days. This does not mean that it snows every other day. It may snow lightly for three for four days straight, and then not snow again for a week. From July-October, precipitation occurs about once every four days. During the summer however, we have to be careful interpreting rainfall data since convective rain showers have limited spatial coverage. What this means to the backcountry traveler is that it rains in the area much more frequently than rainfall data from one or two fixed climate stations indicates. Therefore you should anticipate the development of thunderstorms in the area on most summer afternoons. As a result of frequent thunderstorm activity, prudent climbers and hikers should be concerned about and monitor cloud-to-ground lightning during the late afternoon and evening hours. Climbers should also note that during the summer, overnight lows can be quite cool. The average minimum temperature for July and August at the Moose climate station is +5º C (41º F). This means that above 2500 m (8,200 ft), the overnight temperatures frequently dip below freezing. The coolest overnight temperatures will typically occur on cloudless nights when the winds are light.

Most people are aware that during the winter, Yellowstone is one of the coldest locations in the continental USA where daily observations are recorded. Table 8.1 lists monthly mean and extreme temperatures for the months of November through April from the climate station at Lake Yellowstone (2369 m or 7,770 ft).

It is readily apparent from this table that winters are cold in Yellowstone and that from time to time temperatures can become life threatening. If you are going to travel in this region in the winter, you should be equipped to deal with temperatures below -30º C (-22º F). Since the Yellowstone area is often referred to as the nations icebox, we compared average monthly temperature data from Lake Yellowstone with several other well know "cold spots" around the country. By comparison, Lake Yellowstone has slightly colder mean monthly temperatures than Fraser, Colorado (2610 m or 8,560 ft), although the mean monthly minimum temperatures at Fraser are a degree or two colder than Lake Yellowstone. We also compared Lake Yellowstone with International Falls, Minnesota. Lake Yellowstone is significantly warmer than International Falls, at least from December through February.

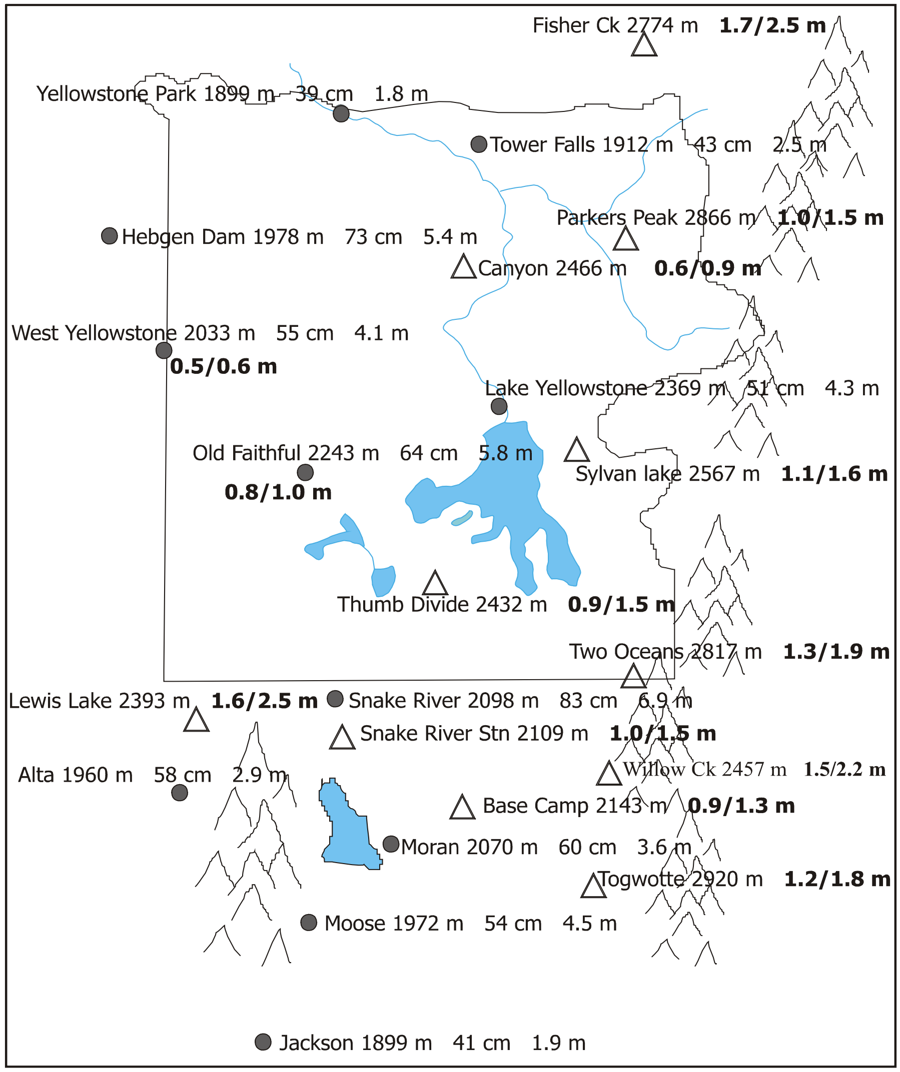


Figure 8.14- Climate data (solid circles) and snotel data (triangles) for stations in Yellowstone and Teton National Parks. Climate data consists of: station name, elevation (m) annual precipitation(cm), and annual snowfall. Snotel data consists of: station name, elevation, first of the month snow depth for February/April. There are several sites where a climate and snotel station are co-located.

In March, International Falls warms up considerably faster than Lake Yellowstone, and remains warmer throughout the summer. As a result, Lake Yellowstone has an annual temperature that is 2.7º C (4.9º F) cooler than International Falls. In order to see how winter temperatures in Yellowstone compared against higher elevation sites, We compared Lake Yellowstone to Climax, Colorado (3463 m). Monthly averaged winter temperatures indicate that the Yellowstone area is slightly colder than Climax, despite being 1090 m (3,600 ft) lower.

Table 8.1 Lake Yellowstone winter temperatures

	Ave. High	Ave. Low	Extreme High	Extreme Low
November	+1 °C	-11 °C	+17 °C	-34 °C
December	-4	-16	+9	-42
January	-6	-19	+7	-46
February	-2	-18	+13	-46
March	+1	-16	+14	-42
April	+6	-10	+18	-32

So why are winter temperatures in the Yellowstone area so cold? It is a combination of the elevation, near continuous snow cover, and most of all, the fact that the plateau is surrounded by mountains. These mountains essentially trap cold air over the plateau allowing deep temperature inversions to form. Even though Yellowstone is not considered your typical valley, during the winter cold air becomes stagnant like in any other valley. The coldest winter temperatures occur when the northern Rockies are under the influence of a surface high with a large amplitude 500 mb ridge positioned over western Canada.
Once cold air becomes trapped in Yellowstone, the only way it can be replaced by relatively warmer air is if a deep trough or low moves into the area from the west. In this situation cold continental air that has been sitting over the northern Rockies is exchanged with warmer maritime air from the West Coast.

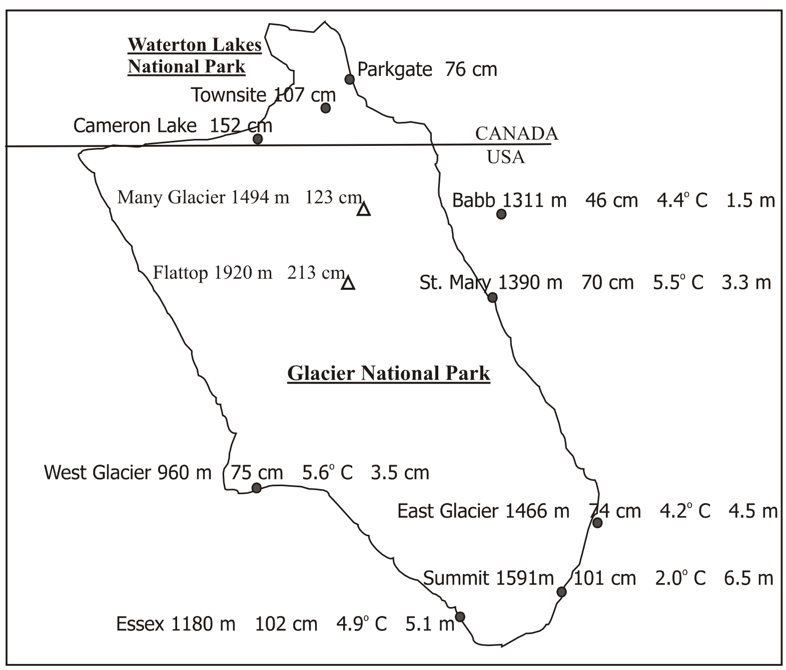


Figure 8.15- Climate (circles) and snotel data (triangles) for Glacier and Waterton Lakes National Parks. Climate data consists of: station, elevation (m), annual precipitation (cm), annual temperature (C), and annual snowfall (m). Snotel data consists of: station, elevation (m), and annual precipitation (cm).

<u>Glacier and Waterton Lakes National Parks</u>

These two parks incorporate some of the most rugged terrain in all of the northern Rockies. As was alluded to earlier, this area is one of the wetter spots in the northern Rockies. Annual precipitation at higher elevations is on the order of

150 cm (60 in) with the heaviest precipitation occurring during the winter. The precipitation gage at Flattop Mountain (1920 m) in Glacier NP for example, receives 49% of its annual precipitation between the months of November and February. Figure 8.15 shows climate and precipitation data for a few stations in and near Glacier NP. The relatively low annual precipitation at West Glacier (960 m) is due to its low elevation. The large contrast in precipitation between Flattop and Many Glacier is interesting. It probably reflects differences in elevation and more importantly, differences in the surrounding topography. Monthly averaged precipitation for these two sites indicates that the largest difference occurs in the winter. We would speculate to say that there is probably considerably more orographic lift occurring in the vicinity of Flattop than around Many Glacier.

Table 8.2 provides a sample of temperatures a hiker or climber could expect during the summer at intermediate elevations within Glacier NP. The proper way to interpret this data is as follows: The range of high temperatures during May is roughly between 19^o and 2^o C (66^o and 36^o F). What is readily apparent is that the range of high temperatures is much larger than the range of low temperatures. At this particular elevation the overnight lows are for the most part below freezing until mid-June. The warmest high temperatures occur when the axis of a high pressure ridge is located to the east of Glacier NP. This allows warm air from the Great Basin to move into western Montana. However, temperatures can be quite cool at night when winds are light and the sky is free of clouds. The coolest high temperatures occur when a synoptic-scale trough or low moves over the region. Keep in mind that temperatures in May and early June are dependant on the amount of snow on the ground. At higher elevations in Glacier NP, the first accumulating snowfall of the season typically occurs during the second half of September.

Table 8.2: Temperature range at Flattop Mountain (1920 m)

Month	High Temp. Range	Low Temp Range
May	19^o to 2^o C	3^o to -8^o C
June	20 to 3	5 to -3
July	25 to 9	10 to 1
August	27 to 8	10 to -1
September	24 to 3	8 to -5

Strong winds at higher elevations are a common occurrence in both Glacier and Waterton Lakes, especially during the winter when the polar jet lies overhead. The east side of Glacier NP as well as communities along Highway 89, frequently experience strong chinooks during the winter. Environment Canada (Canadian Weather Service) estimates that chinooks on average occur four to five times per month on the eastern side of Waterton Lakes. These chinooks cause rapid melting of the snowpack as well cause rapid rises in leeside temperatures. Chinooks are so efficient at melting snow because strong winds cause a continuous flow of dry air over the snowpack. In addition to chinooks, low-level winds are often accelerated as they flow through the narrow canyons of Glacier and Waterton Lakes. A good indicator of the frequency and strength of the winds at higher elevations are the numerous small cirque glaciers located on the eastern side of the Continental Divide. Strong winds blow snow from the windward slopes over the crest of the divide where it is deposited in the cirques and bowls of the eastern slopes.

Thunderstorms are common throughout the summer months, in large part due to the warming of the plains to the east.   Unlike the Rockies to the south, northern Idaho and northwest Montana have a relative high frequency of synoptic lows and troughs move over the region during the summer months. These disturbances often work in conjunction with convection to produce periods of moderate to heavy rain across much of the region.

Forecasting: Chinooks and downslope windstorms

Here are some suggestions to help you forecast chinooks. First of all recall that the conditions that lead to a downslope windstorm or chinook are: moderate to strong summit-level winds, the wind direction should be close to being perpendicular to the long-axis of the mountains, and finally the air in the middle troposphere has to be stable.

Step One- Choose one of the forecast models (ETA, AVN) from one of your favorite web sites. Pull-up the 700 mb wind speeds. The reason you want this level is because it is close to summit level for most of the larger mountain ranges in the western USA. As a rule of thumb, the winds at this level should be around 20 ms^{-1} (45 mph) or greater.

Step Two- Check on the 700 mb wind direction if you were not able to in step one. If 700 mb winds speeds (wind barbs or wind vectors), you can use the 700 mb height field as a default. Remember that at this height (3300 m), the direction of the wind is parallel to the height isolines. You will easily be able to tell the wind direction, which for the western USA should be from the northwest to southwest. Determining wind speeds from the height field is more difficult. The closer isoline spacing the stronger the winds, but estimating actually speeds takes some practice.

Step Three- Now you need to determine if the middle troposphere is stable or not. Most likely you will not be able to access a vertical cross-section of surface-to-500 mb temperatures. If you can get a forecast or current sounding from a nearby station, use it. You want to see if the lapse rate between 850 mb and 500 mb is less than about 6º C km^{-1}. In lieu of a sounding or vertical cross-section, pull-up the 850 mb temperatures, noting what the temperatures are in the area of interest. Repeat this for the 500 mb level. Since the average height difference between 850 and 500 mb is about 3800 m (± 200 m), if the temperature difference is less than 23º C (41º F), then the layer is stable.

This procedure may be difficult to do depending on the availability of model data. One suggestion is to wait for a chinook to be forecasted in an area of interest to you, then try this procedure to see what key weather elements the forecasters are looking at.

<u>Banff and Jasper National Parks</u>

The weather and climate of the northern tip of the Rockies is quite similar to what was just described for Idaho and Montana. The bulk of moisture that eventually falls as snow and rain in is transported from the Pacific Ocean. The largest winter storms of the season occur when a trough or low forms over or north of Vancouver Island,. This produces strong southwest winds over British Colombia. Since there is considerable mountainous terrain between Banff-Jasper and the Cascades, the wettest region in this part of the Rockies lies to the west (upstream) of the Continental Divide. This includes the Bugaboo's, Glacier National Park, the Selkirk and Columbia mountains; where annual precipitation is on the order of 125-150 cm (50-60 in). In Banff and Jasper National Parks, annual precipitation is on the order of 100-125 cm (40-50 in). In contrast the climate stations in the towns of Banff and Jasper, which are located to the lee (east) of the Continental Divide, only average about 40 cm (16 in) of annual precipitation.

Most of the climate stations in eastern British Colombia have a winter and summer maximum in precipitation. Winter precipitation is heaviest from Nov-Jan, this is followed by a two-fold decrease in monthly precipitation in February and April. In May, precipitation starts to increase, with the second maximum occurring in July-August. The February-April precipitation minima is due to the fact that the polar jet lies for the most part lies well to the south of the region. Due to the decrease in precipitation during the second-half of winter, the snowpack at the mid-elevation sites reaches its maximum depth usually sometime in February. To the east of the Continental Divide, there is a single precipitation maximum that corresponds to the summer convective season. Thunderstorms are a common throughout the region. Many of the larger cumulonimbus clouds develop over the Rockies then move eastward onto the plains of Alberta during the evening.

In the elevation range of the climate stations in the region (500-1000 m), mid-winter temperatures range from -5^o C to -15^o C (23^o to 5^o F), obviously it can get much colder than this when the region is under the influence of high pressure and the skies are free of clouds. During the summer temperatures at this same elevation range from 10^o C to about 25^o C (50^o to 77^o F), which means that at an elevation of 3000 m (9,800 ft) expect temperatures on the order of -4^o C to 10^o C (27^o to 50^o F).

One noteworthy feature of this stretch of the northern Rockies are the size of some of the glaciers in Jasper National Park, especially when compared to the Rockies of Montana and Wyoming. The annual precipitation in Jasper N.P. is roughly equivalent to what occurs in northern Idaho and Montana, however, the annual temperature in Jasper is considerably cooler. This brings up an important point, many mountain glaciers were much larger in preceding centuries. What we see today are much smaller remnants which continue to slowly melt and recede.

To the east of the Continental Divide, chinooks are very common during the cooler months of the year. These strong winds cause a very large rise in temperatures at the base of the Rockies, along with a rapid snowmelt.

Northern Rockies Weather Summary

* Northern Idaho and northwest Montana are considerably wetter than the rest of the range. This is a result of a much higher frequency of synoptic disturbances moving over the northern region during the summer and winter.
* The eastern ranges have a May-June precipitation maximum, northwest Wyoming and eastern Idaho have a fairly even distribution of precipitation throughout the year, while the north has a pronounced winter maximum.
* Downslope windstorms and chinooks are common in winter at the base of the eastern Rockies.
* Summer-time Convective rainfall and cloud-to-ground lightning are common over much of the region.

* Winter temperatures in some of the higher elevated valleys and uplands, such as Yellowstone, can periodically be extremely cold.

WEB National Weather Service

Billings	**www.crh.noaa.gov/billings**
Missoula	**www.wrh.noaa.gov/missoula**
Great Falls	**www.wrh.noaa.gov/greatfalls**
Pocatello	**www.wrh.noaa.gov/pocatello**
Riverton	**www.crh.noaa.gov/riw**
Spokane	**www.wrh.noaa.gov/spokane**
Boise	**www.wrh.noaa.gov/boise**

Forest Service National Avalanche Center - **www.avalanche.org/~nac**

Bozeman Avalanche Center- **www.mtavalanche.com**

Glacier Avalanche Center- **www.glacieravalanche.org**

Jackson Avalanche Center- **www.jhavalanche.org**

Canadian Avalanche Association- **www.avalanche.ca**

Canadian Weather- **www.weatheroffic.com** or **www.cmc.ec.gc.ca**

Professional Profile
An Interview With Nolan Doesken

Nolan Doesken is the assistant state climatologist for Colorado. He has been involved in collecting and analyzing climate data in Colorado for over 20 years. Nolan has had a fascination with snow since his youth. In 1996 he co-authored a book on the fundamentals of snow meteorology and snow measurement techniques. Check out the Colorado Climate Center at: http://climate.atmos.colostate.edu

Q1- Why are the central Rocky Mountains renown for dry powder snow?
ND- Basically it is a combination of our high elevation and interior continental location. A lot of snow falls at temperatures well below the freezing point and from airmasses that have been stripped of low-level moisture from upstream mountain ranges. The result is that these clouds are somewhat lacking in supercooled water, so ice crystals are not so heavily rimed as is common in warmer maritime areas. The result is snow that is poor for making snowballs, but great for skiing and snowboarding.

Q2- Is there very much seasonal snowfall variability across Colorado?
ND- We have known for many years that seasonal snowfall patterns often favor either the northern Rockies or the southern Rockies, but rarely both. It is however a fairly recent discovery that these changing patterns are often associated with El Nino and the Southern Oscillation and the resulting changes in ocean currents and sea surface temperatures thousands of miles away. Here in Colorado

we often see our northern mountains (Winter Park, Vail, Steamboat, etc.) have excellent snow years while the southern mountains (Wolf Creek, Purgatory, etc.) are getting lousy snows. When our southern mountains are getting buried, northern Colorado snowpacks are often thin. There are winters like 1977, when the entire region is dry, and there are likewise occasional winters when the whole state is buried in deep snow. But as a rule, one regions bounty is another's woe.

Q3- What are some of the difficulties in measuring snowfall amounts?
ND- Measuring snow seems so simple to the casual onlooker- just take a ruler and stick it in the snow. That's fine and good until you begin to comparing data from different location and different observers. The reality is that measuring snow in a consistent and accurate fashion is much harder than people realize. The reasons are pretty obvious, when you think about it. Measuring snow is like shooting at a moving target. Snow melts, snow settles over time (densification), and of course it blows and drifts. Three good observers can measure the same snowfall and come up with distinctly different answers. We have found that the more frequently an observer takes a snow measurement, the <u>more</u> snow they are likely to report.

The Mountains of New England

The weather in the mountains of New England like most of the rest of North America is dictated primarily by mid-level westerly winds. This should not be taken to mean that the winds are always out of the west, however what it does mean is that the long-term average wind direction in the middle-troposphere is from the west. In actuality there are a number of important flow patterns (primarily winter time) that you should be able to recognize in Figure 8.16. 1) northwest or northerly flow from Canada, 2) southwest or southerly flow from the Ohio valley or mid-Atlantic Seaboard, 3) easterly or southeasterly flow from the Atlantic Ocean, and; 4) east to northeast flow from a Nor'easter. These Nor'easter's are not very common, however they produce large amounts of rain or snow. Many of these storms are remnants of tropical storms that have re-intensified as they move over the warm water of the Gulf Stream. They typically develop very rapidly and hence are called cyclogenic 'bombs' by forecasters.

Remember that cooler air moves into New England when the flow originates from Canada. When the dominate flow direction is from the southwest (trough), the air mass over New England will be warmer and somewhat wetter when compared to north or northwesterly flow (ridge). Storms that move in from the Atlantic are wet and start-off warm but temperatures may cool as the flow becomes more north or northeasterly.

<u>Adirondacks</u>

Since the Adirondacks cover a large portion of northern New York, there is considerable variation in the distribution of rain and snow. For example, annual precipitation varies from 130 cm (51 in) over the southwest and central Adirondacks to about 100 cm (39 in) in the vicinity of Lake Placid. Annual snow fall is distributed in a similar fashion, the southwest and central mountains typically receive from 380-460 cm (150-180 in), while the eastern slopes receive on the order of 250-305 cm (100-120 in). There is a slight increase in monthly precipitation from June through

September, but overall precipitation is remarkably uniform over the course of a year. In January for example, snowfall occurs on 12-15 days out of the month. In July, expect measurable rain to occur 8-

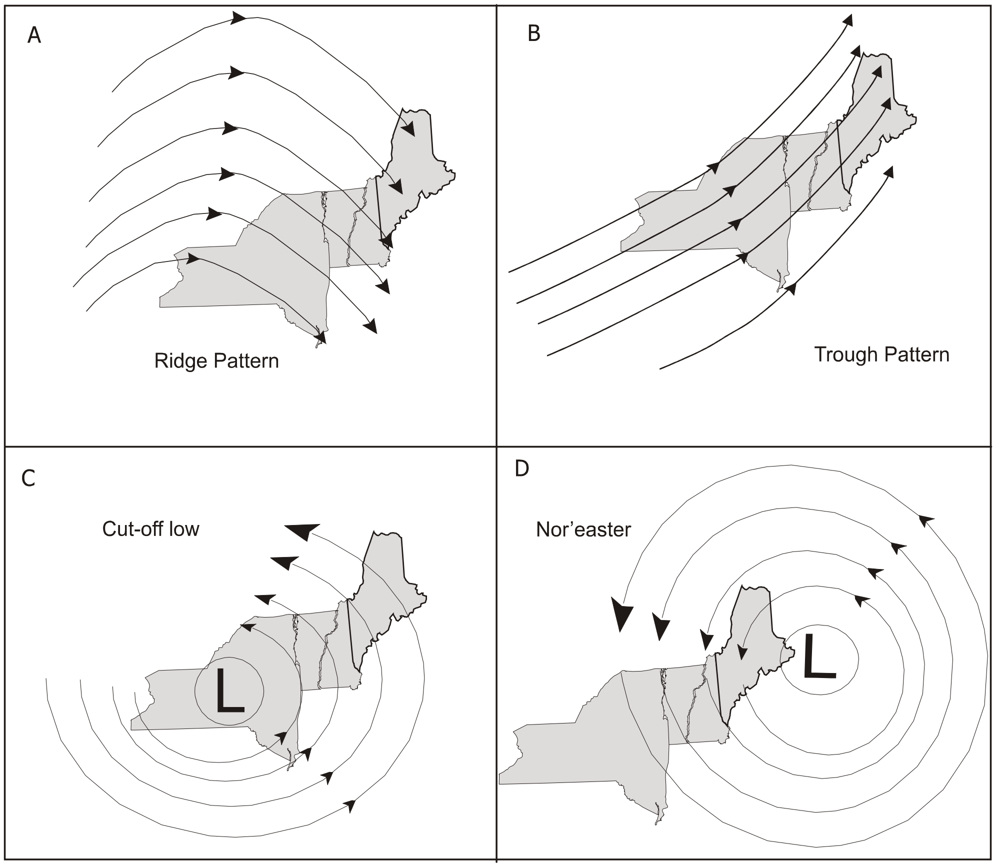


Figure 8.16- Four fairly common flow regimes of New England.

12 days out of the month. During the summer a considerable amount of rain falls during thunderstorm activity.

Temperatures across this region are a function of elevation, two weather stations on opposite sides of the range but at the same elevation will have similar monthly and seasonal temperatures. On a daily basis however, there can be noticeable temperature differences across the range as a result of cold or warm air masses moving into or out of the region. The majority of the climate stations in this region are located at an elevation of about 500 m (1,700 ft). At this elevation typical January temperatures range from -1º to -15º C (30º to 5º F). In addition, the number of days in January where the minimum temperature is below -18º C (< 0º F), is on the order of 12 to15. In July expect a temperature range of 10º to 26º C (50º to 80º F), fortunately very few days get hotter than 32º C (90º F). At higher elevations allow for cooler temperatures than what is listed above, especially in the summer.

<u>White Mountains</u>

The White Mountains of New Hampshire are in many respects the centerpiece of New England mountains, in large part due to the fact that there are eight mountains that exceed a mile in height (1600 m). Since the early 1930's a non-profit weather observatory has been located atop

Mount Washington (1915 m or 6,280 ft). Through the years the Mt. Washington Observatory (MWO) has gained a reputation as a credible organization who's mission is to document the harsh climate of the Presidential Range, and to disseminate this information to the public.

An examination of the climate record of northern New Hampshire clearly shows the influence of elevation on annual precipitation, as displayed in Table 8.3. Like all of New England precipitation is distributed quite evenly over the 12 months of the year, although at the MWO November and December receive the highest precipitation. Average annual snowfall is 645 cm (254 in) at the observatory but somewhere around 200 cm (80 in) in the surrounding low-lands, away from the Presidential Range. At MWO the December through March monthly average snowfall is about 104 cm (41 in). Like any other mountainous area, there is considerable year-to-year variation in snowfall in the Presidential Range. In addition, the actual amount of snow that lies on the ground at any given time not only depends on the amount that fell from the sky, but how much of the snowpack has been re-distributed by the wind after it fell. This is apparent to anyone who has seen or been to the top of Mt. Washington during the winter. Snow accumulates in depressions in the landscape and in the forest much better than it does in wind exposed areas.

Table 8.3: Climate data for northern New Hampshire

Station	Elevation (m)	Annual Precip. (cm)	Ave. Jan Temp. (C)	Ave July Temp (C)
Benton	366	93	-8°	19°
Berlin	283	107	-10°	19°
Errol	390	98	NA	NA
1St Conn. Lk.	506	113	-13°	17°
Jefferson	376	122	NA	NA
Lancaster	262	93	-10°	19°
Mt. Washington	1915	251	-16°	9°
North Conway	162	122	-8°	20°
Pinkham Notch	612	150 est.	-10°	17°

Temperatures in the White Mountains are primarily a function of terrain configuration (valley versus ridge) and elevation. In Chapter 3 you learned that temperatures in general decrease with increasing height, the main exception occurs in the presence of a temperature inversion. When a inversion does occur, the rule of thumb which states that air temperature decreases with height is temporarily invalid. Since temperature inversions occur frequently during the winter months throughout the mountainous terrain of New Hampshire (this applies to all of the mountainous terrain of

New England), they are an important factor in the wintertime climate of the region. What all this means to the mountain traveler is that during the winter the air temperature at higher elevations is not as cold as you may think it is based on the current temperature at the base of the mountains (not accounting for windchill). Notice in Table 8.3 how the average January temperature for Mt. Washington is not that much colder than temperatures at surrounding low-land stations. Also notice that during the summer the temperature difference is significantly larger. Overall, there will be a larger difference in air temperature between the base of the mountain and the summit during summer rather than during the winter.

Without a doubt the most important meteorological element in the White Mountains is the wind. Above treeline the seasonally averaged wind direction is northwest-to-southwest. This does not mean that significant winds do not blow from other directions as well, however when they do occur they are typically of short duration. The windiest months are October through April where monthly average wind speeds run about 33% higher than during the summer. Keep in mind however that strong or very strong winds can occur on any given day of the year. In a review of high winds that occurred on Mt. Washington between 1980 and 1997, Keim and Edminster (1998) found the 18 year average of wind gusts greater than 56 m/s (125 mph), was 8.7 per year, while the average for gusts greater than 63 m/s (140 mph) was 2.2 per year. We want to emphasize that these are gusts of wind, in other words- short burst (a few seconds) of very strong winds. There is a big difference between sustained winds, which are averaged over several minutes, and gusts. When the sustained winds are high, there is a greater chance that strong gusts will occur as well.

The obvious question that has to be addressed is: why is Mt. Washington so windy, and is it unique to the region? There are several considerations: first, the mountains of New Hampshire and northwestern Maine separate two contrasting climatic zones, the maritime zone to the east and the continental zone to the west and north. During the winter months the continental zone is much colder than the maritime, this produces higher pressure over northern New England and Canada. Since air moves from higher pressure to lower pressure, the lower tropospheric winds are from the west or northwest across much of the region. Secondly, the polar jet stream is frequently located over New England, as illustrated in Figure 8.16. The strongest winds in the polar jet lie well above the tops of the mountains of New England, but from time to time very strong winds can be found in the middle and even lower troposphere.

On December 4, 1980 a 81 m/s (178 mph) wind gust was recorded at the observatory atop Mt. Washington. The 850 mb height field and sustained wind speeds for this event are displayed in Figure 8.17. Notice that the flow is north-to-northwest over New England at this level (850 mb is close to the height of the mountains in this region). Prior to and at the time of this event, very cold air was located in Quebec with much warmer air over coastal New England. This helped the intensify the low that is positioned over Nova Scotia. Free atmospheric winds at 850 mb were on the order of 30-35 m/s (65-80 mph). Although not shown, the 500 mb low center was positioned near Long Island, while the axis of the polar jet stream was located along a line from Buffalo to Philadelphia. Not only were the winds at Mt. Washington very gusty, but the average daily wind speed for December 4th , was a phenomenal 53 m/s (117 mph). Strong winds continued into December 5th, but not with the same intensity.

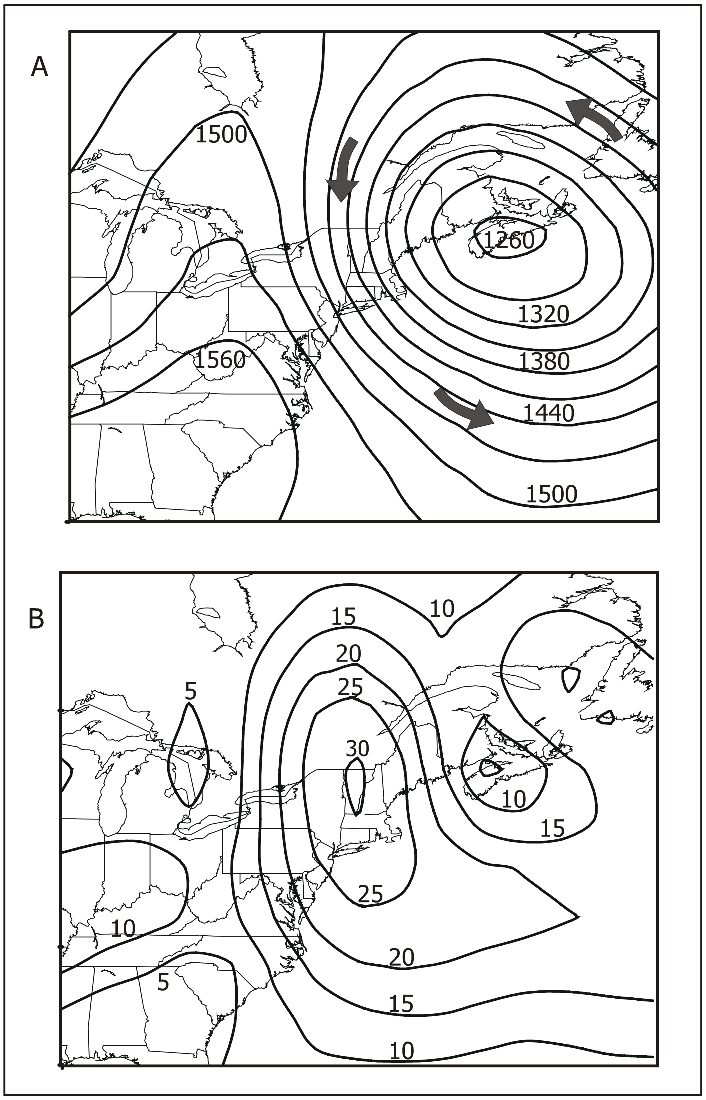
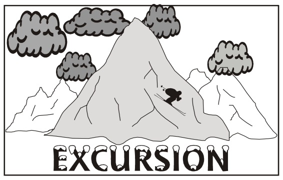


Figure 8.17- December 4, 1980 windstorm. A) 850 mb heights (m), B) sustained wind speeds (m/s). A peak wind gust of 81 m/s (178 mph) was measured at the Mt. Washington observatory during this storm.

In retrospect, during the night of December 3, a strong temperature inversion centered around 800 mb (1900 m or 6,200 ft) developed over New England. This height corresponds closely with the elevation of the highest peaks in the White Mountains. What appears to happen is that strong free atmospheric winds, which are generated by the synoptic-scale pressure gradient, are amplified as they flow between the ground and the inversion. How such strong gusts are generated we do not believe anyone fully understands at this time. What we do know is that most of the big wind events that have been recorded by the MWO, have a duration on the order of 12-24 hours, although longer events do occur. Winter travelers to the higher terrain of the White Mountains should note that strong wind speeds ‘ramp-up’ over a period of 3-6 hours. In other words, you will not get blasted with a 50 m/s (110 mph) gust during a period when the winds are otherwise light. As wind speeds increase however, the likelihood that very strong gusts will materialize also increases.

It is interesting to speculate whether the high winds frequently observed on Mt. Washington are common on other high mountains of New England as well. There is not much in the way of data to support a definitive answer to this question, nevertheless we would speculate that most of the higher peaks of Maine, New Hampshire, and Vermont also have very strong winds. Winds at higher elevations in the Adirondacks on the other hand are probably not as strong or frequent as they are in the mountains to the east, simply because of their more continental location. One important aspect that is extremely difficult to account for is: what role does local topographic affects play in Mt. Washington wind events. This could only be determined if additional wind sensors (anemometers) were placed on a number of higher peaks in the region. This would also make a good high resolution computer model research project.

Acid Rain



In the mid and late 1980's the topic of acid rain was one of the most popular environmental issues in the media. By the mid 1990's however there was little mention of it. So what is acid rain and has the problem gone away? First, acid rain is defined as natural occurring rain or snow that has a pH <5.6. This occurs when dust and organic cloud condensation nuclei are replaced by sulfur dioxide (SO_2) and nitrous oxides (NO_x) . SO_2 and NO_x are commonly emitted into the atmosphere during the burning of coal and other fossil fuels. It should be no surprise than that the northeastern USA was the region with the greatest acid rain problem. The area of concern was eastern Ohio, northern West Virginia, Pennsylvania, large parts of New York including the Adirondacks, as well as parts of Vermont. There was considerable media interest in the fact that so-called "pristine" lakes of the Adirondacks had very high acid levels. This illustrates the fact that atmospheric pollutants area transported considerable distances from their place of origin.

Secondly, the acid rain problem has not disappeared. However, due to stricter regulations on coal during industries since the late 1980's, the amount of SO_2 and NO_x released in the atmosphere each year has been cut considerably. This in turn has lead to a decrease in the amount of acid entering lakes and waterways of the northeastern USA, although the pH in these bodies of water remains low. The environmental impacts of acid rain (and snow) are: the death of certain kinds of fish species, it also destroys certain kinds of bacteria and microorganisms in soils, and it has been know to weaken or kill some species of spruce trees. Sulfates are also a major producer of haze, which causes lower visibility in mountainous regions, even though the sulfates originate hundreds of kilometers away from the mountains.

New England Summary

* Even distribution of precipitation throughout the year.
* Considerably higher amounts of precipitation in the mountains than at their bases.
* Considerable winter fluctuations in air temperatures in response to changing air masses. The coldest air arrives with flow from the north and northwest while the warmest air is transported by southerly or southwesterly flow.
* Warmest temperatures during the summer occur when a quasi-stationary (slow movement) ridge of high pressure is located over over the Great Lakes or New England.
* Thunderstorms primarily occur in June, July and August.
* Strong winds are a common occurrence at higher elevations especially during the winter months.
* Occasionally powerful storms develop off the coast of New England, producing strong easterly winds and transporting large amounts of moisture into the region. These are not good times to be in the mountains.

WEB Mt Washington Observatory- **www.mountwashington.org**

Avalanche Center- **www.tuckerman.org**

National Weather Service

Albany- **www.erh.noaa.gov/er/aly**

Burlington- **www.erh.noaa.gov/er/btv**

Portland- **www.erh.noaa.gov/er/gyx**

Caribou- **www.erh.noaa.gov/er/car**

9

REGIONAL WEATHER SURVEY, PART III
Himalaya, Karakoram, Andes, and Alps

Chapter Highlights:

- ✓ Season-by-season description of weather in the Himalaya and Karakorum
- ✓ The Monsoon revealed.
- ✓ Weather and Climate south of the border- The Andes
- ✓ Mountain Meteorology, where it all started- The Alps

In this chapter we will explore the weather and climate of the worlds highest mountains. The primary hindrance to such an endeavor is the fact that there is not very much weather data from these regions; past or present. Despite these limitations you will find this overview helpful in not only understanding the weather and climate of these regions, but in planning trips as well.

Himalaya and Karakoram

The Himalaya ('abode of snow') and the Karakoram ('black gravel') mountain ranges are adjacent to each other, frequently however the weather is radically different from one range to the other. On the large-scale (synoptic) the primary control on weather in these mountains are latitude and distance from the Indian Ocean. On the local-scale, weather is controlled by elevation and the height of the surrounding terrain. The most important topographic feature in all of Asia is the Tibetan Plateau. Without this enormous high elevation land mass, the weather and climate of Indian Subcontinent, Central Asia, and even Southeast Asia would be radically different than it currently is.

The Himalaya stretch from northern Burma (Myammar) and Arunachal Pardesh, across the northern fringes of the Indian plain, finally terminating in northeast Pakistan (Figure 9.1). This arc of mountains is about 2400 km (1,500 mi) in length and forms the southern edge of the Tibetan Plateau. The Karakoram lie in northern Pakistan, and form a transition zone between the Indian Subcontinent to the south and Central Asia to the north. Along with the Pamir Mountains which lies due north, the Karakoram form the western edge of the Tibetan Plateau.

The 'dividing line' that separates the Karakoram from the Himalaya is quite nebulous. The general consensus is that the mountains which lie to the south of the Indus River are in the Himalaya, while those located to the north of the river are in the Karakoram. This makes Nanga Parbat (8120 m or 26,620 ft) which lies about 15 km (9 mi) south of the Indus River a part of the Himalaya, while the mountains 100-200 km (60-120 mi) due east are considered part of the Karakoram.

Both of these ranges are of course further divided into smaller sub-ranges. The Himalaya for example, are often broken down into five ranges based on height and proximity to the Indian Plain. We won't be using this classification very much, nevertheless you should be aware that it exists. Note that the Karakoram lie further north (34-38º) than the eastern Himalaya (27º). K2 for example lies about 8º of latitude further north than Mt. Everest, with a straight-line distance of 1300 km (800 mi) separating the two mountains. In addition, the eastern Himalaya lie within 700 km (430 mi) of the Bay of Bengal, which means that they are considerably closer to a large source of moisture than the Karakoram. This results in the eastern Himalaya being considerably wetter than the Karakoram. It is an interesting fact however, that the glaciers in the Karakoram and extreme northern Himalaya, are much bigger than anything found in the eastern Himalaya. Obviously temperatures play as major a role in glacier meteorology as does snowfall.

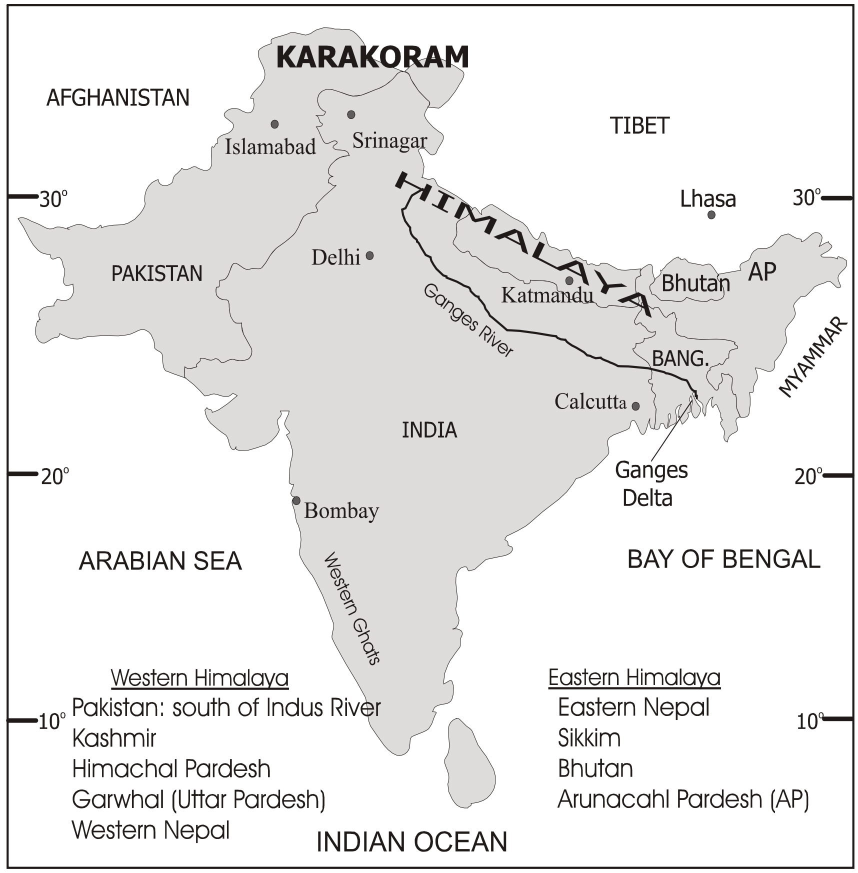


Figure 9.1- Geography of the Indian Subcontinent.

The weather over the entire Indian Subcontinent, of which the Himalaya and Karakoram form the northern border, can be broken down into two basic regimes: the winter and summer monsoons. During the winter monsoon the low-level winds are out of the north, while in summer they are out of the south-to-southwest. Travelers from outside the region know that the "best time" to trek or climb in this part of the world is May or October, and they vaguely know that it has something to do with the summer monsoon. In this chapter the discussion will focus on: how the summer monsoon works, precipitation patterns across the region, location of the predominate storm tracks, and why May and October are the optimum months to travel in the Himalaya, and why summer is the best time to travel in the Karakoram.

Monsoons

The term 'monsoon' is derived from an Arabic word which means- <u>season</u>. In the 10th-13th centuries, Arab traders and merchants sailed up and down the east coast of Africa and around the Arabian Peninsula. They quickly learned that the winds in the western Indian Ocean tend to blow *predominately from one direction for about six months, after which they would reverse direction for the remaining six months. They termed this seasonal shift in winds as the 'monsoon'. The importance was that if you wanted to sail to India and conduct business, you had to do it during the southwest monsoon (summer). When it came time to return to Africa you had to wait for the northeast monsoon (winter).*

During the 19th century, as the science of meteorology began to emerge, a number of meteorologists in India noticed the connection between the monsoon winds and precipitation. Since those early days, considerable research has been aimed at not only gaining a better understanding of how the monsoon works, but being able to predict the amount of rain it produces as well. There are other regions of the world that experience monsoonal weather, chiefly Southeast Asia. As noted in Chapter 8, Arizona, New Mexico, and Colorado experience a weak (in comparison to India) monsoon during the second-half of the summer as well.

Winter

During the winter months high pressure forms over Siberia due to that region's extreme continentality and high latitude. With high pressure over Siberia and low pressure over the Indian Ocean, low-level air (0-3 km above ground level) tends to flow from north-to-south over Central Asia and the Indian Subcontinent. In the middle and upper troposphere the winds blow out of the west, as both the polar and sub-tropical jet streams are very strong. There are times when both of these jets merge over Central Asia, and there are other times when they are distinct. In this part of the world the sub-tropical jet (STJ) is typically found between 20-35ºN, that is it lies over the southern edge of the Tibetan Plateau, never far from the Himalaya. During the winter months the 4500 m (14,700 ft) high Tibetan Plateau is extremely cold; in fact due to longwave radiational cooling, it is considerably colder than the air in the free atmosphere at the same elevation. This cold source creates a large temperature contrast (gradient) between the air over Tibet and the air over the Indian Ocean. This helps intensify the STJ during the winter. The core of strongest winds in the STJ is typically around 200 mb (12 km winter and 13.5 km summer), in which the average wind speed is about 35-40 m/s (75-90 mph), however winds in excess of 80 m/s (175 mph) are not uncommon (Figure 9.2). Below the jet core, wind speeds decrease considerably. At an elevation of 8-9 km (5-5.6 mi), the average free atmospheric wind speed is roughly 50% of that found in the jet core. This means that if your climbing at 7000 m (23,000 ft) in the Himalaya and a 60 m/s (130 mph) STJ lies over head, you should expect wind speeds (ignoring topographic affects) on the order of 25-30 m/s (55-60 mph).

The polar jet stream almost always remains north of 30-35º N, so it generally only affects the western Himalaya, however it does have a significant impact on the weather in the Karakoram, Pamirs, and Tien Shan ranges. Troughs and lows that developed along the polar jet transport moisture into this region from the Mediterranean and Arabian Seas. These disturbances often produce light to moderate widespread precipitation across the Karakoram and western Himalaya, but generally have little influence over the eastern Himalaya. As these disturbances move through the region they typically produce cooler temperatures aloft, as cold air from Iran and Afghanistan is transported southeastward. Troughs and lows also form along the sub-tropical jet stream. These disturbances produce widespread snow in both the western and eastern Himalaya.

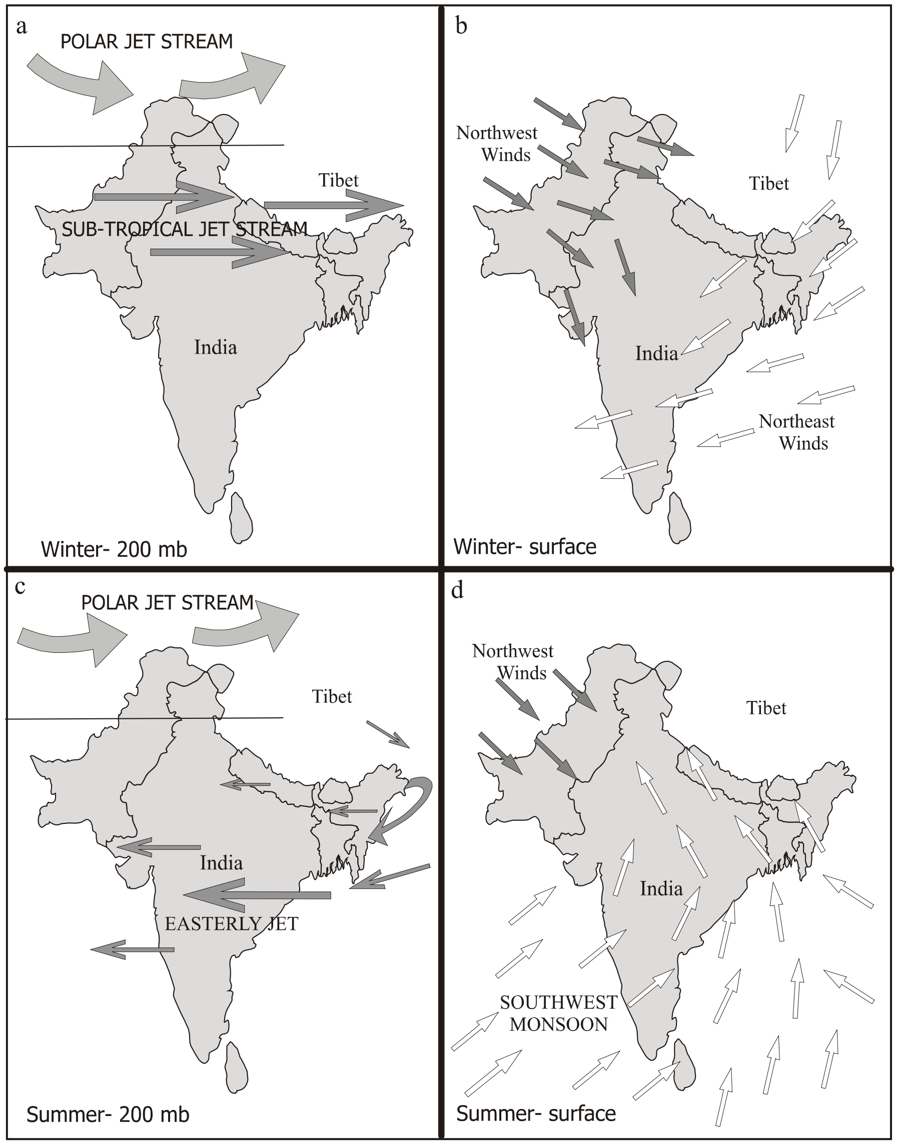


Figure 9.2- Winter and summer wind flow patterns over the Indian Subcontinent, at 200 mb and the surface.

Climbing and trekking is not out of the question during early winter in the eastern Himalaya. The primary factors to consider are cold temperatures and strong winds at higher elevations. The amount of snow that falls varies considerably from year-to-year. In years with large amounts of winter snow, many of the higher passes will be closed due to a deep snowpack. Climbing in the Indian Himalaya and Karakoram is pretty much a non-option in the winter due to moderate to heavy snow in the mid-elevation ranges as well as very cold temperatures. Trekking at lower elevations in the Karakoram is an option, however you should be prepared for cool temperatures, and considerable amounts of snow above 3500 m (11,500 ft).

Spring

Starting sometime in late March, in response to the increase in solar radiation and the subsequent development of low pressure over the Bay of Bengal, there is a dramatic increase in precipitation over the eastern Himalaya. This increase in precipitation is only a precursor to the heavy rains that accompany the summer monsoon. In addition, convective activity and thunderstorm frequency over the northern Indian plain increases between March and May. Several studies have shown that frequency of thunderstorms is at its highest when the STJ lies directly over the plains of northern India.

One aspect of the March and April weather that climbers in particular should note, and as seen in Figure 9.3, is that the STJ tends to weaken considerably during these months. This means that there will be days when the winds at higher elevations will be light-to-moderate, followed by days when the winds re-intensify. It is important not to remember that just because the upper-level winds have been light for several days, does not mean that they might not strengthen as a jet streak moves over the region. The data used to construct Figure 9.3 is from the Lhasa upper air data set. It is the closest long-term record available near the eastern Himalaya. Each data point is a weekly average, which means that the wind speeds experienced by climbers on any given day can be considerably higher or lower. The most obvious information that this figure conveys is the seasonal trend and the fact that wind speeds increase considerably with increasing elevation.

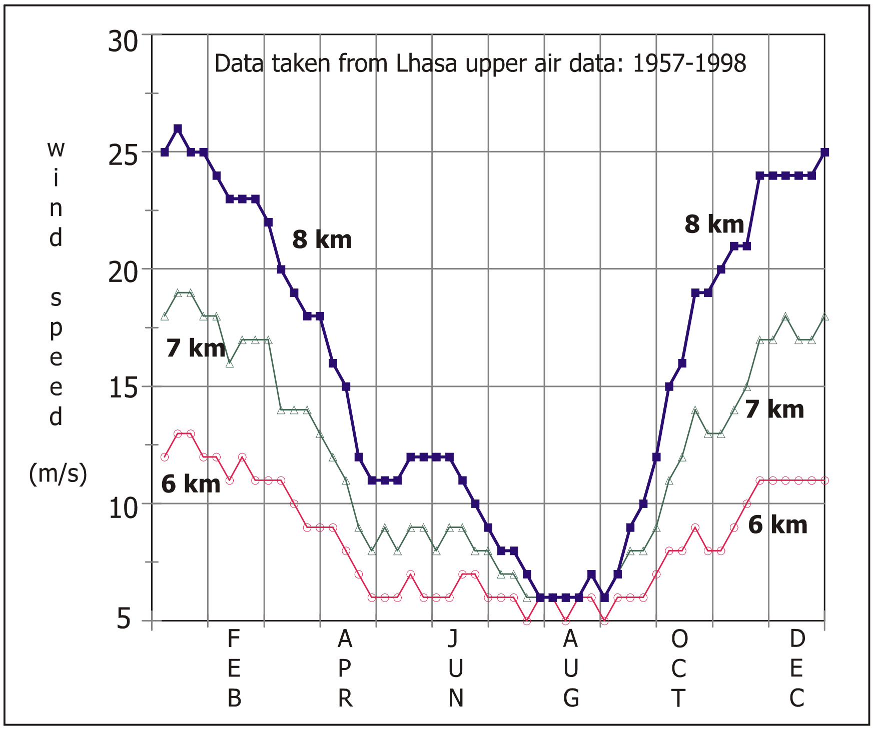


Figure 9.3- Weekly-averaged wind speeds for eastern Himalaya.

By the end of April or early May, as a direct result of the weakening of the Hadley Cell circulation system, the STJ virtually disappears for the summer. This marks the beginning of the pre-monsoon season in the Himalaya. In Table 9.1 May wind speeds observed at the 8 km (26,200 ft) level, are sub-divided into speed classes (taken from Lhasa upper air data). Climbers take note that speeds greater than or equal to 15 m/s (33 mph) have a frequency of occurrence of about 22%. The pre-monsoon season is more of a transition period between a relatively dry winter and a very wet summer. This means there can be a string of cloud free days, followed by several days that are mostly cloud with precipitation a possibility. By way of example, one of the authors and his wife spent 10 days in mid-May one year, hiking in the Garhwal Himalaya of northern India (the very northern part of the state of Uttar Pardesh). We experienced several days of rainy weather, followed by 3 or 4 days with clear blue skies, which was subsequently followed by another period of cooler rainy weather. These storms were more than afternoon thunderstorms, they were disturbances in the upper level westerlies

Table 9.1 Frequency of wind speeds during the month of May in the eastern Himalaya at 8 km.

speed (ms^{-1})	Frequency (%)
0-5	8.6
6-10	30.9
11-15	38.4
16-20	16.5
21-25	4.6
>25	1.0

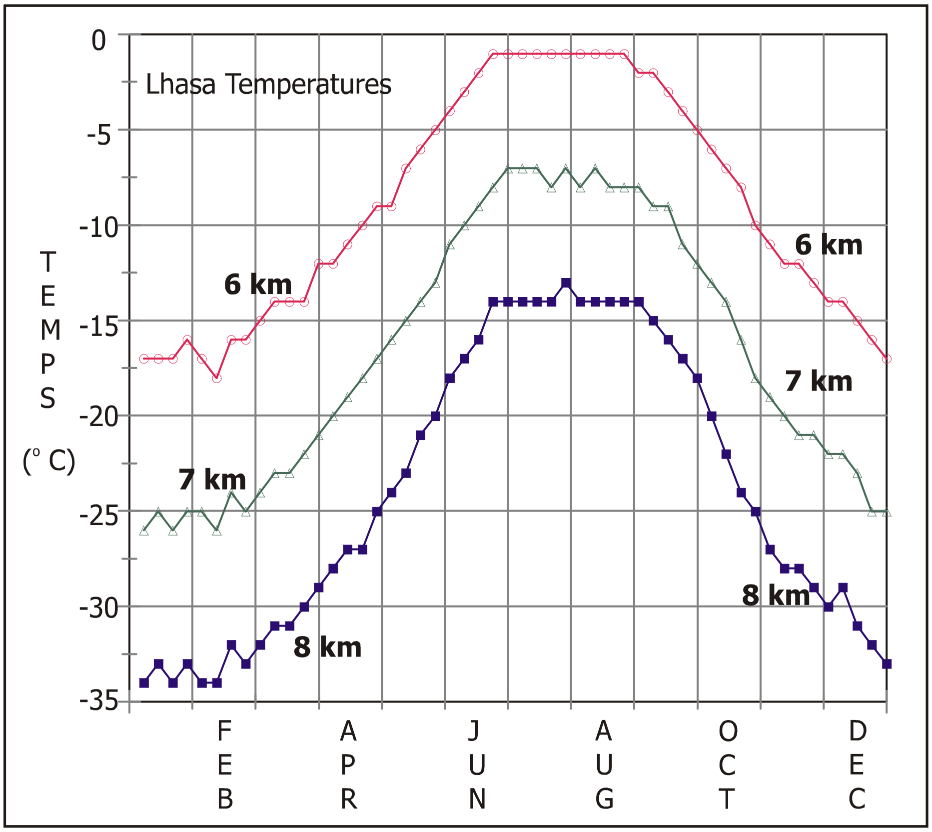


Figure 9.4- Weekly-average air temperatures over the eastern Himalaya.

that had a considerable amount of embedded convection.

By April temperatures over the Himalaya start to warm-up dramatically. By mid-May the freezing level in the free atmosphere over the eastern Himalaya can rise to 4000-5000 m (13,100-16,400 ft) at times (Figure 9.4). May is considered the hottest month in the northern Subcontinent because of the increase in surface heating as the sun moves higher in the sky, and because compared to June-September, cloud cover is limited. This period corresponds to significant snow melt in the Himalaya, which causes the rivers and streams to rise substantially. In addition, the increase in shortwave radiation and warmer temperatures causes an increase in the number of avalanches. During the late spring, a considerable amount of dust and dirt is carried into the lower troposphere as a result of increased surface winds. These winds are produced by strong up-drafts and down-drafts that develop as a result of vigorous convection. As a consequence, the air quality and visibility over much of the Himalaya in late spring is not as good as at other times of the year.

Springtime weather in the Karakoram is not that much different than winter weather, the main difference being the gradual increase in air temperatures. By the middle of May, air temperatures in the 2500-4000 m (8,200-13,100 ft) elevation range are ideal for trekking. Unfortunately there are very few precipitation gages in the Karakoram, so it is hard to get a clear picture of the amount of precipitation that occurs during this time period. However, the number of synoptic-scale storms moving through the region decreases dramatically in May.

Summer

By late May the Tibetan Plateau and northern Indian plain heat-up to such a large extent that low pressure forms over the region. This in turn results in the northward shift of the thermal equator (region of warmest temperatures), which causes winds over the central and northern Indian Ocean to blow from the south. These winds are in essence very large-scale thermally generated winds that result in large part from the temperature contrast between the Indian Ocean and the Asian continent. When the low-level winds become southerlies and the sub-tropical jet stream disappears, the stage is set for the development of the summer monsoon. This also marks the end of the spring/early summer Himalayan climbing season. Rain produced by the summer monsoon is responsible for about 80% of the Indian Subcontinent's annual precipitation. The areas that receive the largest amounts of rain are the windward slopes of the Western Ghats (mountains located in southwest India) and the area around the Ganges Delta, including the eastern Himalaya and the hills of Assam.

During the summer monsoon surface winds over southern India are predominantly from the southwest, while the winds over central and northern India tend to be from the south or southeast (Figure 9.2). The bulk of the moisture carried inland from the Indian Ocean is found in the surface layer, which is about 1.0-1.5 km (0.5-1.0 mi) deep. Once the sub-tropical jet disappears in May, it is replaced in the upper troposphere by a large area of high pressure, called an anticyclone. This anticyclone produces easterly winds which extend from just above the monsoon layer into the lower stratosphere. A easterly wind 'jet' does form over central India (15^{o} N), but overall it has little influence on the Himalaya. In fact, over the Tibetan Plateau and the Himalaya, mid-tropospheric wind speeds are typically less than 10 m/s (22 mph), as seen in Figure 9.3.

It is important to understand that the start of the monsoon rains <u>do not</u> occur at the same time all over India. In a normal year heavy rain begins sometime around the first week of June in both the Western Ghats and in the Ganges Delta, eastern Nepal, Sikkim, and Bhutan. Over the next 3 to 4 weeks the leading edge of the monsoon rain slowly works its way up the Ganges River from the Bay of Bengal. As a result, monsoonal rains do not begin in the western Himalaya until late June.

There is considerable day-to-day variation in the amount of rain received across northern India and the Himalaya during the monsoon season. Periods of very heavy rain are generated by troughs that travel with the easterly jet, which have a tendency to move from the Bay of Bengal up the Ganges River, similar to what happens at the onset of the summer monsoon. Sometimes these troughs recurve back toward the northeast as they encounter the polar jet stream. There are some years when the summer monsoon never fully develops, rainfall is greatly reduced and a large part of the Indian Subcontinent experiences a drought.

The region of the Himalaya that receives the heaviest rains stretches from central Nepal eastward across Bhutan into Arunachal Pardesh. The number of sites where precipitation is collected is pretty sparse in this part of the world, however the Terai (jungle) and foothills of Himalaya receive the bulk of the precipitation. Within the higher peaks the distribution of precipitation is complex, as illustrated by the work of Higuchi *et al* (1982) in the Khumbu Himal of eastern Nepal. Along a transect up the Dudh Kosi River valley, June-October accumulated precipitation decreased with elevation (spanning a elevation range of 2800-4500 m). <u>On the mountains surrounding the Dudh Kosi River valley</u> however, precipitation was observed to be 4 to 5 times larger than what fell <u>in</u> the valley. On the ridges and summits, the heaviest precipitation occurred during the day, probably as a result of strong upslope and valley winds. In the valley itself, precipitation was most frequently observed between 6 pm and midnight, in response to convergent mountain and drainage winds.

As a rule of thumb, the further north one travels within the Himalaya during the summer monsoon, that is the closer you are to Tibet, the drier the conditions. For example, on the north side of the Annapurna Range, summer precipitation is a fraction of what falls in the vicinity of Pokhara to the south. If you are planning to trek during the summer monsoon, count on very wet conditions, this includes high water in rivers and streams, muddy trails, and mudslides to name a few. The good news about travel during this time of year is that the winds are for the most part very light. Typically the winds at the 500 mb level (5500 m) over the eastern Himalaya are from the east at 5-10 m/s (10-

20 mph). Even at 300 mb (9500 m) the winds are not much stronger. In Himachal Pardesh and Kashmir the upper level winds are from the west and tend to be considerably stronger than what occurs over the eastern Himalaya. Freezing levels (free atmosphere) during the summer are typically around the 5000 m (16,200 ft) level. Above this elevation you should expect considerable quantities of fresh snow.

Since monsoon rains move northwest from the Bay of Bengal up the Ganges River, a legitimate question to ask is: how far north do they usually extend? Unfortunately it is not possible to draw a line of a map and say with any certainty that the monsoon rains stop here. In general, Kashmir is considered the northern most extent of the monsoon, however there is considerable variation from year to year. In some years abnormally large amounts of summer precipitation occur in the Karakoram and throughout central Pakistan. For example, one of the authors spent a significant amount of the summer of 1986 hiking around in the Karakoram, and overall it was abnormally cloudy and wet. Typically there were three or four days of rainy weather (at lower elevations) followed by the same number of days of clear or partly cloudy skies. Was the higher than normal amount of clouds and precipitation due to the monsoon? Most likely it was. This was also the summer when there were a large number of deaths on K2 (see Chapter 1).

Figure 9.5 shows annual precipitation and the percentage of annual precipitation that occurs during the summer monsoon (June-September). You should note that there is a dramatic increase in precipitation south of Kashmir, due to the monsoon. As noted earlier, as one travels closer to Tibet the more arid the landscape becomes. Mean annual precipitation in Namche Bazar (3353 m) for example is on the order of 94 cm (37 in), while a little further to the north on the lower Khumbu Glacier (4900 m) it is estimated to be around 50 cm (20 in).

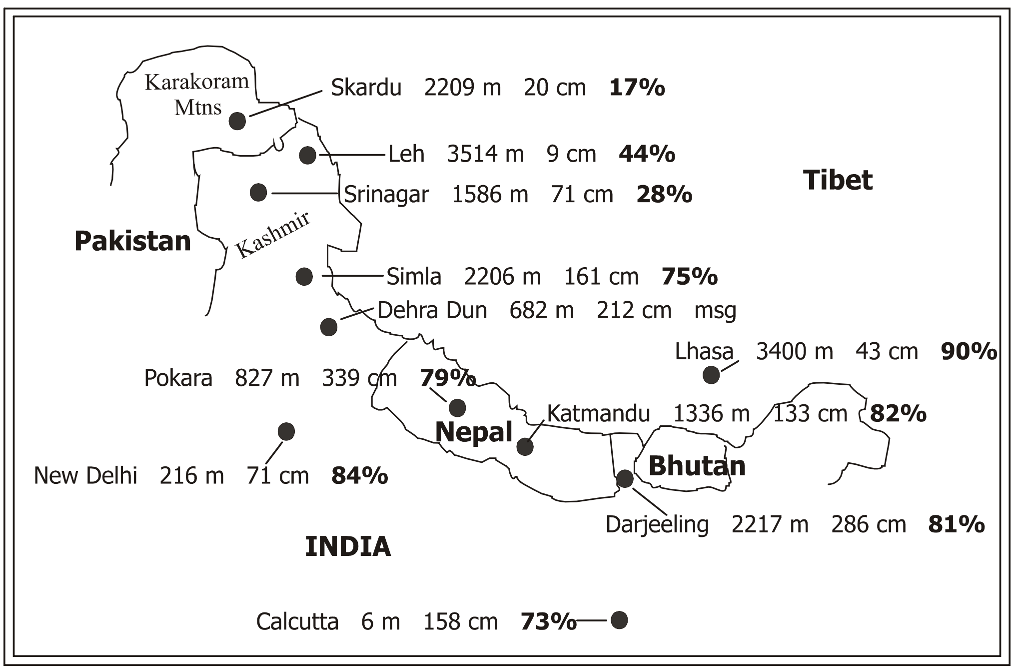


Figure 9.5- Annual precipitation for select stations in the Himalaya and northern India. Order of data is: station, elevation (m), annual precipitation (cm), and percent of annual precipitation which occurs from June-September. Missing data= msg.

Summer weather in the Karakoram is dominated by the polar jet stream. When the polar jet lies over the range, at a minimum you can expect strong winds at higher elevations and modest amount of cloud development. If a low or trough moves into the region, widespread precipitation should be expected as well. The polar jet of course is not a fixed feature over the region during the summer, it migrates between 30º-60º N. This means that the Pamirs and Tien Shan as well as the Karakoram can get hit with some very powerful storms during the summer months. Overall the upper level winds in the Karakoram are considerably stronger than they are in the eastern Himalaya during the summer. Table 9.2 displays the mean monthly precipitation and temperature data (1961-1990 normals) for Skardu (2209 m). Although these values are not representative of higher elevations, it does show the seasonal trends. It is evident that the bulk of the annual precipitation occurs from January-to-May. At higher elevations precipitation occurs anytime of the year there is moderate amounts of moisture in the middle troposphere. At elevations below 3000 m (9,800 ft), during the summer expect hot dusty conditions by day, with mild temperatures at night.

Table 9.2 Skardu Climate Data

	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	**Ann**
mean precip (cm)	2.1	2.4	4.0	2.6	2.6	0.9	0.9	1.1	0.7	1.0	0.6	1.4	**20.3**
mean max temp. (C)	3.1	5.5	11.9	18.4	22.7	28.4	31.4	31.2	27.0	20.0	12.5	5.8	**18.2**
mean low temp. (C)	-7.8	-4.7	1.5	6.7	9.7	13.7	16.7	16.4	12.0	4.8	-1.8	-5.9	**5.1**

* Data provided by Pakistan Meteorological Office

Autumn

Autumn (late September through early November) weather in the Himalaya is prime time for trekking and climbing, because precipitation is generally light and skies can be cloud free.
The main concern for climbers is the re-establishment of the subtropical jet stream during the month of October. Look back at Figure 9.3 and carefully note how by mid-September wind speeds dramatically increase. There is no set week when the STJ re-appears, but when it does, the onset of strong upper level winds is often pretty rapid. The silver lining is that the strength of the upper-level winds tend to fluctuate from day-to-day.

Strong STJ winds occur in association with either storms or clear skies. Occasionally a trough or low will move over the Himalaya during this period, but they are much less frequent than at other times throughout the year.

Table 9.3 Frequency of wind speeds during the month of October in the eastern Himalaya at 8 km.

speed (ms^{-1})	Oct. 1-15 Frequency (%)	Oct. 16-31 Frequency (%)
0-5	0.6	1.0
6-10	9.2	5.4
11-15	32.0	16.7
16-20	37.9	31.2
21-25	15.7	25.6
>25	4.6	20.1

The frequency of October winds over Lhasa are displayed in Table 9.3. These winds should be a good indicator of what climbers can expect in the eastern Himalaya. It is obvious that the winds intensify significantly throughout the month.

During this time of year temperatures start to cool as a result of the lower sun angle but also as a result of the rapid heat loss of the Tibetan Plateau due to a lack of moisture in the atmosphere. Figure 9.4 gives some indication of temperatures that you might expect at a given elevation. Notice how temperatures plummet starting in mid-September.

The eastern Himalaya are unique in that it is one of the few mountainous regions of the world that are influenced to some degree by hurricanes (known in the Eastern Hemisphere as typhoons). Typhoons develop in the Bay of Bengal and produce large amounts of heavy rain in Bangladesh or along the east coast of India. There are two typhoon seasons: April-May and October-November, which of course correspond to the pre-monsoon and post-monsoon periods. We have not seen any kind of study that documents the influence of these storms on the eastern Himalaya, but we would speculate that on occasions considerable amounts of residual moisture works its way into eastern Nepal, Sikkim, Bhutan, and Arunachal Pardesh. When this occurs it could certainly inconvenience trekkers and climbers, however keep in mind that these storms frequently kill tens of thousands of people in Bangladesh and India, so whatever inconvenience you experience is trivial in comparison.

Many travelers to the Himalaya have seen and photographed banner clouds, especially in the Khumbu region of Nepal. Hindman (1995) reports that banner clouds on Everest and Cho Oyu develop and dissipate on a diurnal cycle, reaching their maximum extent in the afternoon. The bulk of the moisture for these clouds is derived from the moist valley winds and convective thermals that pump moisture into the strong westerly winds. It should be noted that a reduction in the size or the dissipation of a banner cloud at night does not necessarily indicate that the strong westerly winds have diminished. The disappearance of the banner cloud in large part is due to the absence of the low-level moisture source.

Autumn in the Karakoram can bring some periods of very good trekking weather. At high elevations its too cold and windy to climb, but below 3500 m (11,500 ft) temperatures are moderating from summertime highs. Be prepared for the occasional synoptic-scale storm that moves over the region and produces wide spread precipitation.

Himalaya Weather Summary

* Wettest period: June-September
* Driest period: October-November
* Sub-tropical jet stream produces high winds at summit-level from mid-October through mid-April. It is important to remember that strong winds can develop anytime the STJ is over the Himalaya.
* Eastern Himalaya is overall considerably wetter than the western Himalaya.
* Western Himalaya is more subject to winter storms with heavier snowfall at mid-elevations when compared to the eastern Himalaya.
* Weakest winds (easterly at mid and upper elevations) occur during the summer monsoon season.
* Due to the transport of low-level moisture into the eastern Himalaya from the Bay of Bengal, prolific cloud formation during the warmer months of the year often occurs at lower and mid-elevations before it occurs at higher elevations.

WEB: Indian Institute of Tropical Meteorology

www.imdmumbai.gov.in/imd.htm

Karakoram Weather Summary

* Wetter period: November-April
* Drier period: normally May-October but depends on northern extent of summer monsoon.
* The polar and occasionally the subtropical jet stream dominate the weather. When the polar jet is located over the range, expect at a minimum strong summit level winds, although precipitation usually accompanies the better developed troughs and lows.
* Unlike the Himalaya, there is no extended period when the polar jet is absent from the region.
* Since the Karakoram are located some considerable distance from an ocean, the bulk of the the moisture is transported from the west and southwest in the middle troposphere. As a result, the bulk of precipitation in the Karakoram falls out at higher elevations (>4000 m).

WEB: Pakistan Meteorological Office
http://met.gov.pk

The Andes

Although certainly not the highest mountain range, the Andes are without question the longest continuous chain of mountains to be found on earth. This range stretches roughly 8000 km (5000 mi) from northwest Venezuela (10º N) through Colombia, Ecuador, Peru, Bolivia, Argentina, finally terminating in southern Chile (54º S). Due to the 64º latitudinal range, the Andes and adjacent regions contain every conceivable climate regime; from some of the wettest regions (Patagonian region of southern Chile) on earth to some of the driest (Atacama Desert in northern Chile). Weather stations are few and far between in the Andes, therefore it is difficult to make a detailed weather assessment. Nevertheless the material in this section provides a basic overview of principal regional

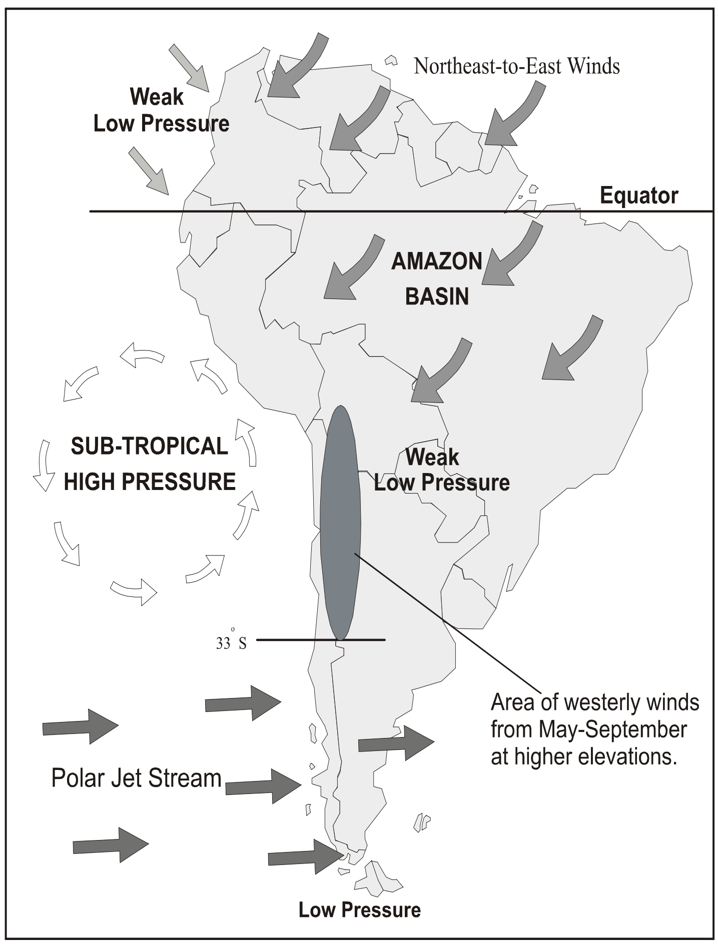


Figure 9.6- Prominent meteorological features over South America during May-September (Southern Hemisphere winter).

weather patterns as well as relevant climate data.

Since the Andes are oriented north-to-south and since the predominate wind direction is either east (in the tropics) or west (mid-latitudes), precipitation and temperatures vary more from east-to-west across the range, than over the same north-to-south distance. The Andes can be classified into many sub-ranges, however one of the standard classifications is to differentiate between the Cordillera Occidental ("western mountains") and the Cordillera Oriental ("eastern mountains"). In the central Andes there is a relatively broad plateau between the C. Oriental and the C. Occidental, that ranges in elevation from 3500 m to 4300 m (11,500-14,000 ft). This plateau is referred to as the Altiplano ("high plateau") or the Puna, and it reaches its greatest dimensions in southern Peru (12^o S), Bolivia, and northern Argentina (26^o S). Elevation is one of the primary controls of weather in the Andes of course, but the influence that the Pacific Ocean and the Amazon Basin have on the range cannot be overstated. One of the important climate elements in the Pacific Ocean is the Peru Current, which flows from south-to-north (Figure 9.6). This current transports cold water which originates in the sub-polar regions as far north as Ecuador. The presence of cold water directly adjacent to the coast has a large influence on the weather of the coastal regions of Ecuador and Peru as well as northern Chile. Since evaporation from a cold body of water is significantly less than from a warm body of water, the air over the coastal zone tends to be drier than what would normally be expected for these latitudes. Equally important to the control of weather in western South American is the presence of a semi-permeant cell of high pressure located over the eastern Pacific. This sub-tropical high, produces a strong trade-wind inversion at a height of about 500-1000 m (1600-3200 ft) above the surface of the ocean. This high migrates north and south parallel to the coast over the course of a year; it is furthest north (20^o S) in June and furthest south in December (35^o S). As a consequence of both the cold Peruvian Current and the trade-wind inversion, the coastal zones of Peru and northern Chile (to about 26^o S) are extremely arid. There are places in the Atacama Desert of northern Chile for example, where measurable rain occurs roughly once every 10 years. In one of those 'freaks of nature' it should be noted however that this same arid coastal zone is frequently blanketed by marine stratus clouds. Due to the low height of the trade-wind inversion, these clouds are shallow and therefore are not capable of forming rain droplets.

Between 33^o-45^o S, the climate along the Chilean coast and the western slopes of the Andes is similar to the climate of western Oregon and Washington. The primary control of weather in these mid-latitudes is the polar jet stream (Southern Hemisphere equivalent to the polar jet stream which produces high winds and stormy weather over North America). The southern tip of Chile (45^o S) has a very wet and windy climate because it is surrounded by water. The Circumpolar Ocean which encircles Antarctica, is a breeding ground for storms due to the temperature contrast between the relatively warm water and the very cold air over the Antarctic icecap.

The Amazon Basin forms the eastern border of the northern and central Andes. The low and mid-level winds over the Amazon are primarily from the east. These easterly winds transport large amounts of moisture from the Amazon to the eastern slopes of the Andes. Because of the extreme aridity in the coastal zone, the C. Oriental of Ecuador, Peru and Bolivia tend to be much wetter than the C. Occidental (Figure 9.7). Since rain and snowfall data is very scarce in the high Andes, it is very difficult to establish accurate precipitation maps. We do know however that there are a number of very wet regions. The eastern slopes of the Ecuadorian Andes are very wet as will be described below. Another wet area is the western slopes of the Andes in southern Chile, south of 45^o S, where annual precipitation is on the order of 5-8 m (16-26 ft) in some places. Keep in mind that due to the

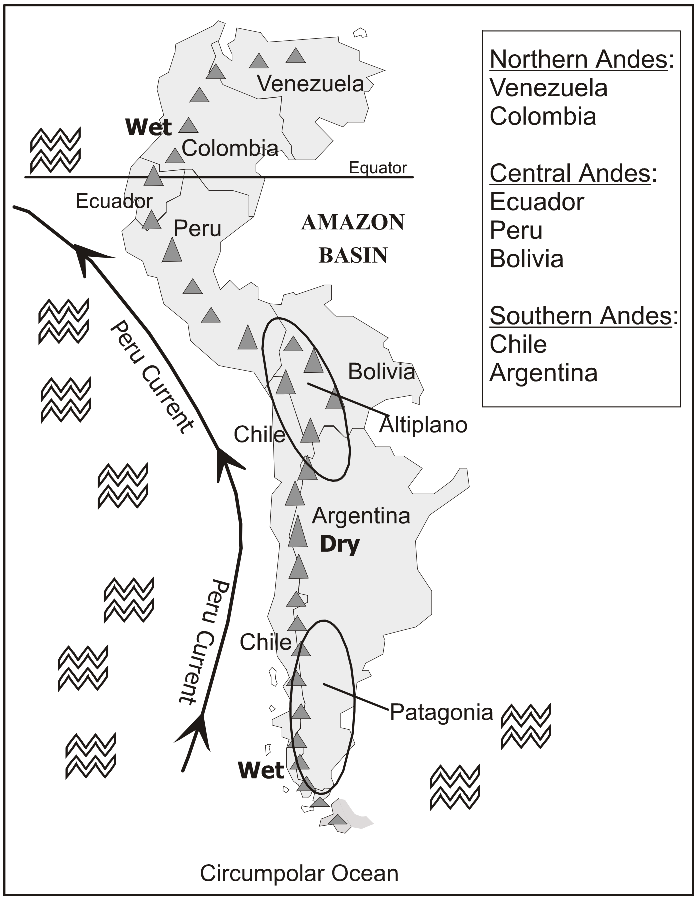


Figure 9.7- Geography of the Andes.

transport of low-level moisture, the 500-2000 m (1,600-6,400 ft) elevation zone on either side of the Andes receives the bulk of the precipitation.

The primary climbing season in the central Andes (Ecuador south to northern Chile and Argentina) is May-September, which corresponds to the Southern Hemisphere winter. This is one of the few mountainous regions in the world where the best climbing weather occurs in the winter. The reason for this is due to the low-latitude position of the central Andes. During the summer months of Nov-April, the central Andes receive some 60-90% of its annual precipitation. From May-September, the moisture laden easterly winds diminish, producing a significant 'dry period'. There are a number of exceptions to the aforementioned rule of thumb as will be noted in the following sections.

Note: *In the Southern Hemisphere, air flows in the opposite direction around HIGHS and LOWS when compared to the Northern Hemisphere. Therefore, air moves counterclockwise around HIGHS and clockwise around LOWS. Likewise the Coriolis deflection is to the* <u>*left in the Southern Hemisphere*</u>*. Figure 9.8 displays a generic 500 mb height field which would be valid from May through September (winter).*

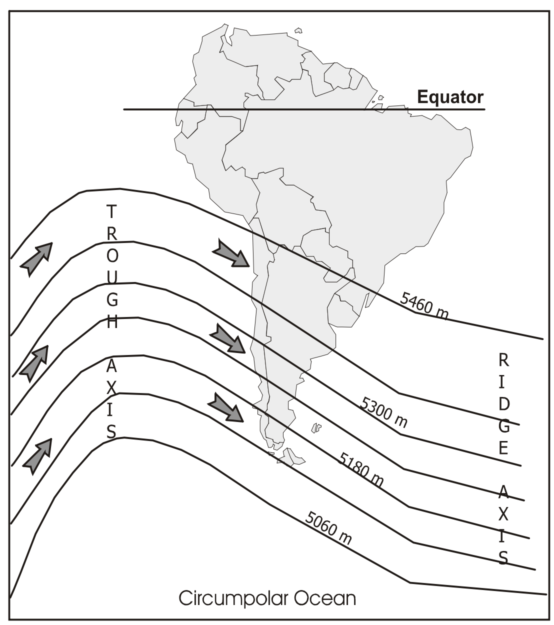


Figure 9.8- A typical 500 mb pattern over South America during the winter (May-September). Notice the difference in the orientation of troughs and ridges with respect to the Northern Hemisphere.

Ecuador

For a small country Ecuador has a very complex climate, in part due to its position on the equator, and secondly because it is a narrow wedge (350 km) of elevated terrain sandwiched between the Pacific Ocean and the Amazon Basin. One of the easiest ways to identify climate zones is by noting the types of vegetation that grow in an area, as well as the overall appearance of the landscape. Variations in climate zones are apparent as one travels south along the Pan American Highway from Otavalo through Quito, to Ambato (this whole region is referred to as the highlands). In the vicinity of Otavalo the landscape is quite dry, the native vegetation consist of xerophytic plants and scrubs. Quito is considerably greener than the region to the north, with an annual rainfall on the order of 1.2 m (47 in), most of which occurs between October and May. Just south of Quito is a fertile valley which contains extensive farms and ranches. Another 75-100 km (45-60 mi) further south however the hills turn brown and the landscape is semi-arid. In this region annual rainfall is only about 0.45 m (18 in). Continuing further south the landscape becomes even more arid.

Climate data for Quito is given in Table 9.4. It is apparent that there is little seasonal change in air temperature at Quito, and for that matter across the entire highlands. The decrease in rainfall and the number of days per month with thunderstorms between June and September is evidence of the reduction in moisture transport from the Amazon Basin. In this region the majority of the rain occurs in the afternoon and early evening hours, in response to convective processes. A secondary dry period is evident from December through early February.

Table 9.4- Monthly climate data for Quito, Ecuador (2820 m)

	Temperature (ºC)		Precipitation	Thunderstorm
	Mean Max.	Mean Min.	(cm)	Days
Jan	22	8	9	7
Feb	21	8	13.5	7
March	21	8	14.8	12
April	21	9	16.5	16
May	22	8	11.0	13
June	21	8	4.9	8
July	22	7	2.9	4
Aug	23	7	3.5	5
Sept	23	7	8.4	7
Oct	22	8	13.5	13
Nov	21	8	9.9	10
Dec	21	8	9.4	7

The wettest regions of Ecuador are the lower elevation zones (500-2000 m or 1,640-6,500 ft) on the windward slopes of the C. Oriental and C. Occidental. For example, the eastern slopes below Cotopaxi and above the jungle town of Tena (500 m), are extremely wet. Tena averages about 620 cm (243 in) of rain per year, however in 1969 it received an astronomical 890 cm (350 in). These large rainfall rates are no doubt related to the fact that Tena is located at the base of a 4000 m (13,100 ft) high mountain barrier (Figure 9.9). Quito and the valley 50 km (30 mi) to its south are a wet spot in the midst of a other wise dry highland valley. It appears that this rainfall anomaly is due in large part to the fact that moisture from the east is able to move into the highlands between the “gap” formed by Cayambe (5790 m) and Antisana (5704 m) to the north and Cotopaxi (5897 m) to the south.

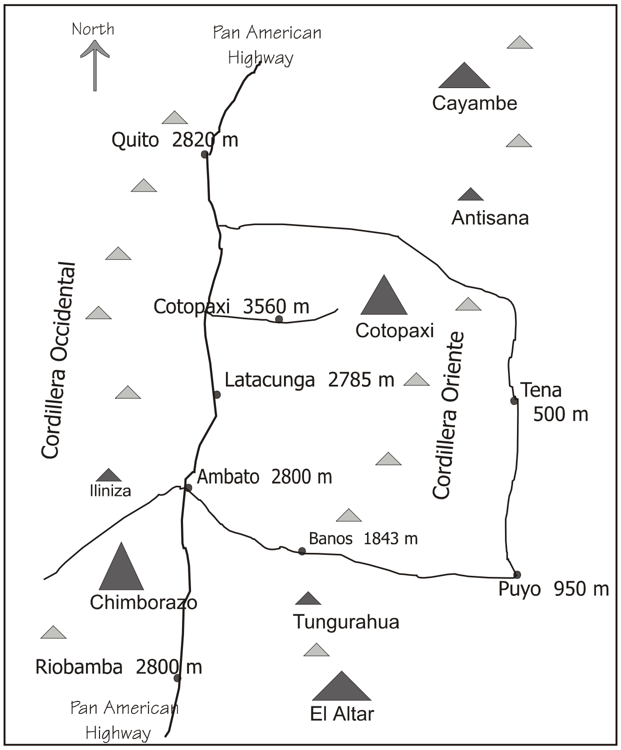


Figure 9.9- Geography of Ecuadorian Highlands.

Switching to the Pacific side of the highlands, the lower slopes of the C. Occidental directly to the west of Quito, typically receive about twice as much annual precipitation as the capital. Starting about another 150 km (92 mi) south however, the western slopes of the Andes (<2500 m) tend be dry, reaching desert-like conditions in northern Peru.

Table 9.5 shows a comparison of average monthly precipitation for five highland stations and the town of Puyo, which is located at the base of the C. Oriental. Note that the Cotopaxi station listed in this table is at the base of the mountain on the west side, not high on the mountain.

Ecuador Climbing Weather

The traditional climbing season in Ecuador is dictated by wet and dry seasons with seasonal temperature changes of little consequence. The driest months in the central highlands are July and August followed by a short dry period centered around January. This should not be taken to mean that skies are cloud free during these months and that perfect climbing weather is assured.

Table 9.5- Average Monthly Precipitation in Ecuador (cm)

	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	**Ann**
Banos	7.2	7.4	9.8	13.1	15.0	19.3	17.8	15.8	13.0	9.1	6.6	6.5	**140**
Cotopaxi	8.9	10.5	14.4	13.3	12.4	8.1	4.6	4.0	8.4	10.6	10.3	8.7	**114**
Latacunga	2.8	5.4	5.4	6.5	3.8	3.2	1.7	1.6	2.8	5.5	5.5	4.1	**48**
Puyo	30.1	29.8	42.9	46.5	40.8	45.7	38.9	34.6	36.5	38.2	36.4	33.2	**454**
Quito	9.0	13.5	14.8	16.5	11.0	4.9	2.9	3.5	8.4	13.5	9.9	9.4	**117**
Riobamba	2.6	4.2	5.4	4.9	3.0	4.0	1.7	1.7	2.9	4.8	4.6	3.0	**43**

Episodes of precipitation and high winds do occur during these periods. Keep in mind that the further east a peak is located, the higher the annual precipitation, which means that there is a greater chance of precipitation during the so called “dry periods”.

In addition, some of the heaviest rain in the jungle occurs from April through July. Even though precipitation in the jungle is on the decline during August-September, considerable amounts “spill over” into the eastern Andes. Cayambe, Antisana, Tungurahua, and El Altar can receive large amounts of precipitation anytime of the year. Rachowiecki and Wagenhauser (1994) report that the highest summit success rate for climbers on the higher peaks of the C. Oriental, occurs in December and January. Analysis of weather data indicates that between mid-December and late January, high pressure forms over northern Brazil. As a result winds over the eastern Ecuadorian Andes tend to be from the north and considerably drier than easterly winds which dominate for most of the rest of the year. This December and January drier period is clearly evident in Table 9.5 at the stations of Banos and Puyo.

Cotopaxi (the mountain) receives substantially less annual precipitation than either Cayambe or Antisana. This is due to the fact that the elevated terrain directly to the east of Cotopaxi intercepts the bulk of the low-level moisture than moves in from the Amazon. As a result of the complex interplay between precipitation and winds, there are two traditional climbing seasons: June through early September and mid-December through late January. March through May should generally be avoided for climbing because of abundant precipitation.

The snow line varies from storm to storm but generally lies between 4000-5000 m (13,100-16,400 ft) . Temperatures in the vicinity of 4700 m (15,400 ft-approximate hut level on Cotopaxi and Chimborazo), typically range from -5º to -10º C (23º to 14º F), with little seasonal variation. At 6000 m (19,680 ft) temperatures range from -15º to -20º C (5º to -4º F). Winds at higher elevations are predominately from the east (northeast-to-southeast) for most of the year.

Peru

Since the low-level westerly winds which originate over the Pacific at this latitude are very dry, virtually all of the rain and snow that falls in the mountains of Peru has to come from the Amazon Basin. As a result, the eastern ranges receive considerably more precipitation than the western ranges. Thunderstorms are a common occurrence during the wet season as well. In the C. Blanco and C. Huayhuash ranges of northern Peru, 70-80% of the annual precipitation occurs from October-April, with January-March being the wettest months (Table 9.6). Prime climbing and trekking weather occurs from May-September.

The mountain ranges around Cuzco (3320 m) the C. Vilcabamba and the C. Urubamba, form the northeastern edge of the Altiplano. About 80% of the annual precipitation that falls in this region occurs between November-April. Keep in mind however, that due to the proximity of these mountains to the Amazon, the weather during the May-September climbing season is not as dry and cloud-free as is it elsewhere in the Peruvian Andes

Table 9.6- Precipitation data (cm) for Peru.

	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	**Ann**
Cuzco	14.9	11.5	9.7	3.8	0.7	0.3	0.4	0.6	2.4	4.7	7.0	10.9	**67**
Huaraz	11.6	10.3	12.6	8.2	2.7	0.3	0.2	1.7	3.1	7.4	7.1	8.6	**74**
Huanuco	5.1	6.5	6.4	2.9	1.0	0.4	0.3	0.6	1.4	3.2	4.3	5.9	**38**
Tingo Maria	43.0	41.8	36.7	28.5	21.9	15.0	14.1	11.9	17.1	30.3	31.4	36.2	**328**

. Also note that the cities and towns of Cuzco, Huaraz, and Huacnuco have considerable amounts of terrain to the east, which of course shields them from receiving large amounts of precipitation. The town of Tingo Maria on the other hand lies at 664 m (2,100 ft) on the eastern slopes, hence it receives the full impact of Amazonian moisture.

Bolivia

The primary destination of trekkers and climbers in Bolivia is the C. Real, which is located between Ancohuma (6427 m) in the north, and Illimani (6462 m) in the south. As elsewhere in the central Andes, May-September (winter) is the dry season, with 80% of the annual precipitation occurring between November-April (summer). During November-April cumulus clouds typically develop in the early afternoon, while most rain showers tend to occur between 5-10 pm. You should note that during the dry season, due to cloud free skies, lower sun angles and low relative humidities, the diurnal temperature range on the Altiplano is on the order of 20º C (36º F), and as large as 30º C (54º F) on occasions. This means the day time highs can be relatively warm, but night time lows are quite cold. In addition, thermally generated winds (valley, mountain, glacier) tend to be a little stronger during the dry season.

It does occasionally snow on the Altiplano during the winter, but generally never more than about ten centimeters (4 in). Short periods of stormy weather during the winter may delay or cancel some flights in and out of La Paz's El Alto airport from time-to-time. Table 9.7 displays monthly average temperature, precipitation, and surface wind direction at El Alto. During the winter, winds over the Altiplano are generally from the west. When these winds are strong, they frequently create large dust storms over the arid landscape. The town of Oruro (3700 m) has a reputation for being a windy dirty outpost of civilization during the winter months because of these storms.

Table 9.7- Climate data for El Alto Airport, Bolivia (4100 m)

	Temperature (ºC) Mean Max.	Mean Min.	Precip. (cm)	Prevailing wind direc.
Jan	13	3	13.0	east
Feb	13	3	11.0	east
March	14	3	7.2	east
April	14	2	4.7	east
May	13	0	1.3	west
June	13	-2	0.6	west
July	13	-2	0.9	west
Aug	14	-1	1.4	west
Sept	14	0	2.9	east
Oct	16	2	4.0	east
Nov	16	3	5.0	east
Dec	14	3	9.3	east

The arid C. Occidental forms the western edge of the Altiplano, and also serves as the border between Chile and Bolivia. What little precipitation that does fall in this region, occurs between October-April. Sajama (6542 m) is the highest peak in Bolivia and is the main attraction in the C. Occidental. An automated weather station was installed on Sajama's summit in October 1996, in order for researchers to get a better understanding of high altitude tropical meteorology (Hardly *et al* 1998). Although the data record is short, observations during 1997 show that minimum daily temperatures on the summit ranged between -13º C and -20º C (9º and -4º F) while maximum daily temperatures ranged from -5º C to -12º C (23º and 10º F). From May-September winds at the summit were primarily from the north or northwest. The strongest winds occurred at night and frequently exceeded 10 m/s (22 mph).

Travelers to the southern Altiplano will observe a very arid landscape. This region is lies directly to the east of the South American subtropical high. This means that there is almost no transport of moisture from the Pacific, and little transport from the southern region of the Amazon Basin as well. What precipitation that does occur (estimated to be 10-20 cm per year), falls between November and April. Due to the lack of available moisture, there are many 6000+ m peaks in this region that are devoid of permanent snow and ice. From May through September the winds are primarily from the west or northwest, and are frequently quite strong.

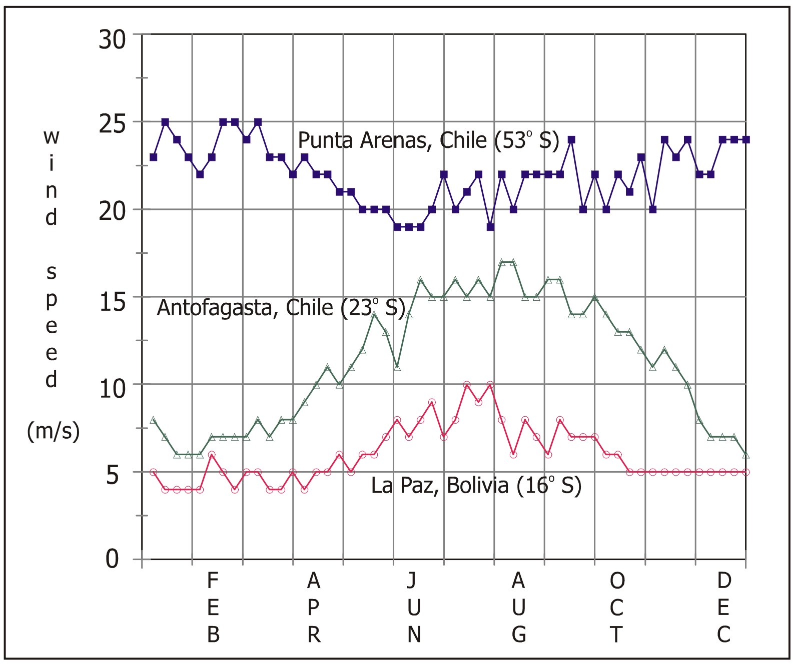


Figure 9.10- Weekly-averaged wind speeds for three South American upper air stations. Data is for 6 km (19,700 ft) above sea-level.

Chile and Argentina

Between 30º-45º S there is a pronounce increase in precipitation as one travels from north-to-south. This section of the Andes is a transition zone from the arid subtropical climate in the north, to the wet maritime climate in the south. This means that westerly winds dominate at all levels in the troposphere throughout the year, as seen in Figure 9.10. Take note that the further south you travel, the strong the winds.

Annual precipitation varies from about 1.5 m (59 in) in the north to 3-5 m (120-200 in) in the south. In addition, there is a very sharp decrease in precipitation on the leeward slopes of the Andes in Argentina. Unlike the central Andes, this region receives the majority of its precipitation between May and August. Table 9.8 displays climate data for Cristo Redentor, Argentina (3830 m), a semi-arid station in proximity to Aconcagua (6959 m). It is evident from Table 9.8 that October-April is the dry season, which corresponds to the southern-most position (35º to 40º S) of the South American sub-tropical high The traditional climbing season for Aconcagua is December-March (Biggar 1996), which as you can see from the wind speed data taken at 6 km over Antofagasta, Chile (Figure 9.10), is the period with the weakest winds of the entire year.

Table 9.8- Climate data for Cristo Redentor, Argentina (3830 m)

	Temp. (°C) Mean	Precipitation (cm) Monthly Mean	# Days > trace
Jan	4.0	1	5
Feb	3.6	1	4
March	1.8	1	6
April	-0.6	2.2	6
May	-4.5	9.6	9
June	-5.9	4.4	10
July	-6.7	5.6	10
Aug	-6.5	6.4	11
Sept	-5.5	2.3	8
Oct	-3.5	1.9	9
Nov	-1.0	1	7
Dec	2.4	1	6

South of 45° S the climate is very wet and windy, similar to southeast Alaska (panhandle). The heaviest precipitation occurs between 500-1500 m (1,600-4,800 ft), on the western slopes of the Andes. Precipitation is pretty evenly distributed throughout the year. There is a slight October-December decrease in precipitation in the climate record at Evagelistas, (elevation 55 m) Chile, but not by much. Estimated annual precipitation on the Patagonian Icefields is 300-500 cm (120-200 in). The permanent snowline is about 2000 m (6,500 ft) in the north of this region but closer to 1000 m (3,280 ft) in the south.

Westerly winds blow at all elevations throughout the entire year, as evident in Figure 9.10 at Punta Arenas. As was alluded to earlier in this section, the Circumpolar Ocean is a breeding ground for strong low-pressure systems. The southern Andes are the only elevated terrain for thousands of kilometers, as a consequence, as these storms move on-shore, large amounts of precipitation are generated. Travelers to Fitzroy and Torres Del Paine National Park in Argentina, take note; even though these parks lie to the lee of the crest of the Andes, they frequently get hammered by major storms. The driest season to visit southern Patagonia is January-to-March, when the storm track is closer to Antarctica than to Patagonia. This also corresponds to the period when the South American sub-tropical high reaches its southern most latitude. Climbers should allow plenty of extra time for possible weather delays.

Andes-Climbing and Trekking Seasons

Colombia: November to March
Ecuador: June to early September, mid-December to January
Peru: June to September

Bolivia: June to September

Chile/Argentina:
north of 27º S: June to September
between 27º-45º S: October to March. Aconcagua: December to March
south of 45º S: January to March, give or take a month.

WEB	Brazilian Meteorological Office	**www.inmet.gov.br**
	Chile Meteorological Office	**www.meteochile.cl**
	Peru Meteorological Office	**www.senamhi.gob.pe**
	Ecuador Meteorological Office	**www.inamhi.gov.ec**

The Alps

The study of weather in the Alps began almost two hundred years ago, with some of the earliest weather stations dating back to around 1820. By the middle of the 19th century national weather services were being organized throughout the Alpine countries. As a result there is an 150+ year continuous record of weather observations in many regions of the Alps, and in some cases there are sporadic records that date back much further.

Beginning in southeast France and extending across Switzerland, northern Italy, Austria, Slovenia and Croatia, the Alps form a 800 km (500 mi) arc across southern Europe . Like any other large mountain range, the Alps are divided into numerous sub-ranges. Several of the more important sub-ranges are the: Jura of northwest Switzerland, the Bravarian Alps of southern Germany, and the Dinaric Alps of Slovenia and Croatia. The Pyrenees Mountains which straddle the border between Spain and France, are not considered a part of the Alps.

Like most mountain ranges the Alps, act as a climate divide; in this case they separate the Mediterranean climate to the south, from the temperate mid-latitude (west) or semi-continental (east) climate to the north. The Alps themselves have three geographic climate zones: the northern slopes, the inner Alpine valleys, and the southern slopes. The northern slopes block cool northwest or northerly winds from reaching the Mediterranean, conversely the southern slopes limit the amount of warm Mediterranean air reaching continental Europe. As a result the northern slopes tend to be cooler and cloudier than the corresponding elevation on the southern slopes. The inner Alps, especially the valleys, tend to be considerably drier than any either the southern or northern slopes. Temperatures in the winter in these alpine valleys can be colder than any other place in the Alps at the same elevation, due to persistent temperature inversions. Overall, the southern slopes are sunnier and have fewer days with precipitation than the northern slopes. However, when precipitation does occur over the southern slopes, the intensity is frequently quite high.

The predominate 500 mb flow patterns are illustrated in Figure 9.1. The so called split jet patter which is in 9.11a, is fairly common during the cooler months of the year, especially in the spring when the high pressure forms over the eastern Atlantic Ocean. When this pattern occurs in the winter, expect sunny days with cold clear nights. When it occurs in the summer expect sunny days with little precipitation, although some convective showers are always possible. On the southern slopes expect some widespread precipitation if the southern jet works its way into the northern Mediterranean basin.

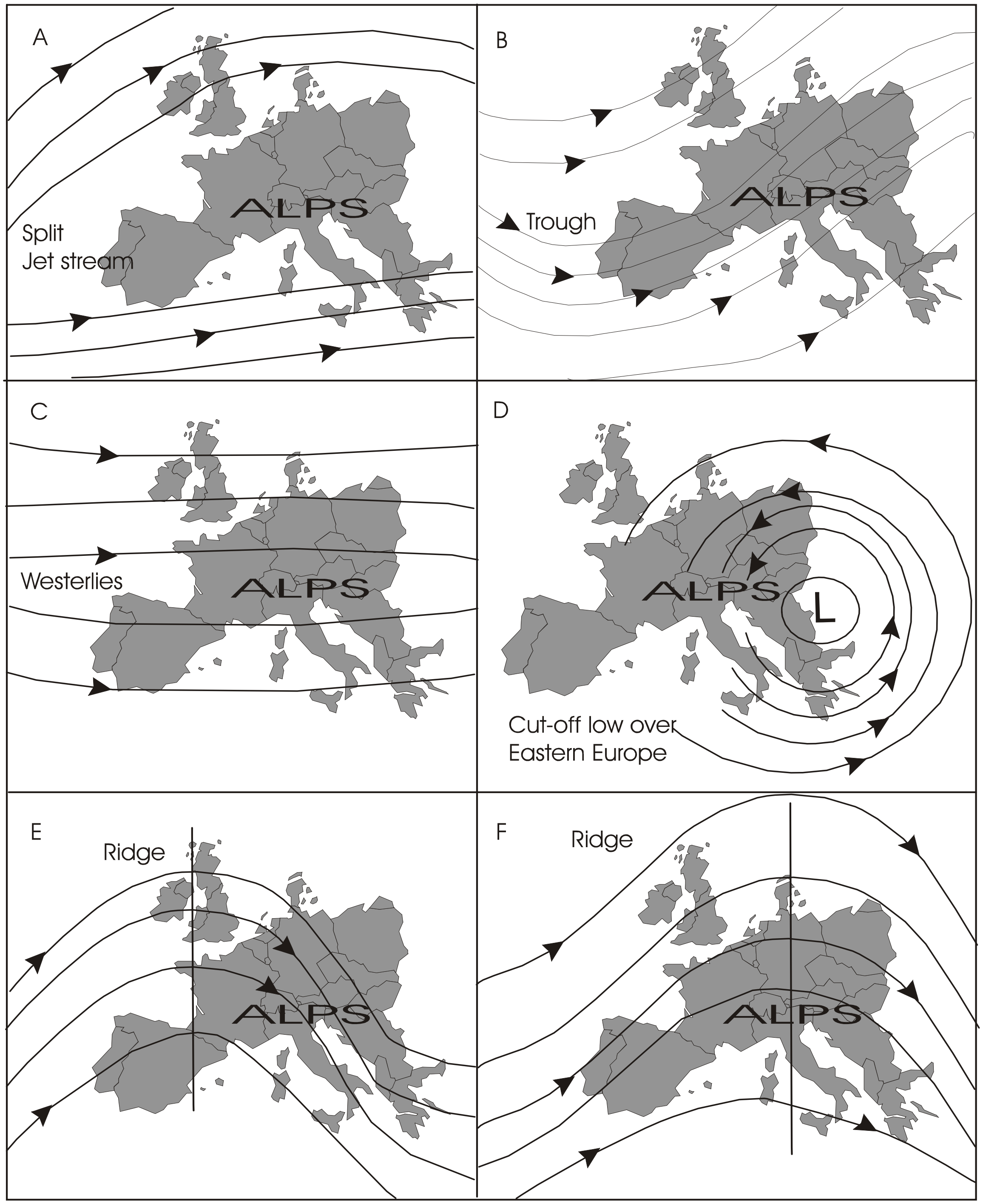


Figure 9.11- Depiction of six common 500 mb flow patterns over Europe.

If the jet remains over the southern Mediterranean, expect dry and sunny conditions over the Alps. When a trough or low is positioned to the west of Portugal, air flows into the Alps from the southwest (Fig.9.11b). This pattern is a major precipitation producer in southeast France, northern Italy and the Dinaric Alps. This pattern occurs more often in the spring and autumn than at other times of the year. When the flow is westerly as shown in Fig.9.11c, weather in the Alps is variable. If the westerly winds are strong, some moisture will be transported into the western Alps. If on the other hand the westerlies are weak, little stratiform precipitation can be expected. This is one of those weather patterns where you also have to look at a surface weather map to see the position of any fronts. Remember that precipitation occurs when several hundred kilometers on either side of fronts.

When a well developed cut-off low is positioned over eastern Europe in the winter (Figure 9.11d), you can expect cold temperatures in the northern Alps, particularly in Austria and Germany. This pattern does not occur very often in the summer, but when it does expect cooler than normal temperatures for that time of year. Since the middle atmosphere becomes cooler with this flow pattern, convection can be enhanced as long as the supply of moisture is not completely shut-off. In Fig.9.11e notice that the ridge axis is west of the Alps, this produces northerly to northwesterly winds over the mountains. Some precipitation could be expected on the northern slopes as moisture is transported from the North Atlantic and North Sea. If the ridge continues to move eastward, the trough that was originally in the eastern Atlantic moves into Europe, resulting in a pattern similar to Fig.9.11b.

Downslope wind storms, which are know as foehn in the Alps, are quite common throughout many regions. They occur on either the northern or southern slopes, depending on the geostrohpic wind direction over the Alps. Foehn occur during the cooler months of the year, with April have the highest frequency. During winter and spring, cold air to the north of the Alps flows through the gaps in the mountains near Lake Geneva, producing a drainage flow know as the bise. When cold air drainage occurs around the western Alps in France, it goes by the name of mistral. In the eastern Alps, a bora is produced when cold air that forms over eastern Europe flows through the lower terrain that separates the Alps from the Dinaric Alps. When a bora occurs, temperatures along the eastern Adriatic coast cool significantly.

Due to numerous deep valleys that dissect the Alps, valley as well as mountain winds are well developed and are a major influence on the weather and climate of towns and villages located in these valleys. It is not uncommon for mountain winds to attain speeds of 10 m/s (22 mph) in the early morning hours, as cold air flows down valley, out of the mountains.

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Precipitation and Temperatures

Along the northern slopes of the Alps average annual precipitation ranges from 100 to 150 cm (39-59 in), with several climate stations receiving well over 200 cm (88 in). There is a distinct summer precipitation maximum throughout this region except in the far west, where the precipitation is more evenly distributed throughout the year (Figure 9.12). Mountain travelers should remember that even though July-September are the wettest months, that the bulk of these summer rains consists of a short burst of convective showers which most often occur during the late afternoon and evening hours. In the winter, most of the precipitation is associated with large synoptic-scale storms that move over the region from the west or northwest. These storms produce wide-spread light snow (or rain at lower elevations) and typically have a duration of one to two days.

Frei and Schar (1998) studied the distribution of annual precipitation along a north-to-south transect of the Alps, roughly from Munch to Trento. They found that areas 50-80 km (30-50 mi) upstream of the base of the Alps, had enhanced amounts of precipitation because of the far-reaching upstream influence of elevated terrain. Their study also indicated that precipitation appears to reach a maximum at an elevation of about 800 m (2,600 ft) on the northern slopes and at about 1200 m (3,900 ft) on the southern slopes. Above these elevations annual precipitation decreases noticeably.

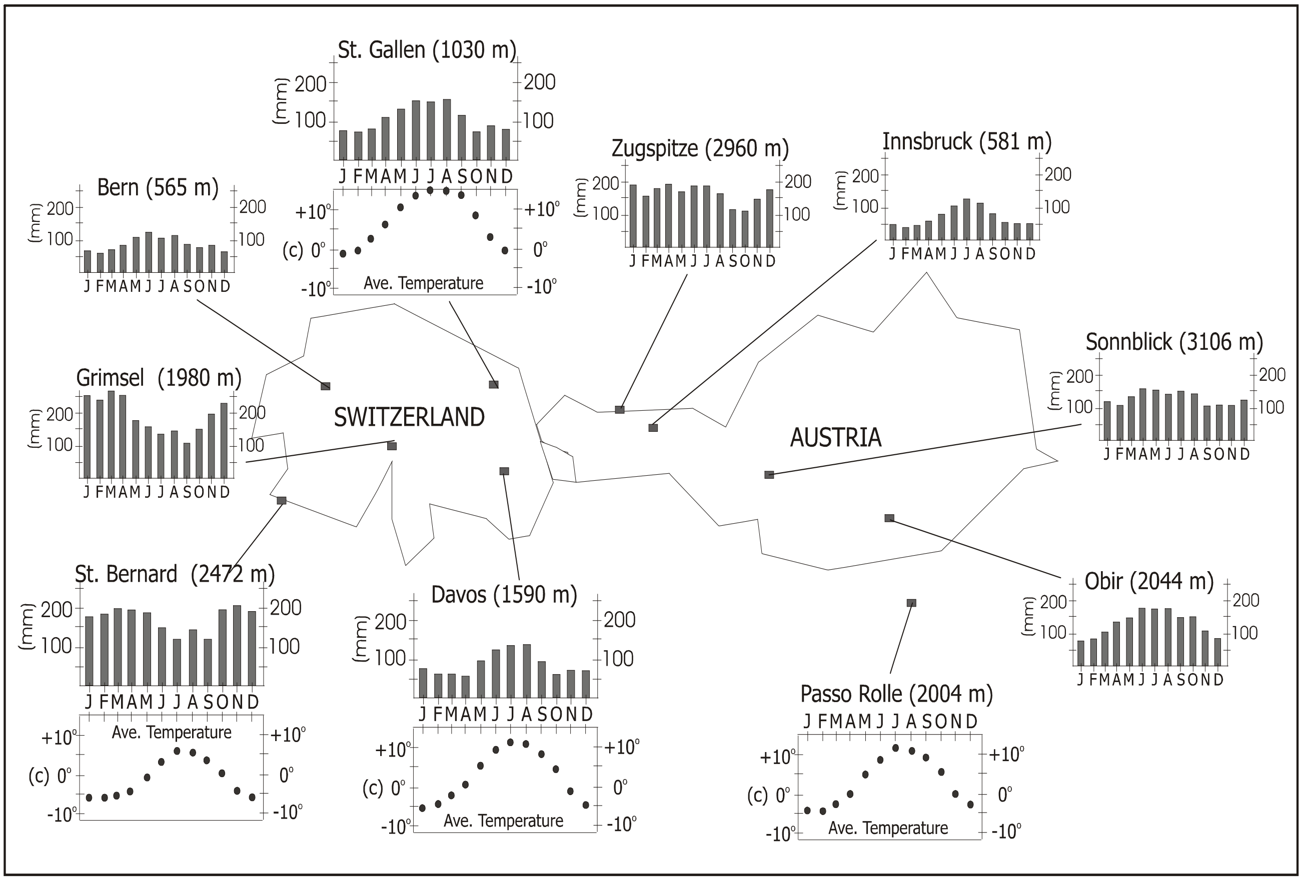

Figure 9.12- Monthly precipitation (mm) and air temperature (C) for select stations in the Alps.

It is worth mentioning that from September through April, a large number of low pressure systems are generated in and near northern Italy, just to the south of the Alps. Low pressure centers form over this region for two reasons: during the cooler months of the year, low-level cold air flows around either end of the Alps. This cold air adjacent to warm air already positioned over the Mediterranean Sea, produces a large temperature gradient, which facilitates the generation of a front. At the same time, as middle and upper level air flows over the Alps and descends over northern Italy, it tends to create an incipient low because the air has a considerable amount of counterclockwise rotation. As these incipient lows move into a region where there is large temperature contrast (gradient), they intensify and become mature systems. The high frequency of these lows during the cooler months of the year, is why parts of the southern slopes of the Alps are some of the wettest regions in continental Europe. The term used by meteorologist for this phenomena is lee-side cyclogenesis, and it also frequently occurs directly to the east of the Rocky Mountains as well.

In the mountains of the southern Alps, precipitation tends to be heaviest in the spring and autumn. Numerous flash floods as well as river floods occur in northern Italy in spring and autumn. In spring, heavy rain can accelerate alpine snow melt, producing floods. In autumn, heavy rain in the mountains often leads to flash flooding. During the summer months expect the usual late-afternoon and evening rain showers with the occasional thunderstorm.

Table 9.9 shows monthly average temperature data from three Swiss stations that are located at different elevations in the Bernse Alps. This data gives a range of high and low temperatures for summer month, and is applicable all across the Alps. On any given day of course, temperatures in the western Alps can differ substantially from those 500 km to the east, however Table 9.9 gives an indication what you can expect in terms of average conditions.

Table 9.9- Alps temperature data

station	(ºC)	May	June	July	August	Sept.
Jungfrau (3580 m)	lows	-12 to -6	-9 to -3	-6 to 0	-6 to 0	-8 to -2
	highs	-7 to -1	-2 to +2	-2 to +4	-2 to +4	-3 to +3
Gutsch (2287 m)	lows	-2 to +2	-1 to +4	+1 to +7	+1 to +7	-1 to +6
	highs	0 to +6	+5 to +11	+8 to +14	+8 to +14	+7 to +13
Interlaken (580 m)	lows	+3 to +9	+7 to +13	+9 to +14	+9 to +14	+6 to +12
	highs	+14 to +21	+17 to +24	+18 to +26	+17 to +25	+15 to +22

Alps Weather Summary

* Precipitation- heaviest along the northern and southern mountains. Due to the barrier effect, the central mountains are considerably drier.
* Summer-time convection occurs over the entire range. The highest frequency of heavy rain and thunderstorms is in late afternoon and evening.
* Gap winds are frequent in many regions during the winter. This winds are best developed when central Europe experiences an extensive cold spell.
* Foehn are common along at the base of the northern and southern mountains, depending on the direction of the large-scale winds. This downslope windstorms are best developed during the cooler months of the year.
* Due to the presence of large and deep valleys throughout much of the Alps, moderate and occasionally strong valley and mountain winds are frequent.

Web Austrian weather/climate **www.zamg.ac.at**
Swiss Meteorological Office **www.westwind.ch**
German Meteorological Office **www.wetterzentrale.de**
British Meteorological Office **www.meto.govt.ul**

Appendices

Appendix One: Answers to quiz questions

Introduction:

1. Rising elevation on a stationary altimeter indicates: (b) decreasing pressure. This information can be important because *it may* indicate the approach of a low pressure system (i.e.-storm).

2. True/False: Air within a low pressure weather system generally moves toward the center of the low ans upward? True- see Chapter 3.

3. True/False: As a general rule of thumb: wind speeds decrease with height in the lower atmosphere? False- winds typically increase with height in the troposphere.

4. What is the windiest season in the Presidential Range of New Hampshire? (c) Winter. You do not have to know anything about this region to hazard a good guess-wind speeds are almost always highest during the winter months in any location, because of the temperature contrasts.

5. Cloud-to-ground lightning has the highest frequency of occurrence between the hours of: (b) 3-7pm. The actual peak time various to some degree from one region to the next, however the best answer is b.

6. True/False: Wave clouds and a mountain cloud cap indicated high winds near the summit of a mountain? True- you will find that more often than not, winds are moderate to high near the summit of a peak that has a cloud cap or is producing wave clouds.

7. True/False: Due to mixing of air in the atmosphere, a climber at 5000 m (16,400 ft) in the Alaska Range experiences roughly the same air temperature as a climber at the same elevation in the Himalaya? False- the temperature on each mountain of course depends on a number of factors, however, statistically (or climatologically), the climber in Alaska will be colder than his/her counterpart at the same elevation in the Himalaya.

8. Large thunderstorms typically develop over what time period: (d) 1-2 hours- you may find this hard to believe because you have observed large thunderstorm cloud systems live a lot longer than several hours. However, most of those thunderstorm systems (complexes) are made-up of many individual storms-hence they appear to have a much longer lifetime.

9. During the summer, air temperatures _______ as a major low pressure system approaches: (b) Cool down- This is true because cool air surrounds the low pressure center.

10. True/False: Most 'ground blizzards' occur after new snow has fallen? True.

11. On a night with no clouds and little wind, pick the location that will have the coldest morning temperature: (c) Floor of a valley- this is so because the radiational cooling of the valley causes the cool dense air to sink toward the bottom of the valley during the night.

12. True/False: Precipitation (rain or snow) always increases with increasing elevation? False- in mid-latitude mountain ranges the amount of snow or rain per storm, per month, per year typically increases from sea-level to some height, after which the amount decreases with further increase in elevation. The height of maximum precipitation depends on a number

of factors: air temperature, proximity to large moisture source, size of the mountain range to name a few.

13. True/False: The primary climbing seasons in Ecuador are May-September and January? True- see section on Andes in Chapter 8 for explanation.

14. True/False: Climate statistics are not useful in expedition planning since the weather on any given day can be dramatically different then the long-term normals? False- If you answered this question with a ‘true’, you should be shot. Of course the weather can be different then the climatic normals, however, climate statistics give you some idea of what type of weather you can expect- rainy season verses dry season, temperature ranges, windy season verses light winds, etc.

15. True/False: Water in the atmosphere always freezes when the air temperature is at or below 0º C (32º F)? False- many people may have gotten this question wrong because we are taught in school that water more or less turns into a solid when it reaches its freezing point. This is correct as long as there are ice or freezing nuclei in the water. In the atmosphere, in order for liquid water to turn into a solid, it most do so on some type of foreign material (dirt, dust, aerosols). If there are not enough of these ice nuclei in a parcel of air, the water will continue to cool below 0º C (32º F) as a liquid. See Chapter 5 and Appendix 3.

16. A large cumulus cloud generates the following types of ‘wind’: (d) all of the above. Updrafts and downdrafts are best developed within the cloud tower, horizontal winds can form near the ground as downdrafts make contact with the earth’s surface and are forced to ‘spread out’. See Chapter 5.

17. True/False: Wind chill temperatures <u>increase</u> with <u>decreasing</u> wind speeds? True- note carefully how the question is worded. Wind chill temperatures decrease (get colder) as wind speeds increase, so as wind speeds decrease- the wind chill temperature increases (warms).

score - for introduction quiz.

# <u>correct answers</u>	<u>Action required</u>
15-17	Climb on!!
11-14	Good show! - you know what your doing.
8-10	Your on the right track - with a little work you will be a weather god.
4-7	Your weather skills need improvement - read on!
0-3	Stay at home, lock the doors and watch TV!

Chapter 2 quiz

1. Name the four types of fronts: cold, warm, stationary, occluded.

2. True/False: The Coriolis force only changes wind speed, not wind direction? False- it changes or deflects wind direction. To the right in the Northern Hemisphere and to the left in the Southern Hemisphere.

3. Oxygen is the: 1st, 2nd, 3rd, 4th most abundant gas found in a sample of air. Second.

4. A rising altimeter signifies what? A drop in pressure- in cases when the rise (drop) is large, it may indicate a developing storm.

5. Air moves in a _______________ direction around an are of low pressure in the Northern

Hemisphere. Counterclockwise. (*Anticlockwise* for our British and Commonwealth readers)

6. True/False: More oxygen is available to a climber on the summit of K2 than to a climber on the summit of Mt. McKinley? False- more oxygen is available to you on McKinley than on K2.

7. A cold air mass moves: a) over b) through c) under, a warm air mass. c) under.

8. A westerly wind is moving in what direction? From west-to-east.

Chapter 3 quiz

1. True/False: The sky appears blue because oxygen decreases with elevation? False- The sky's color is a result of sunlight scattering off of nitrogen molecules. The blue sky gets darker with increase in elevation because of a decrease in air density- essentially there is a reduction in the amount of scattering that occurs.

2. ___________ is the transfer of heat from a hot to a cold body. Conduction.

3. Clouds absorb and emit ______________ radiation. Longwave.

4. True/False: In the Northern Hemisphere, south facing slopes generally receive less shortwave radiation than north facing slopes? False- south facing slopes receive more SW radiation.

5. For a given air temperature, the wind chill temperature ___________ as the wind speed increases. Decreases (gets colder).

Chapter 4 quiz

1. Name the three jet streams? Arctic, Polar, Sub-tropical.

2. True/False: Glacier winds only occur at night? False.

3. Barrier jets form when the lower atmosphere is _________. Stable.

4. True/False: Downslope wind storms only form when the upstream wind is nearly parallel with the mountain range? False- the winds should be perpendicular to the mountains.

5. True/False: Valley winds develop faster than slope winds? False- slope winds generally develop faster than valley winds because there is less air to heat over the slopes than in a valley.

6. True/False: The stronger the bond between ice particles, the more difficult it is for the wind to transport it? True.

7. Slope winds take about ________ minutes to develop after sunlight 'heats' the air. Roughly 30 minutes.

8. True/False: Downslope windstorms are common in the western Rockies? False- they are common along the eastern slopes of the Rockies.

Chapter 5 quiz

1. True/False: A cloud consists of many tiny liquid droplets and/or ice crystals held in suspension? True.

2. True/False: A wave cloud is only made-up of water droplets? False- it can consists of water droplets or ice crystals or both. Since wave clouds occur at higher elevations, they will however tend to be ice crystals.

3. Cumulus clouds are often __________ than they are wider. Taller.

4. True/False: Lightning is really cool? False- It will burn ya!

5. Supercooled liquid water exists at a temperature ______________ freezing. Below.

6. Orographic precipitation is primarily controlled by __________ speed. Wind speed.

7. True/False: As pristine ice crystals fall through a cloud and merge with other crystals they form what are called ____________. Aggregate snowflakes.

8. True/False: Rainbows only form when the angle of the sun with respect to the water droplet is 22^o ? False- rainbows form at many angles with respect to the sun. Sun halo's produced by ice crystals form at an angle (arc) of 22^o from the sun.

9. In general, cooler air temperatures result in ___________ amounts of snowfall. Smaller.

10. Cumulus clouds consist of updrafts and ___________. Downdrafts.

Chapter 6 quiz

1. True/False: Weather balloons are launched at the same time no matter the location? True- at 0Z and 12Z.

2. Name at least two NWP models: Eta, NGM, NOGAPS, AVN/MRF, RUC, etc.

3. True/False: Radar can be used for long-term weather forecasting? False- radars are used to monitor current conditions and aid forecasting for several hours into the future.

4. What is freezing rain? Rain that falls as a liquid but subsequently freezes as it makes contact with the ground, trees, wires, etc.

5. True/False: Climate data is of little value when you are planning a major overseas expedition? False.

6. True/False: Lightning is not a hazard in the Rockies? False- if you got this wrong you should be made to walk the plank!

7. True/False: The mid-latitude storm track is closely associated with the sub-tropical jet stream? False- its associated with the polar jet stream

8. What range of wavelengths (visible, infrared, microwave) are used in satellite imagery to view/track clouds at night? Infrared.

Appendix Two: Skis and Snowboards

Have you ever wondered how skis or a snowboard slides across the snow? This topic lies within the realm of snow physics, nevertheless it has a lot of commonality with boundary-layer meteorology, hence we offer it to our readers.

Let's start this discussion by relating two experiences that all skiers (boarders) have: 1) when at rest skis have the tendency to resist motion to a certain degree (i.e.-stick); 2) when the temperature of both snow and air is cold, skis have reduced glide, sometimes very little glide at all. The common theme in these two cases is friction. Physics teaches us that there are two kinds of friction: static (at rest) and dynamic (in motion). When you are at rest on level ground it takes a certain amount of energy to move the first meter, however, it takes less energy to move the second meter. This results from the fundamental fact that static friction is larger than dynamic friction. Friction is the resistance to motion between two material objects (similar to viscosity in fluids). In our example friction takes place between the bottom of the ski (p-tex) and the snow particles in the upper layers of the snowpack. If you think that friction is temperature dependent, you are absolutely correct, the colder the snow, the greater the friction between ski and snow. Nordic skiers know this all too well. Skate skiers have a difficult time producing glide when the snow and air temperatures drop much below -10º C (14º F). At these cold temperatures diagonal skiers may have better luck since the kick zone of the ski requires a considerable amount of friction anyway.

There is an additional consideration that must be factored into the physics of sliding: water. Due to frictional heating between the ski and snow, snow particles directly underneath the ski melt, providing a microscopic layer of water that lubricates the ski and reduces the dynamic friction. The "cool" thing about this process is that it occurs when both the bottom of the ski and the snow are below freezing. This is possible because the weight of the skier creates very high pressures on the snow particles, which lowers the melting temperature. Amback and Mayr (1981) have show by empirical testing that the thickness of the water film created by frictional heating of a ski ranges between 5 and 20 microns (one micron equals 10^{-6} m or 0.001 mm or 0.00004 in), which is too small to be visually detected. Their work suggests that the thickness of the water film is temperature dependent; as temperatures rise, the water film increases in thickness. Applying glide wax to the p-tex reduces the roughness of the bottom of the ski, reducing friction, resulting in more speed per unit of energy expended (your energy and the pull of gravity). Racers like to ski on hardpack (not the same as wind slabs) or ice because with a very smooth surface, the force of friction is small, allowing for quick acceleration. In summary, when a ski slides over an uneven (rough) snow surface, frictional heating due to movement and the weight of the skier creates a thin layer of water which provides the tip and tail of the ski with a zone of reduced friction. A ski that slides over a smooth snow surface experiences a limited amount of frictional heating (because the coefficient of dynamic friction is small), resulting in a thinner layer of water at the ski-snow interface.

There are a number of additional considerations which affect ski performance, the first is snow depth and the second is the thickness of the water film. Snow depth is important because even if the thickness of the basal water film is optimal, but the ski is having to push snow out of the way, it's glide is reduced. In Spring and Summer, there is an optimal time in the morning when all of these factors come into alignment, providing some great skiing. Conversely it is possible to have too thick of a basal water film, this reduces glide because of the suction (capillary force) between the base of the ski and the surface of the water. One way to reduce this potential problem is to rill the base of the ski (introduce longitudinal grooves into the wax). This helps reduces the thickness of the water film.

Now for a trivia question, what part of a downhill ski do you think warms the greatest during a run? Colbeck and Warren (1991) took a handful of thermocouples and placed them all over the base of a pair of skis and then made a number of runs. Their data indicates that the ‘warmest’ part of the ski is the area just behind the heel. This is what one would expect from frictional heating, because this is the part of the ski that has the largest downward force (skiers weight). They found that this area of the ski is 1-2^o C (2-3^o F) warmer than either the tips or the tails.

Appendix There: Microphysics (continued from Chapter 5)

Condensation: vapor⇒liquid

Recall that a parcel of air is limited in the amount of moisture it can hold to a large degree (thermal pun intended), by its temperature; the warmer the air the more moisture it can hold before it reaches saturation. Cloud droplets (liquid phase only) form when the RH of a parcel reaches 100%. Saturation occurs when moisture is transported into the parcel increasing the RH, or when the parcel cools to its dew point temperature. Of course a combination of the two can occur as well.

Once a parcel reaches saturation *it does not* spontaneously produce cloud droplets. Due to molecular energy constraints, in order for water vapor to condense, some type of small foreign nuclei must be present. These nuclei are “dust-like” particles, with a typical diameter on the order of 10^{-7} m (0.0001 mm or 0.0000039 inches), and are called cloud condensation nuclei (CCN). Sources of CCN are small soil particles which are carried into the middle troposphere by strong surface winds. Other sources of CCN are volcanic eruptions, forest fires, decaying vegetation, as well as sea salts derived from evaporating sea spray. Small water droplets (10^{-5} m or 0.00039 inches) form when water vapor condenses on the outside of a CCN. A parcel of air with a large number of CCN, will tend to form a cloud with a large number of small droplets. A parcel with fewer CCN will form a cloud with fewer droplets, however, these droplets are larger in size than the droplets contained in the cloud that started out with a larger number of CCN. This turns out to be an important consideration for further droplet growth. Once the RH drops slightly below 100%, condensation stops. Conversely there are times when the supply of moisture exceeds the number of available CCN, and the RH in a cloud parcel exceeds 100%. When this does occur, it is called supersaturation. This is an important process because it leads to very rapid droplet growth when either the parcel gains additional CCN or when larger droplets fall through the parcel from above.

Once small droplets are created they grow primarily by condensation. We have already stated that if the RH drops below 100% condensation stops. Once it stops, all of the small droplets quickly re-evaporate and the condensation cycle does not start until the air reaches saturation once again. So how do larger cloud droplets form in such an environment? Laboratory studies have shown that the water vapor pressure over a small droplet is a function of its radius of curvature (indicator of how curved a surface is). The radius of curvature for a flat water surface, for example, is *infinity* (i.e.-a *very large* radius of curvature), while a small cloud droplet has a *small* radius of curvature. Samples of CCN from different types of clouds reveal a wide variety of shapes and sizes. Sea salt crystals are relatively large compared to “dust-like” CCN, in fact, a unit of oceanic air has fewer CCN than continental air. Studies have shown that the time it takes a cloud to form precipitation sized droplets is considerably less in oceanic clouds than in clouds that form over land. Large droplets typically only form when the cloud contains some larger (giant) CCN or if part of the cloud becomes supersaturated. In other words, if all of the original droplets are of roughly equal size, it is very difficult for larger droplets to develop. Most warm clouds never generate enough large droplets during their life cycle to create precipitation. Cloud droplets essentially stay suspended in the cloud because their fall velocities are so small that any upward cloud motion compensates for the downward pull of gravity.

Deposition: vapor⇒solid

The term deposition is used in the physical sciences to mean the direct transformation of water vapor into ice, without going through the liquid phase. The formation of surface hoar frost is a prime example of deposition The formation of ice crystals in clouds is very similar to the formation of cloud

droplets. In order for deposition to occur an Ice Nucleus (IN) must be present. Ice nuclei most often consist of microscopic clay particles. Early in the 20^{th} century researchers found that embryonic ice crystals preferred IN that had a similar crystalline structure as the ice itself. Shortly thereafter it was recognized that the structure of silver iodide closely resembled that of ice. Since that time silver iodide has been used as a cloud seeding agent. Cloud seeding was a hot research topic in the 1950's and 1960's. The impetus for this work was twofold; to control severe weather (large hail, tornados) by limiting cloud size as well as the size of hydrometeors contained within *Tcu* and *Cb*. And secondly; to "stimulate" the development of precipitation in small cumulus clouds, especially over drought prone regions.   Although cloud seeding is not as popular today as it was, in large part due to legal issues, this technique is currently used to speed-up the dissipation of ice fog in some areas.
The formation of ice crystals occurs at temperatures well below freezing, in fact the preferred temperature range is around -12^o to -18^o C (10^o to 0^o F). Empirical studies have shown that cloud parcels with temperatures just below freezing have a more difficult time producing ice crystals due to molecular energy considerations, while clouds that are very cold have a difficult time producing ice crystals because of the low moisture content. Ice crystals take-on a wide range of shapes and sizes despite having similar crystalline structures. The actual shape of a crystal (plate-like, column, spherical), depends on the pressure, temperature and moisture content of the cloud in which they are forming. As with water droplets, a cloud initially containing a wide size distribution of small ice crystals, is more likely to produce precipitation sized crystals (snowflakes), because the larger crystals grow at the expense of the smaller crystals.

Freezing: liquid⇒solid
Ice crystals can also form when liquid droplets enter a region of sub-freezing temperatures. This may occur for example, when clouds containing droplets cool as they are lifted over a mountain or up a frontal boundary. Further crystal growth occurs as a result of deposition or if liquid water is present through a process called riming. You may be surprised to learn that liquid water can exist at temperatures considerably below freezing, when it does it is called supercooled liquid water (SLW). There are times when due to a lack of impurities within a droplet or the when the distribution of droplet energy restricts the formation of embryonic ice crystals, that liquid droplets do not freeze until the temperature of the water is between -10^o C and -15^{o}C (14^o to 5^o F)
Riming is a process by which ice crystals collect small SLW droplets which instantaneously freeze as they collide with the crystals (contact freezing), forming what is called graupel. This usually occurs when the original ice crystal is fairly large and starts to fall through the lower levels of the cloud which contain SLW. Note that the riming or icing of aircraft works on the same principal. When a plane flies through a cloud containing SLW, the wings act as giant contact nuclei, the result is a thick layer of ice accumulating on the wings. Spreading de-icer on the wings prior to take off limits the ability of SLW to bond to the wings.
Since most clouds that are capable of producing precipitation, contain both cloud droplets and ice crystals, we need to consider what occurs when both are present. Without going into the specific physics involved, we will simply state that the <u>water vapor pressure over ice crystals is lower than over liquid droplets</u>*.* Why is this important ? It is important because in a cloud that is <u>near</u> saturation (as most clouds are), the ambient water vapor pressure causes cloud droplets to evaporate. The water vapor that is evaporated from the cloud droplets migrates to the ice crystals where it is added to the mass of the crystal via deposition. This process is recognized to be important for the rapid growth of precipitation sized ice crystals in many clouds.

Glossary

adiabatic-- Refers to a parcel of air that does not exchange mass or any form of energy with the air surrounding the parcel. This concept is an idealization on the part of meteorologist, however it works quite well for most applications. A parcel that rises adiabatically can cool because it expands, and conversely a parcel that is adiabatically sinking can warm as a result of compression.

advection-- the *horizonta*l movement (transport) of an air mass or very large parcel.

ambient air-- Air that surrounds a designated unit of air, like a parcel, and which has different properties, such as temperature or moisture, from the parcel.

boundary layer-- A layer in the lowest part of the troposphere of variable thickness (roughly 1 km or 0.6 miles) which is influenced by the daily heating/cooling of the earth's surface. Winds within the boundary layer are influenced by surface friction.

buoyance force-- A vertical force that results from a parcel of air being either less dense than the air surrounding, resulting in rising motion; or more dense, resulting in a sinking motion.

continentality-- Regions located in the interior of continents, far removed from the ameliorating affects of the oceans. These regions experience cold winters and hot summers. (e.g. Northern Great Plans, Interior of Alaska)

convection-- A parcel that rises (updraft) as a result of positive buoyance. Convection can occur near the ground as the earth's surface is heated, or in the middle of the atmosphere when a cooler layer of air is positioned over warmer air.

coriolis force-- A force resulting from the rotation of the earth. It causes winds to deviate to the right in the Northern Hemisphere, and to the left in the Southern Hemisphere.

flow separation-- Refers to the separation of a stream of air from the ground due to a localized region of higher pressure.

front-- A boundary between a cold air mass and a warm air mass. There is a strong temperature gradient perpendicular to the front as well.

geopotential height-- Away from the surface, meteorologist use maps of geopotential height or just 'heights', instead of pressure values (isobars). These height lines are drawn from a semi-horizontal level of constant pressure, for example the 500 mb heights are taken from the height that the 500 mb pressure level is above mean sea level. High heights represent high pressure, low heights lower pressure.

geostrophic winds-- Winds in the atmosphere that occur above the boundary layer, resulting from a balance between the pressure gradient force and the Coriolis force. The direction of geostrophic winds is parallel to the lines of equal geopotential height.

Hadley cell-- A north-to-south circulation cell that as a rising branch near the equator and a descending branch around 30^o N/S. The Hadley Cells export energy from the tropics and transports it to the mid-latitudes in both hemispheres.

hydrometeors-- A term used to designate all types of rain drops and snowflakes.

isobars-- Lines of equal pressure drawn on a weather map.

jet streak-- The area of highest wind speeds within a jet stream. Jet streaks will move through a jet stream often producing synoptic-scale storms below.

lapse rate-- The change in temperature with height. The difference in temperature between any two levels in the atmosphere is the lapse rate. Upper-air soundings measure air temperature every few hundred meters, from which temperatures and lapse rates within the troposphere and lower stratosphere are derived

latent heat-- Process of heat energy released or absorbed in a parcel of air as a result of a phase change of water. When water vapor is transformed into a water, or when a water is transformed into ice, heat is given off. Conversely, when ice turns into water, or water into water vapor, heat is taken out of the parcel and used to make the phase change possible. A given amount of water vapor has more energy than water (liquid), which in turn has more energy than ice.

microphysics-- The branch of meteorology that studies the formation of clouds, and the subsequent development of rain and snow.

numerical weather prediction– Using computer models to solve the fundamental equations that rule the dynamics and thermodynamics of the atmosphere.   Models are initialized with current data and then run into the future for a given amount of time.

nowcasts-- Refers to short-term weather forecasts that covers a period of about 6 hours. Nowcasts are generally only issued when weather conditions are changing rapidly.

occlusion-- A type of front in which the cold air has completely wrapped itself around the low pressure center, lifting all of the warm air above the surface.

overrunning-- Process in which warmer moist air moves over the top of cooler dry air. Given enough vertical lift, during the winter this often leads to the production of snow. A common event in or near mountainous terrain.

pattern recognition-- A methodology used by weather forecasters that utilizes the forecasters past experience with a given weather pattern, to predict what is going to happen when that ‘same pattern’ occurs again.

pressure gradient force-- A force resulting from high pressure in one location and low pressure in and adjacent location. Motion in the absence of any other force is from high to low pressure.

rain shadow-- A region downwind of a mountain range that receives smaller amounts of rain (or snow) when compared to the upwind side of range. This occurs because the majority of the precipitation has been forced out of the clouds over the mountains where air is rising.

saturation-- Air is said to be saturated when it contains the maximum amount of water vapor it can hold for a given temperature. When air is saturated its Relative Humidity is 100%.

sea surface temperatures-- Water temperature at or very near the surface of the ocean. SST’s are measured either by ships and buoys, or estimated by sensors mounted on satellites.

scattered or scattering-- Refers to the absorption and re-emittance of electromagnetic radiation by various types of gases in the atmosphere.

shortwave radiation-- Electromagnetic radiation that is visible to the human eye.

snowline-- A transient line that separates an area of snow from snow-free ground. The snowline can vary considerably from one storm to the next. In general, during the summer the snowline gradually moves to higher elevations.

storm track-- Refers to the movement or concentration of synoptic-scale storms, usually along the polar jet stream.

supercooled liquid water-- Water that has a temperature below freezing yet it remains a liquid. The reason this can occur is that in order for water to freeze, it needs to some type of foreign material (aerosols, dusts) or what is called an ice nucleus to initiate the freezing process. If foreign material is absent from a parcel of air, the temperature of water can reach well below freezing. Ice crystals that fall through a layer of SLW, are often rimmed, because the ice crystal acts as an ice nuclei.

temperature gradient-- A region in which there is a large temperature difference. For example, if its 22^o C in Seattle and 18^o C, then the temperature gradient is 4^o C per 285 km. Essential it is the difference in temperature over some horizontal distance (referred to in this book as 'temperature contrast).

temperature inversion-- A layer within the atmosphere where the temperature increases with an increase in height. Inversions typically form within one or two kilometers of the ground and may persist for several hours or be semi-permanent depending on the situation.
trade wind inversion-- Is a persistent temperature inversion which is found over much of the sub-tropics at an altitude of 2000-3000 m (6500-9800 ft) above the surface of the ocean.

tropopause-- Divides the troposphere (lowest section of atmosphere) from the stratosphere.

upper-air data-- This data consists of wind speed, wind direction, temperature, and relative humidity taken every few hundred meters during a weather balloon ascent. An additional name is sounding data. The instrument package attached to the balloon that makes the measurements and transmits data back to the ground station is called a radiosonde or rawinsonde.

virga-- Rain that evaporates or snow that sublimates (changes from solid directly into water vapor) before it reaches the ground. Often viewed from a distance as a dark shaft extending below cloud base.

Recommended Reading

Barry, R.G. 1992. *Mountain Weather & Climate*. London: Routledge. 2nd edition. This book was written for meteorologist but I would recommend it to anyone who is serious about mountain meteorology.

Price, L.W. 1981. *Mountains & Man* Berkley, CA: University of California Press. A book that covers a wide range of mountain environment issues, including mountain weather and climate. Can be found in libraries.

Wallace, J.M. and P.V. Hobbs. 1977. *Atmospheric Science: An Introductory Survey*. Sand Diego, Ca: Academic Press. This is one of several standard college level textbooks. Even though it has some advanced material, there are plenty of topics that a serious student of mountain weather could benefit from. This book can be found in many college bookstores.

Whiteman, C.D. 2000. *Mountain Meteorology: Fundamentals and Applications*. Oxford, England: Oxford University Press. This is a excellent book that makes good use of graphics and photos.

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