
4

Precipitation

All water enters the land phase of the hydrologic cycle as precipitation. Thus in order to assess, predict, and forecast hydrologic responses, hydrologists need to understand how the amount, rate, duration, and quality of precipitation are distributed in space and time.

Global patterns of precipitation were outlined in Section 3.2.2, and the physics of precipitation formation is described in Section D.5. This chapter begins with a survey of the meteorology of precipitation. Following that, we focus on methods of measuring precipitation at a point and of estimating precipitation over a region. Estimates of regional precipitation are critical inputs to water-balance and other types of models used in water-resource management.

Sound interpretation of the predictions of such models requires an assessment of the uncertainty associated with their output, which in turn depends in large measure on the uncertainty of the input values, as discussed in Section 2.5.2. The uncertainty associated with a value of regional precipitation consists, in turn, of two parts: (1) that due to errors in point measurements and (2) that due to uncertainty in converting point-measurement data into estimates of regional precipitation. Thus, a central goal of this chapter is to develop an understanding of these errors and uncertainties.

The remainder of the chapter discusses the climatology of precipitation (including methods of characterizing seasonality and extreme values), gives a brief assessment of inadvertent and intentional human influences on precipitation and their

implications for hydrologic analyses, and concludes with an overview of precipitation quality and the phenomenon called "acid rain."

4.1 METEOROLOGY OF PRECIPITATION

The formation of precipitation requires a four-step process: (1) cooling of air to approximately the dew-point temperature; (2) condensation on nuclei to form cloud droplets or ice crystals; (3) growth of droplets or crystals into raindrops, snowflakes, or hailstones; and (4) importation of water vapor to sustain the process (Section D.5.4). Radiation, mixing, conduction, and horizontal movement from high to low pressure regions can produce cooling, but not at rates sufficient to produce hydrologically significant precipitation. Such rates occur only when cooling is due to vertical uplift: When a parcel of air rises it cools **adiabatically**, i.e., without the loss of heat. (See Section D.2.3.) Rising dry air cools at a fixed rate of $1^\circ\text{C}/100\text{ m}$ (the dry adiabatic **lapse rate**); moist air cools at a variable but lower lapse rate (about $0.5^\circ\text{C}/100\text{ m}$) due to the liberation of latent heat as condensation occurs.

Here, we discuss the three meteorological situations in which significant rates of adiabatic cooling by vertical uplift occur and the characteristic spatial and temporal precipitation patterns associated with each.

4.1.1 Uplift Due to Convergence

As discussed in Section 3.1.3, horizontal air flow is induced by pressure gradients and movement is generally toward regions of low pressure. Thus low-pressure areas are loci of **convergence**; the air converging from several directions is forced to rise, and the rising produces adiabatic cooling. **Frontal convergence** (or **cyclonic convergence**) is characteristic of the mid-latitudes and occurs at the boundaries between **air masses** of contrasting temperature and/or humidity. **Nonfrontal convergence** is largely a tropical phenomenon that occurs within a mass of warm, moist air.

Extra-Tropical Cyclonic and Frontal Convergence

Outside the tropics, between approximately 30° and 60° N and S latitudes, much of the precipitation is associated with **extra-tropical cyclones**. As noted in Section 3.1.2, these cyclonic storms are the principal agents of poleward energy transfer in the midlatitudes.

Extra-tropical cyclones are dynamic wave-like features that form at the boundaries between air masses with contrasting temperatures, moisture contents, and densities. Air circulating for several days over a surface with roughly uniform temperature and moisture conditions takes on the characteristics of that surface and can be identified as a particular type of air mass. Table 4-1 gives the general characteristics of the major air-mass types and their regions of origin. These air masses are of subcontinental scale, covering hundreds of thousands of square kilometers. In general, the centers of air masses are areas of high pressure (**anticyclones**), so they are also the centers of clockwise circulation in the Northern Hemisphere.¹

Figure 4-1 depicts the typical development of an extra-tropical cyclone, a process called **cyclogenesis**. Few actual cyclones follow this sequence exactly, but it provides a framework for understanding the major precipitation-producing processes in the mid-latitudes. Cyclogenesis takes place at an air-mass boundary, or **front**, which is a region about 100

to 200 km wide in which temperature and pressure gradients are relatively large. Atmospheric pressure decreases away from the air-mass centers, and the **isobars** (lines of equal atmospheric pressure) are subparallel. Thus the front develops in a low-pressure trough with relatively cold, denser air to the north and warm, less dense air to the south (Figure 4-1a). If the front is not migrating, it is a **stationary front**. Because the circulations of the air masses are clockwise, wind blows in opposite directions on either side of the front and so creates a zone of **wind shear**. Often, the eddies generated in this zone trigger instabilities in the boundary that appear as waves when viewed in plan.

Cyclonic storms form when the frontal wave develops further, such that the counterclockwise circulation around the wave apex intensifies (Figure 4-1b). To the west of the apex, cold air moves over the surface displacing the warm air, which rises along the frontal surface. This portion of the boundary is now a **cold front**. To the east of the apex, the warm air is displacing cold air at the surface to create a **warm front**, and the warm air rises along the frontal surface here also.

Cloud formation, often leading to precipitation, commonly occurs along one or both frontal surfaces (Figure 4-2a) and at the apex, where the air converging at the center of the counterclockwise circulation must also rise. Typically, cold fronts have relatively steep slopes, about 1 in from 30 to 40, whereas warm fronts have slopes of 1 in from 60 to 120; thus precipitation is usually more intense and areally more concentrated at cold fronts than at warm fronts. Note that the isobars now form a quasi-circular pattern around a low-pressure center at the apex; typically the diameter of a fully developed cyclonic circulation is on the order of 1500 km.

Cold fronts usually move faster than warm fronts, so the evolution of an extra-tropical cyclone typically follows a sequence something like the one shown in Figures 4-1c, d. Where the cold front overtakes the warm front, so that there is colder air everywhere at the surface with warm air above, the front is said to be **occluded**. Although the air-mass contrast at the surface is now weaker, it is still present aloft (Figure 4-2b), and the rising of the warm air over the cold may continue to generate precipitation—in fact, the maximum rainfall intensity usually occurs during the early stages of occlusion (Miller et al. 1983).

¹ In the remainder of this section, cyclonic development will be described for the northern hemisphere. For the southern hemisphere, the directions north and south are to be reversed, and the circulations described are in the opposite sense: cyclones circulate clockwise, anticyclones counter-clockwise.

TABLE 4-1

Characteristics and Origins of Major Air-Mass Types in the Northern Hemisphere.

Air Mass	Characteristics	Source Regions	
		Winter	Summer
Continental polar (cP)	Cold, dry	Arctic Ocean Canada— northern United States Eurasia	Northern Canada Northern Asia
Continental tropical (cT)	Warm, dry	California-Arizona-Mexico Northern Africa-Arabia-northern India	Nevada-Arizona-northern Mexico Northern Africa-Mediterranean- Arabia-central Asia
Maritime polar (mP)	Cold, moist	Northwest Atlantic Ocean	Northernmost Atlantic Ocean-Arctic Ocean
Maritime tropical (mT)	Warm, moist	Northwest Pacific Ocean Central Atlantic Ocean Central Pacific Ocean Arabian Sea-Bay of Bengal	Northernmost Pacific Ocean Central Atlantic Ocean Central Pacific Ocean

Data from Barry and Chorley (1982).

Ultimately, several processes combine to bring the rain production to a halt: (1) the temperature difference between the air masses is decreased by adiabatic cooling of the rising warm air and warming of descending cold air; (2) the pressure contrast that initiated the convergence is reduced; and (3) the inflow of moisture is reduced.

Evolution to the stage shown in Figure 4-1c usually takes 12 to 24 hr, and the process typically is completed in another two or three days (Miller et al. 1983). Because the development takes place in the zone of westerly winds, the low-pressure center at the apex of the cyclonic circulation moves generally eastward, at a speed of about 1000 km day^{-1} , as the cyclone evolves.

Figure 4-3 is a weather map showing a typical extra-tropical cyclonic storm over North America, generating precipitation over an area of some $500,000 \text{ km}^2$. Precipitation associated with extra-tropical cyclones thus covers wide areas, and persists for from tens of hours to days at a given location. Because rates of uplift are relatively low, precipitation intensities are generally low to moderate. As noted above, precipitation at cold fronts usually covers a smaller area and is more intense than warm-front precipitation; on occasion, rates of uplift at cold fronts are similar to those produced by thermal convection (see below) and produce bands of intense thunderstorms.

On a given winter day there are usually about 10 cyclonic storms in various stages of development through the mid-latitudes. The net result of these

storms, in addition to the production of precipitation, is the climatically critical equatorward transfer of colder air and the poleward transfer of warmer air and latent heat of condensation (Figure 3-7).

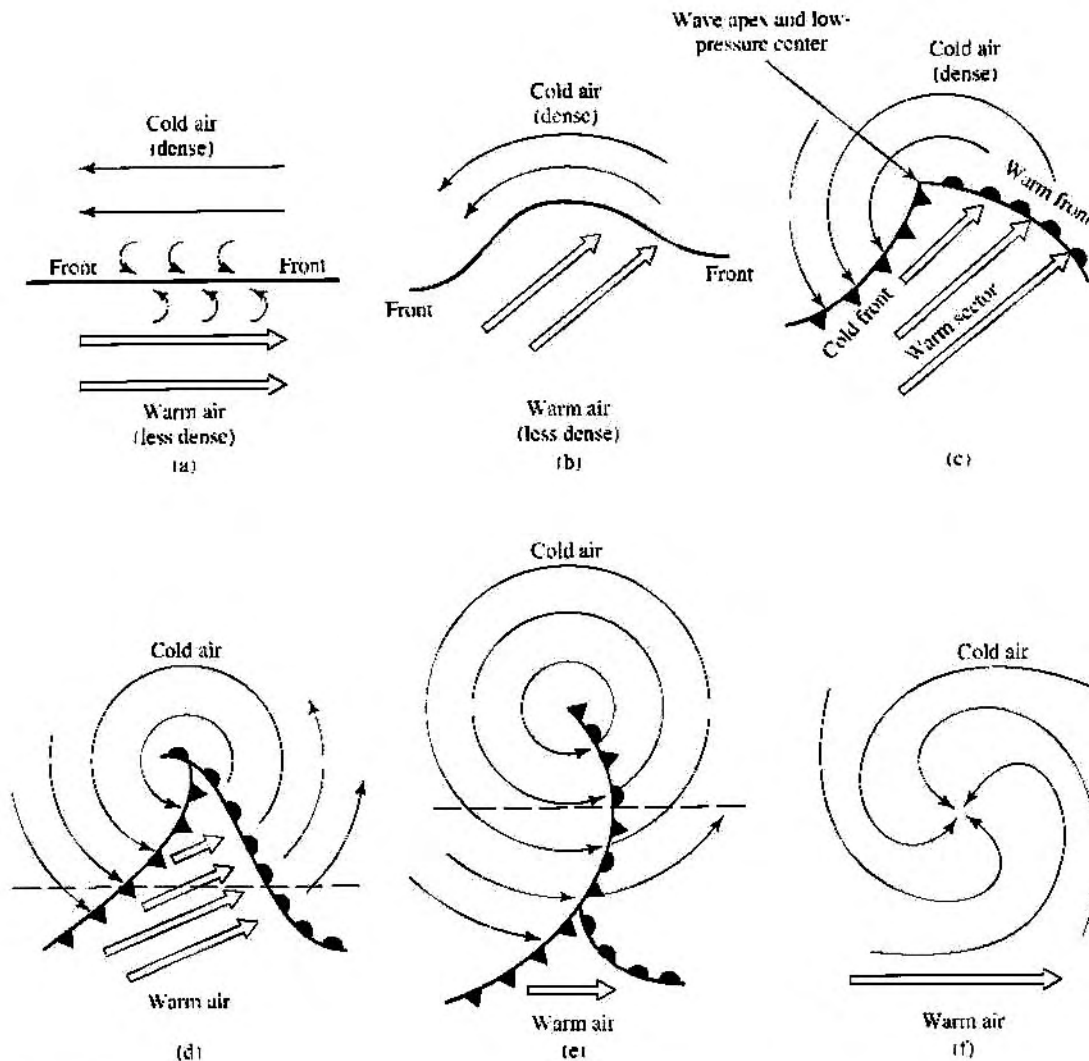
Non-frontal Convergence

Non-frontal convergence is associated with (1) the quasi-permanent intertropical convergence zone (ITCZ; see Section 3.1.3) and (2) temporary storms that form seasonally over the tropical oceans.

Convergence at the ITCZ The low-latitude Hadley cells (Figure 3-8) create a zone of convergence that circles the globe in tropical regions. The ITCZ migrates seasonally and is spatially discontinuous and intermittent (Barry and Chorley 1982), but it is persistent enough to create the equatorial band of heavy precipitation that is apparent in Figures 3-18 and 3-19.

Tropical Cyclones (Hurricanes) Tropical cyclones are cyclonic storms, not associated with fronts, that form over the oceans between 5° and 20° latitude in both hemispheres. They have the potential to develop into extremely intense storms, which are called **hurricanes** (North America), **typhoons** (eastern Asia), **cyclones** (Indian Ocean), and **baguios** (China Sea) (Miller et al. 1983).

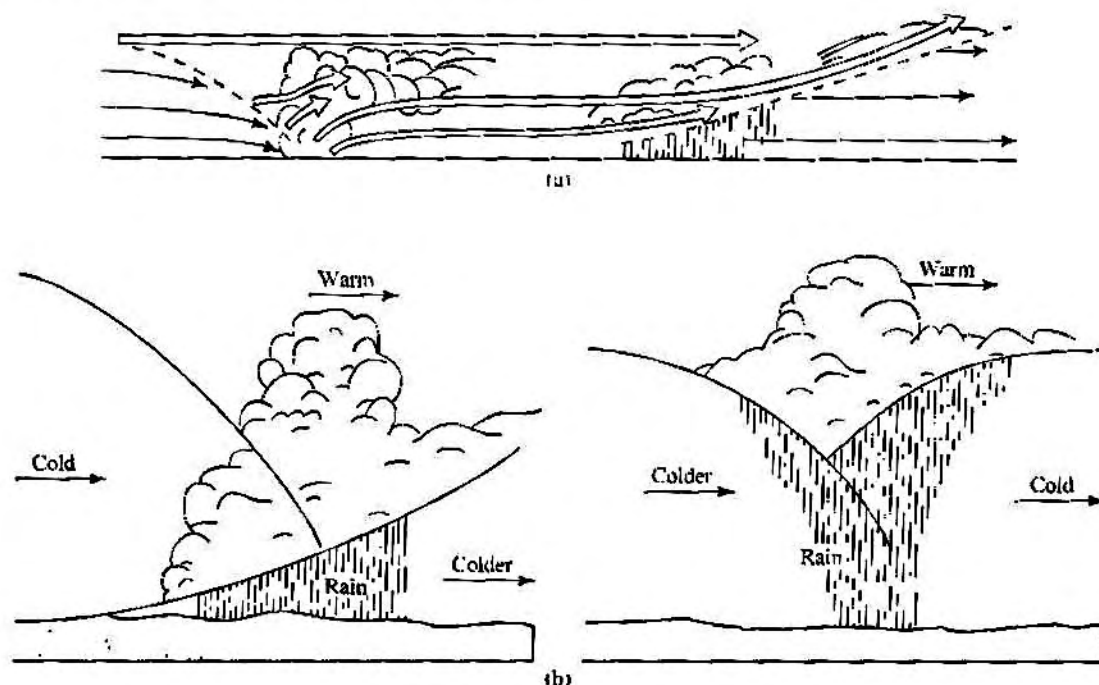
Tropical-cyclone formation begins with a small low-pressure disturbance in a maritime tropical air mass. Sea-surface temperatures of at least 27°C (Miller et al. 1983) are required to induce high rates of evaporation into the converging and rising air.

**FIGURE 4-1**

Typical sequence of development of extra-tropical cyclones (map view). Arrows indicate wind flow. (a) Stationary front between two air masses, showing eddies due to wind shear. (b) Initial wave development from eddy. (c) Intermediate stage with distinct cold and warm fronts and warm sector between. (d) Late stage with occlusion beginning. Dashed line is location of cross section in Figure 4-2a. (e) Front largely occluded. Dashed line is location of cross section in Figure 4-2b. (f) Final stage, with fronts dissipated. Used with permission of Prentice Hall, Inc., modified from *Elements of Meteorology*, 4th ed., by Miller et al., © 1983 by Bell & Howell Co.

the cooling of this air then triggers condensation, and the accompanying release of latent heat further fuels the uplift. Thus the cyclone is fed by evaporation and driven by condensation; if conditions are right, the circulation will intensify until winds near the center reach speeds as high as 65 m s^{-1} (140 mi hr^{-1}). A fully developed hurricane (by convention, a tropical storm becomes a hurricane when its winds exceed 33 m s^{-1}) has the structure shown in Figures 4-4 and 4-5.

Because they form in the belt of easterly winds, the initial movement of hurricanes is usually westward; however, they often move poleward into the zone of westerly winds and may be swept well into the mid-latitudes (Figure 4-6). Because of the reduction in evaporation, their intensity lessens when they move over colder water or, especially, land. However, they can persist for thousands of kilometers over land, moving at speeds of about 5 to 7 m s^{-1} (10 to 15 mi hr^{-1}) and delivering very high rates

**FIGURE 4-2**

Cross-sectional view of extra-tropical cyclone development. Arrows indicate wind flow. (a) Section through warm sector along dashed line of Figure 4-1d. (b) Sections for two types of occluded fronts along dashed line in Figure 4-1e. Reprinted with permission of Prentice Hall, Inc., from *Elements of Meteorology*, 4th ed., by Miller et al., © 1983 by Bell & Howell Co.

of rainfall, as well as destructive winds, over hundreds of thousands of square kilometers.

4.1.2 Uplift Due to Convection

Convective precipitation occurs as a result of adiabatic cooling associated with "parcels" of air that rise because they are less dense than the air that surrounds them. **Thermal convection** is illustrated in Figure 4-7. Line E is the actual temperature in the atmosphere (its slope is the **environmental lapse rate**); the temperature at the surface is initially T_0 . During the day, intense solar radiation heats the ground surface, and locally the air in contact with the surface is warmed by conduction to a temperature T_1 . This cell of heated air is now warmer and therefore less dense than the surrounding air, so it rises. As it rises, it cools initially at the dry adiabatic rate (curve D). If the air is relatively dry, it will continue to rise and cool at the dry adiabatic rate until its temperature reaches the ambient temperature (point S). At this point, the atmosphere is stable: further rising would make the parcel colder and denser than the surrounding air, so convection

ceases at this level. However, unstable conditions will persist if the parcel contains enough moisture so that its dew point is reached during uplift, and cooling occurs at the moist adiabatic rate (curve M). In this case, the parcel's temperature remains higher than ambient, and uplift and condensation continue, often leading to significant precipitation. Figure 4-8 shows cumulus clouds formed due to thermal convection over Cuba.

Thermal-convection cells usually cover areas of a few square kilometers. Rates of uplift due to thermal convection can be very high, reaching from 10 to 30 m s^{-1} (Barry and Chorley 1982), in contrast to about 1 m s^{-1} along a typical frontal surface. However, the frictional drag caused by the falling rain acts as negative feedback and reduces the uplift velocity. Thus, these cells produce very intense rain, often accompanied by lightning, thunder, and hail, covering small areas and lasting less than an hour.

4.1.3 Uplift Due to Orography

In most regions of the world, long-term mean precipitation increases with elevation; this relationship

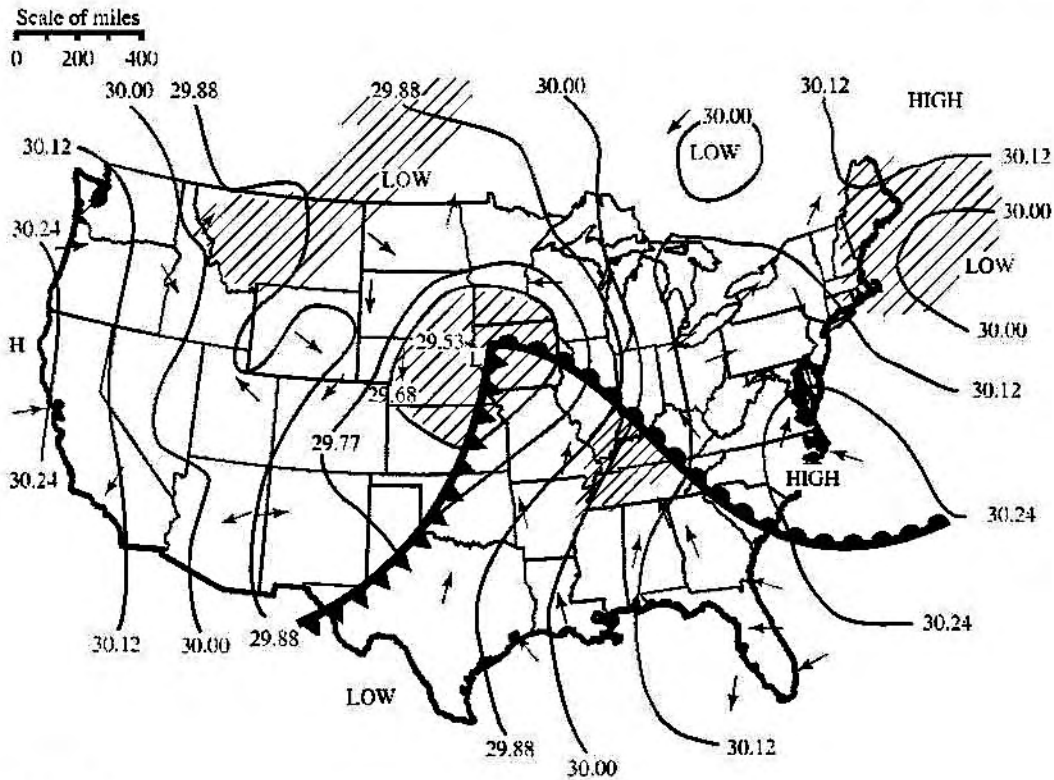


FIGURE 4-3
Weather map showing a cyclonic storm over the United States on 22 April 1970. Solid lines are isobars (pressure in in. Hg). Shaded area is region of precipitation. Arrows show wind direction.

is referred to as the **orographic effect**. Although this effect is widespread, the rates of increase of precipitation with elevation vary widely from region to region (Figure 4-9) and can even reverse at the highest elevations, especially in tropical regions (Barry and Chorley 1982). Barros and Lettenmaier (1994) reported orographic enhancement of precipitation by factors of from 150 to 185% for hills of up to a few hundred meters relief and of more than 300% when relief exceeds 1–2 km.

In the simplest situations, the orographic effect occurs because air moving horizontally encounters a topographic barrier and acquires a vertical component of motion as it flows over that barrier. The rate and degree of cooling are determined by the horizontal wind speed, the wind direction relative to the barrier, the steepness and height of the barrier, and the temperature and humidity conditions that control convective stability.

Clouds and precipitation form on the windward slope of the mountains, and peak precipitation often occurs windward of the topographic crest. The

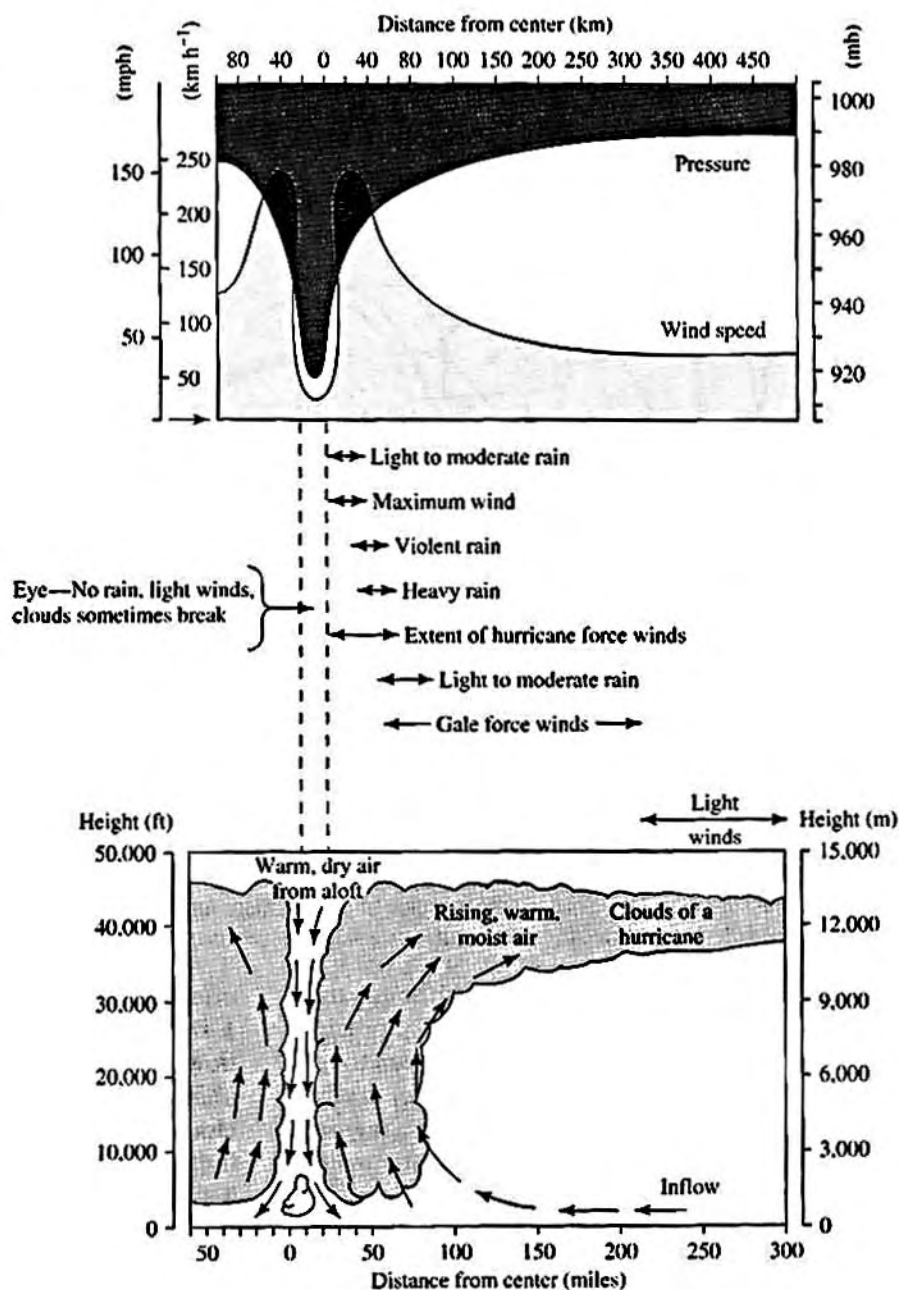
effect commonly persists across the crest for at least a short distance above the leeward slope, but the downward air movement on the lee side causes adiabatic warming. This warming tends to dissipate the clouds and thus turn off the precipitation-producing process and so produces a **rain shadow** (Figure 4-10). Within a region of orographic effect, the precipitation–elevation relation varies as a function of the aspect and inclination of individual slope facets and their relation to local topographic barriers.

The classic orographic situation exists on the central west coast of North America, where several successive mountain ranges are all nearly perpendicular to persistent moist westerly winds from the Pacific Ocean. Figure 4-11 shows the close relation between long-term average precipitation and topography over a distance of 850 km in this region; the reduced precipitation east of the Fraser Valley is presumably due to the fact that the air has lost most of its moisture by this point.

One of the most pronounced orographic effects occurs in conjunction with the monsoon over the

FIGURE 4-4

Vertical section of a fully developed hurricane showing patterns of wind, pressure, and rain. Reprinted with permission of Prentice Hall, Inc., from *Elements of Meteorology*, 4th ed., by Miller et al., © 1983 by Bell & Howell Co.



Indian subcontinent. In summer, a northward shift in the position of the intertropical convergence zone (see Section 3.1.3) and accompanying changes in circulation patterns induce moist southerly winds from the Bay of Bengal. These winds converge and rise over the Khasi Hills in Assam, India (north of Bangladesh), producing persistent heavy rains and giving the city of Shillong (elevation 1598 m) one of the highest average annual rainfalls in the world, 2200 mm, more than half of which occurs in June, July, and August.

In many, if not most, situations, orographic effects are the result of convective, frontal, or cyclonic mechanisms interacting with topography rather than a separate precipitation-generating mechanism. Even in the classic situation shown in Figure 4-11, much of the precipitation comes from extra-tropical cyclones; the extra uplift provided by the mountains increases the cyclonic precipitation on windward slopes and the downdrafts reduce it on leeward slopes (Bruce and Clark 1966).

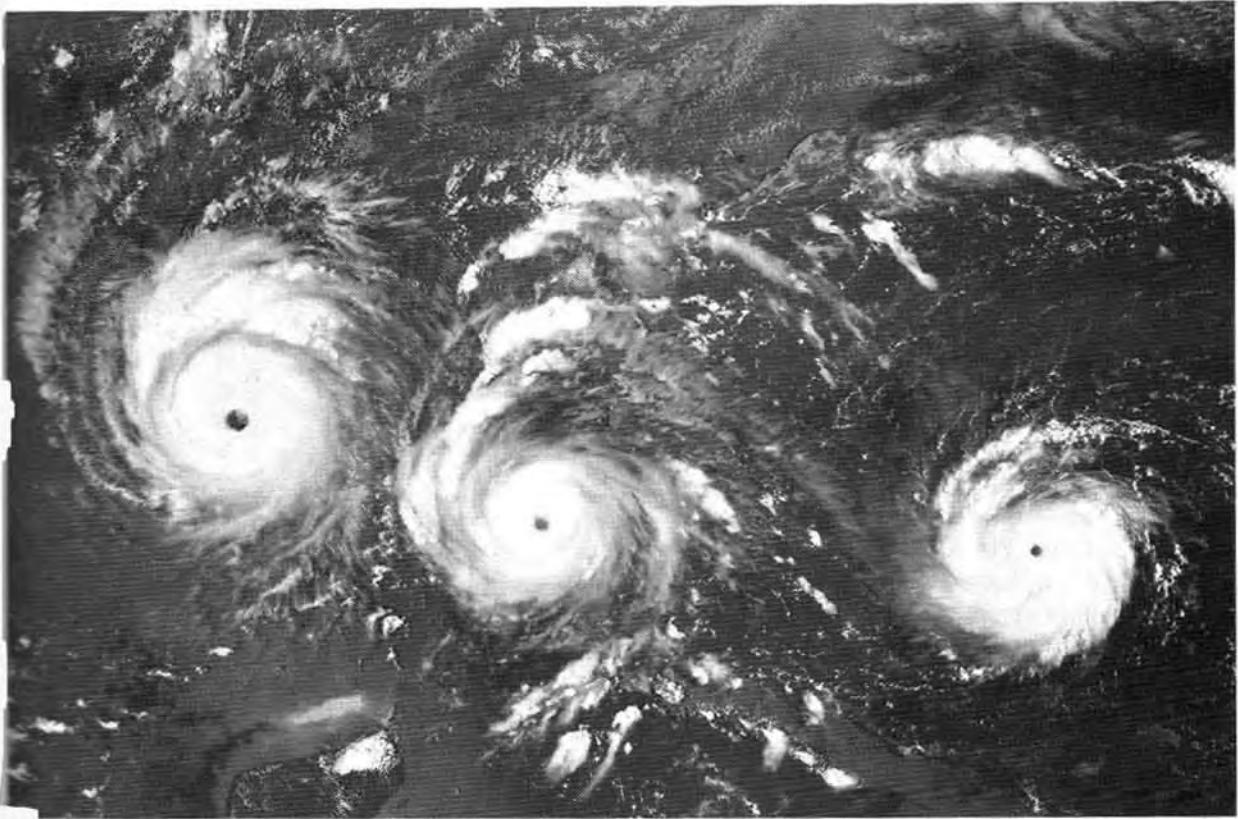


FIGURE 4-5
Satellite images of Hurricane Andrew moving across the southeastern United States, 24–26 August 1992.

Under some circumstances, the frictional resistance offered by topographic rises of only a few meters can cause sufficient uplift to produce precipitation (Barry and Chorley 1982); Figure 4-12 shows cumulus clouds induced by relatively small hills in the southern Great Plains.

An increase in precipitation with elevation can be due to (1) an increase in the fraction of time precipitation falls at higher elevations, which could be called the pure orographic effect, and/or (2) higher intensities at higher elevations when precipitation is occurring at all elevations, which would represent enhancement of precipitation triggered by convection or cyclonic disturbances. The first effect is shown in Figure 4-13 for an intensively gaged New England region; it was also found by Hendrick et al. (1978), who reported a three-fold increase in hours of winter precipitation between 400 m and 1200 m elevation in another area of northern Vermont. However, Hendrick et al. (1978) attributed about

75% of the difference in precipitation at the two elevations to higher intensities at the higher elevation when precipitation was falling at both elevations, so that enhancement accounted for most of the orographic effect in that region.

In regions with significant relief and otherwise similar climate, the orographic effect commonly accounts for much of the spatial variation of precipitation. For example, Dingman et al. (1988) found that elevation accounted for 78% of the variation in long-term mean precipitation in New Hampshire and Vermont. Thus one can often exploit the effect in estimating the average precipitation over a region, as in the hypsometric method discussed in Section 4.3.2. The orographic effect can also carry over into many hydrological relations: Dingman (1981) showed that average streamflows, dry-season streamflows, and floods were all significantly related to elevation in central New England.

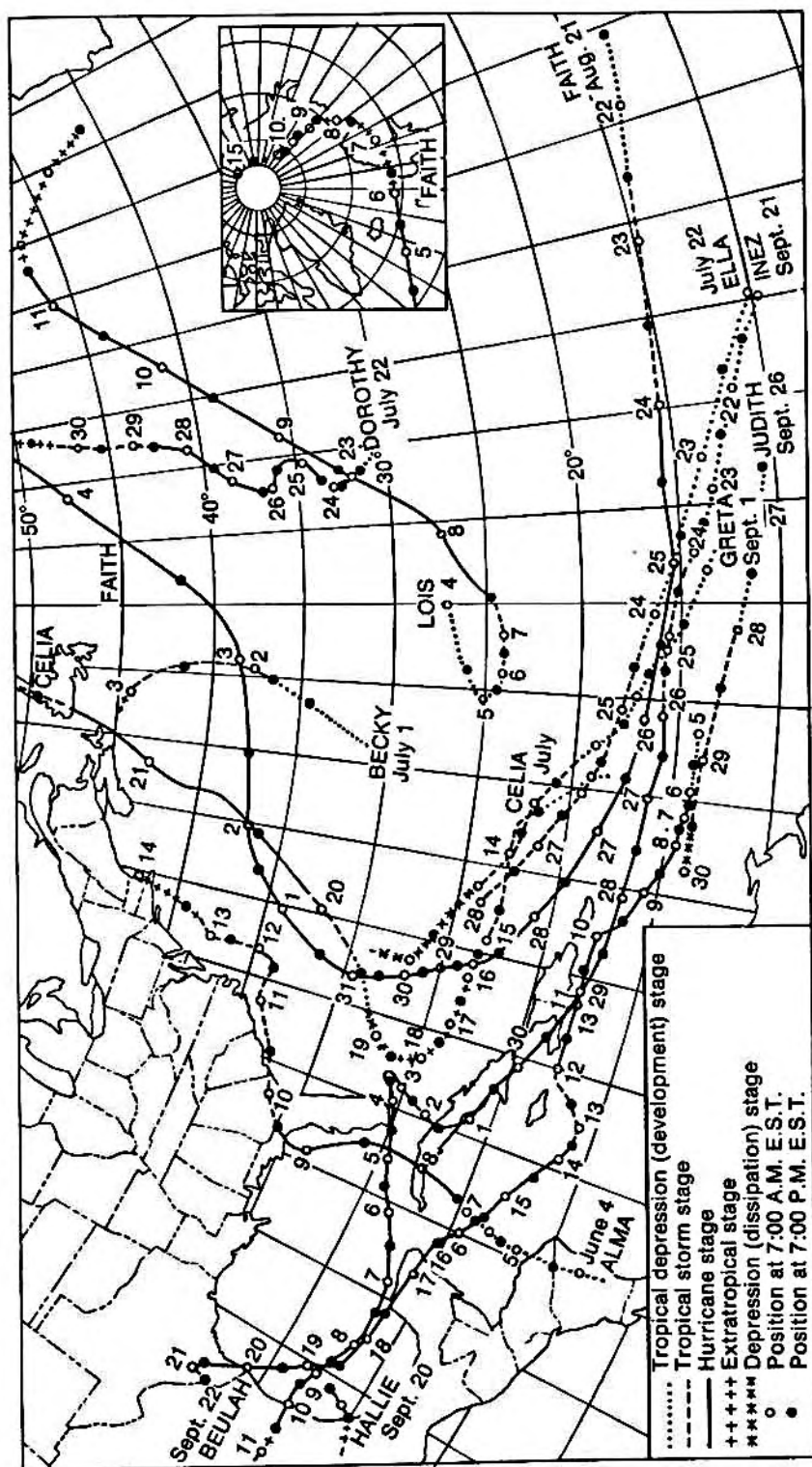
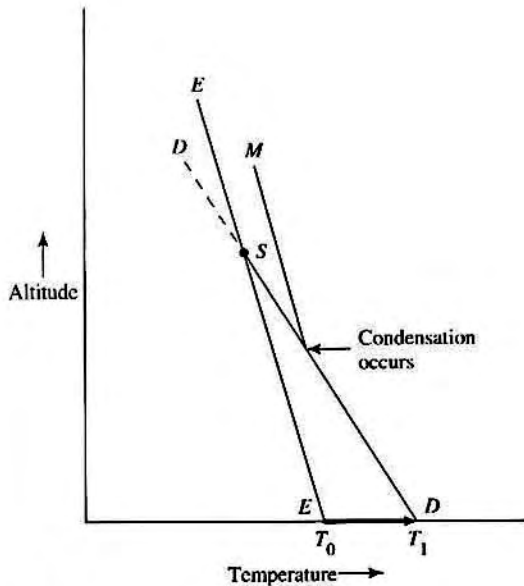


FIGURE 4-6

Tracks of 1966 tropical depressions, storms, and hurricanes and of Hurricane Beulah (1967) in the North Atlantic. Positions shown at 12-hr intervals. Reprinted with permission of Prentice Hall, Inc., from *Elements of Meteorology*, 4th ed., by Miller et al., © 1983 by Bell & Howell Co.

**FIGURE 4-7**

Convection due to surface heating. Line E is the environmental lapse rate (the actual atmospheric temperature as a function of elevation); line D is the dry adiabatic lapse rate; and line M is the moist adiabatic lapse rate. Air at the surface is heated from temperature T_0 to T_1 by contact with the ground. See text for full explanation.

**FIGURE 4-8**

Fair-weather cumulus clouds formed by thermal convection over Cuba. Note that the clouds are present only over the land—the heat capacity of the sea prevents it from heating sufficiently to induce convection. Photo by author.

4.1.4 Critical Temperature for Rain-Snow Transition

Much of the precipitation that falls outside the tropics originally forms as snow or hail in super-cooled clouds (Section D.5.3). The form of precipitation reaching the surface is determined largely by the height of the 0°C -temperature surface: Rain occurs if that surface is high enough to allow complete melting; otherwise the snow reaches the ground.

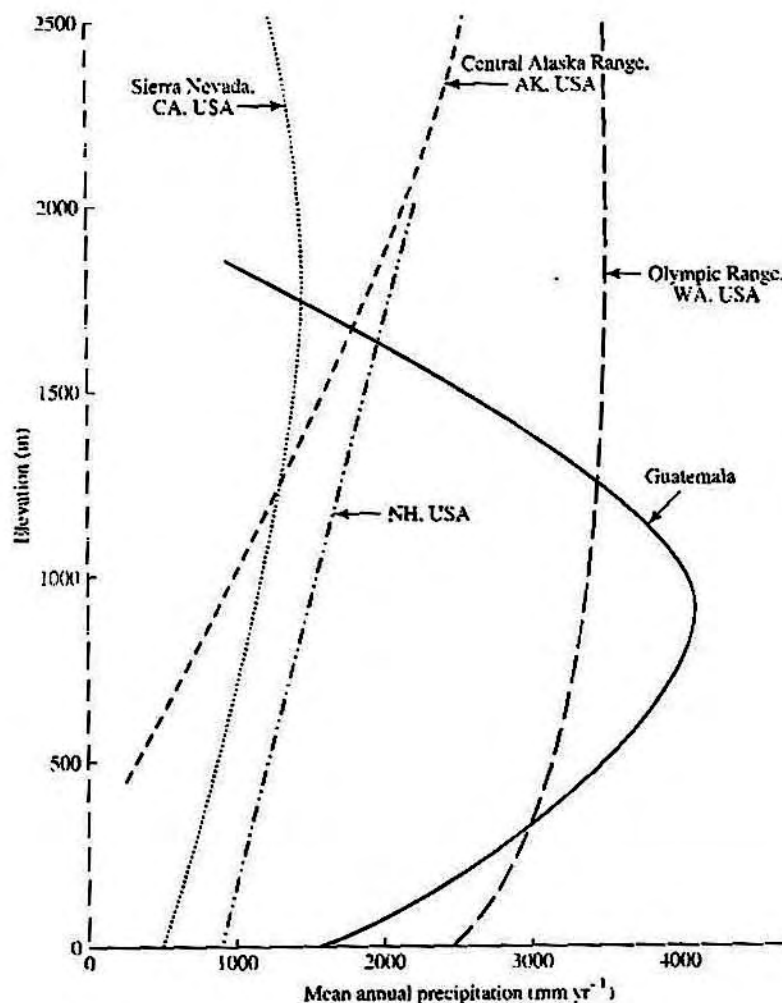
Input data for many watershed models consist of the amounts of precipitation falling in a given time period, along with other surface meteorological information for the period, such as temperature, humidity, wind speed, and solar radiation. Information about the form of the precipitation typically is not readily available; therefore one of the most critical decisions the model must make is whether precipitation is rain or snow. A mistake in this decision will result in an erroneous snowpack water balance and will cause the model either to predict a ground-water recharge and/or streamflow event when none occurred or to fail to predict an event that did occur.

The most natural criterion for this decision is the air temperature at the site of interest. We know that air temperature usually decreases with altitude (Section D.2.2) but, almost always, we have information only about surface temperature. Thus it is not immediately obvious what the temperature criterion should be, and we must turn to empirical information for guidance. Auer (1974) examined about 1000 weather observations in which the surface temperature was recorded and the solid or liquid nature of the precipitation was clearly indicated. (Occurrences of freezing rain were omitted.) He calculated the frequency of rain vs. snow reports at each temperature and constructed the graph shown in Figure 4-14, which indicates that rain is virtually never recorded when temperature is less than 0°C , snow is never observed when temperature exceeds 6°C , and the probabilities of rain and snow are equal when temperature equals 2.5°C .

It should be noted that the geographical applicability of Auer's (1974) analysis is not clear, and it may not apply in all regions. Thus, if at all possible, one should attempt to determine the relationship appropriate for a region of interest by examining local data. The method used in the BROOK90 model to separate rain and snow is described in Box 4-1.

FIGURE 4-9

Relations between mean annual precipitation and elevation in several regions of North America. Data from Barry and Chorley (1982) and Dingman et al. (1971; 1981).



4.1.5 Moisture Sources and Precipitation Recycling

For a given region, the fraction of precipitation that entered the atmosphere as evapotranspiration from that region is called the **recycling ratio**. Determining the value of this ratio is important in assessing the potential for human modification of climate. Prior to the 1940s, it was widely thought that most continental precipitation was recycled. This belief underlay several proposals for altering land-use practices in order to increase precipitation and thus prevent recurrences of conditions like the disastrous Dust Bowl of the early 1930s in the south-central U.S. However, a study by Holzman (1937) showed (1) that most of the precipitation falling on the United States comes from maritime air masses that obtain their moisture from the oceans and (2)

most of the evapotranspiration from the land enters continental air masses which move off the continent without yielding much precipitation; thus the recycling ratio is relatively low.

As is shown in Table 4-2, Holzman's (1937) conclusions have been supported in more detailed studies of the Mississippi River basin and Eurasia. However, two large environmentally sensitive regions have been found to have greater degrees of recycling: the Amazon Basin, where it is feared that the destruction of the rain forest will significantly affect at least the regional hydrologic cycle (Leah and Warrilow 1989), and the Sahel region of sub-Saharan Africa, where there are concerns that recycling produces positive feedbacks that will amplify the desertification processes initiated by over-intensive land use (Savenije 1995).



FIGURE 4-10

View westward toward the Rocky Mountains in Colorado, showing orographically-induced clouds persisting for many tens of kilometers leeward of the mountains before dissipating. Photo by author.

Precipitation recycling also appears to be at least seasonally important in other regions. Evaporation from large lakes, such as the Laurentian Great Lakes, gives rise to localized regions of high precipitation on their downwind shores, especially in late fall and early winter. Moisture evapotranspired from extensive wetlands during summer days may contribute a significant proportion of that season's precipitation in central Alaska.

The variations in estimates for the same regions and the limited geographical scope of the studies cited in Table 4-2 suggest that more research on recycling is needed to refine our knowledge of the likely impacts of land-use changes on climate, especially in arid regions where the recycling ratio may be high (Eltahir and Bras 1996). Modeling studies like that illustrated in Figure 3-38 are one approach to the problem; other approaches are used in the studies listed in Table 4-2.

4.2 MEASUREMENT AT A POINT

Because precipitation is the input to the land phase of the hydrologic cycle, its accurate measurement is the essential foundation for quantitative hydrologic analyses. Unfortunately, there are often many reasons for concern about the accuracy of precipitation

data, and these reasons must be understood and accounted for in both scientific and applied hydrologic analyses.

Almost always, hydrologists are concerned about the amount or rate of precipitation over an area such as a drainage basin, so there are two parts to the problem of accuracy: (1) How accurate are point measurements of precipitation? and (2) How accurately can point measurements be converted to measurements over an area? This section addresses the first question.

4.2.1 Types of Precipitation Gages

Conceptually, the measurement of precipitation at a point is straightforward: Simply place a vessel open to the air (a **rain gage** or **precipitation gage**) at the point of observation, and periodically measure or continuously record the quantity of water it collects. The volume collected is divided by the area of the opening and recorded as the depth of precipitation.²

Nonrecording **storage gages** may be simple straight-sided cylinders or more elaborate devices with funnels and collecting vessels. A **weighing-recording gage** works by introducing the collected water to a vessel on a scale and recording the accumulated weight. A second type of recording gage, known as a **tipping-bucket gage**, introduces the water to one of a pair of vessels with a known small capacity that are balanced on a fulcrum; when one vessel is filled, it tips and empties and records the time of this event, and the other vessel is brought into position for filling. Details of the construction and operation of several types of storage and recording gages were given by Gilman (1964), Brakensiek et al. (1979), and Shaw (1988).

The measurement of snow in precipitation gages is fraught with difficulties. The hydrologist is interested in the amount of water substance that falls, so the catch of frozen precipitation must be melted before measurement; this is often done by placing a charge of antifreeze in storage gages or by activating heating elements in tipping-bucket gages. Less tractable problems arise when snow piles up at the gage orifice and subsequently blows off and, as

² There is evidence that this approach has been in use at least since 400 B.C.E. in India, where the rainfall data were used in managing agriculture (Neff 1977).

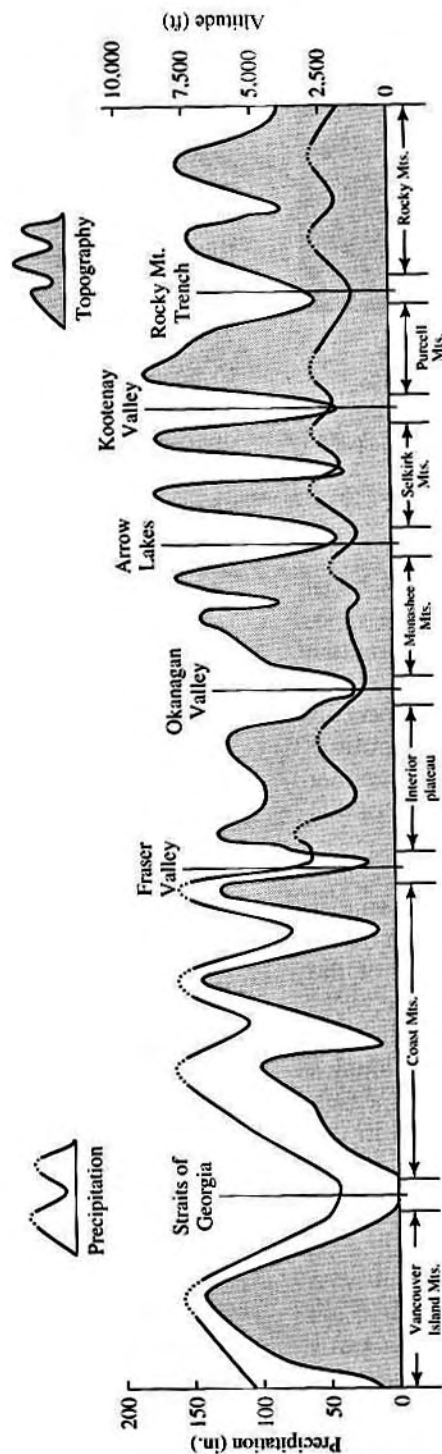


FIGURE 4-11
Relation between mean annual precipitation and topography across southern British Columbia. After Bruce and Clark (1966).