The Rocky Mountains encompass a wide range of climates, from the boreal setting of the Mackenzie River to warm temperate domains of the American Southwest. These climates set the context for ecological, hydrologic, and societal processes in this mountain region. In this chapter, we describe the geography and recent history (the past 100 years) of climate in the Rocky Mountains. We then explore the sensitivity of these climates to ongoing changes in regional and global factors that control climate throughout the Rockies.

This chapter discusses five Rocky Mountain regions divided along physiographic lines and political boundaries; two are in Canada and three are in the United States. The Canadian regions are the northern (Yukon and adjacent Northwest Territories) and southern (British Columbia and Alberta) Canadian Rockies (see the map facing the preface). The U.S. regions are distinguished physiographically as the northern, central, and southern U.S. Rockies (figure 4.1).

**Continental Context**

The Rocky Mountains have a north–south axis that spans the Northern Hemisphere's middle latitudes (roughly lat. 35°–65° N). In this location, the mountains form a significant 2,000–4,000 meter (m), or 7,000–14,000 feet (ft.), high barrier to the general westerly midlatitude flow of the atmosphere. The arrangement of mountains and atmospheric flow carrying moisture from the Pacific
FIGURE 4.1 *Northern, Central, and Southern Rocky Mountains of the United States.* Open circles mark locations of stations used in historical analyses; contour lines indicate elevation above sea level in meters (500 m contour interval).

Ocean generates a classic orographic precipitation pattern, with enhanced precipitation on the windward side and a rain shadow on the lee (figure 4.2).

Actual regional patterns are, of course, more complicated. Winter storms approaching the southern Canadian and northern U.S. Rockies from the Pacific Ocean are laden with moisture, whereas those arriving in the southern U.S. Rockies have lost much of their moisture in crossing the Sierra Nevada and the intermountain West (figure 4.2). On the eastern side of the Rockies, both polar continental cold air from boreal regions and warmer maritime tropical moist air from the Gulf of Mexico are blocked by the mountain front from Alberta to New Mexico. As these winter air masses collide with the mountain front, they move upslope and generate precipitation along the eastern ranges of the Rockies.
In summer, the southern Canadian Rockies continue to receive moist Pacific air. To the south, however, much of the interior western United States is under the influence of either dry continental air or monsoonal flows from the Gulf of Mexico and California. In the U.S. Rockies, the boundary between these air masses depends on incursion of gulf air into the southern Rockies driven by circulation from the Bermuda subtropical high-pressure center (figure 4.2) (Mitchell 1976).

An understanding of source regions for air masses that give rise to climates of the Rockies helps to identify causes of interannual, decadal, and longer-term variability. Regions receiving flows from the Pacific (the Canadian and northern U.S.
Rockies) are strongly influenced by variability in North Pacific Ocean surface temperatures and atmospheric circulation patterns, such as reflected by the strength of the Aleutian low-pressure center. In contrast, the southern U.S. Rockies are more influenced by subtropical and tropical sea surface temperatures and circulation patterns, such as El Niño and the Southern Oscillation (ENSO).

Geography of Climate

Climate is spatially highly variable in mountainous regions. Variability is determined largely by the effects of topography (elevation, slope, and aspect) on precipitation, temperature, solar radiation, and humidity. There are few records of temperature and precipitation in complex terrain compared with adjacent flatlands, and fewer still of solar radiation and humidity. It is usually necessary to infer topographic effects in order to quantify the climate of a region of complex terrain. This is accomplished by generating high-resolution maps of climate (e.g., grid interval less than 5 kilometers [km], or 3 miles [mi.]) with uniform data density over complex terrain using both horizontal interpolation of observations and vertical extrapolation to estimate climate at elevations higher than local observations (Thornton, Running, and White 1997; Daly, Neilson, and Phillips 1994). There are typically not enough observations of solar radiation and humidity to perform direct and accurate interpolations and extrapolations, but reasonable proxies can be derived from the more common observations of temperature and precipitation (Thornton, Hasenauer, and White 2000). Plates 3 and 4 present high-resolution climatologies for the U.S. and Canadian Rockies.

Climatology of the U.S. Rockies

We analyzed regional differences in the mean climatology of the U.S. Rocky Mountains based on a 1 km gridded eighteen-year database of daily surface weather parameters (the U.S. Daymet data set) (Thornton, Running, and White 1997); see plate 3 and table 4.1. Within each region, we also evaluated other climatological statistics, including interannual standard deviation and regressions with elevation (table 4.1).

Spatial Patterns. All regions show substantial spatial variation in mean climate reflecting roles of latitude, elevation, and continental position (plate 3). From south to north, the Rockies broadly become cooler (at the same elevation), receive more moisture (with the increasing influence of Pacific air), and have less solar radiation (table 4.1, plate 3). In the context of the entire western United States, the Rockies differ markedly from mountains to the west. The Coast Ranges, Cascade Range, and Sierra Nevada are more strongly influenced
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**Maximum Temperature (°C)**

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**Daily Total Solar Radiation (MJ m⁻² day⁻¹)**

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**Daily Average Water Vapor Pressure (Pa)**


*Note:* Analysis includes regional mean, interannual standard deviation (SD), interannual coefficient of variation (CV; as fraction, for total precipitation only), and elevation regressions slope and $R^2$ statistic. Elevation regressions were performed on climate parameters averaged over five elevation classes in each region.
by maritime air masses and are generally warmer and wetter and receive less solar radiation than the Rocky Mountains at similar latitudes.

The climatology of daily precipitation is complex because precipitation is not a temporally continuous process. The same long-term average of total annual precipitation could be obtained by either a high frequency of small events or a low frequency of large events. Precipitation frequency at low elevations is highest in the northern U.S. Rockies (35% wet days), reflecting the influence of relatively warm, moist air masses from the Pacific, compared with the central and southern regions (less than 12% wet days during winter). The average precipitation per event ranges substantially, from 1.2 centimeters (cm) per day, or approximately 0.5 inch (in.) per day, in winter at high elevations in the north to 0.25 cm/day (0.1 in./day) in winter at low elevations in the central Rockies.

Shortwave (solar) radiation received at the surface increases from north to south in winter but is nearly constant across regions in summer, when longer day length in the north compensates for lower sun angles. Winter solar radiation as a percentage of summer values ranges from 21% in the north to 36% in the south. Absolute humidity is highest in the northern region and lowest in the central region for all seasons, reflecting the maritime influence in the north. Humidity is slightly higher in the southern region than in the central region because of incursions of moist air from the Gulf of Mexico and the Pacific.

**Elevation Relationships.** Precipitation increases significantly with elevation in all three regions of the U.S. Rockies, with peak precipitation occurring near mountaintops, a pattern characteristic of midlatitude mountainous regions (table 4.1). The northern region has higher annual and January total precipitation for a given elevation than either the central or the southern region. In July, however, the southern U.S. Rockies receive more total precipitation at high elevations than the regions to the north because of the southwestern summer monsoon. Differences in annual totals among regions are greatest at low elevations.

The frequency at which it rains or snows decreases with elevation in the northern region but increases with elevation in the central and southern regions, a pattern that is stronger in January than in July. Higher elevations get more precipitation per event (defined as a wet day) in all regions and all seasons. In contrast to annual total precipitation and precipitation frequency, the largest differences in annual average event size among regions are at the highest elevations.

Temperatures can be colder at the lowest elevations in the central region than at similar elevations in the northern region. This is because of stronger influence from continental air masses in the central Rockies and maritime air
masses in the north. Relationships with elevation for both maximum and minimum temperature (i.e., temperature lapse rates) are strong and linear in all regions and all seasons (table 4.1). Lapse rates for maximum and minimum temperature are in general stronger (larger decrease in temperature with increasing elevation) in summer than in winter, with summer lapse rates for maximum temperature stronger than for minimum temperature. Lapse rates for all seasons are stronger in the southern region than in the northern and central regions. As a result, there is a convergence in temperatures across regions at higher elevation. For all regions, interannual variability of maximum temperature decreases with elevation in winter, whereas that for minimum temperature increases in summer.

Winter solar radiation increases with elevation in the northern U.S. Rockies, but in the southern region it peaks at mid-elevations and then decreases slightly with elevation. This regional difference is in part a result of lower sun angles in winter in the north, which cause lower and mid-elevation sites to be shaded by adjacent mountains. Another contributing factor is the tendency for low-level clouds and fog to form as a result of temperature inversions during winter in valleys of the northern and central regions. In all three regions, peak summer solar radiation occurs at mid-elevations, around 2,500 m (8,000 ft.). Interannual variability of solar radiation increases strongly with elevation in all regions in winter and in the northern and southern regions in summer.

Water vapor pressure (absolute humidity) decreases with elevation in all regions and all seasons, consistent with decreasing temperature and surface air pressure. Decreases in absolute humidity with elevation are largest in summer, whereas those for relative humidity are largest in winter. Interannual variability in absolute humidity decreases with elevation in all regions in winter.

**Climatology of the Canadian Rockies**

A high-resolution gridded analysis for western Canada reveals the distribution of mean annual precipitation and mean January temperature (plate 4; Daly et al. 2001, 2002). These maps show the same pattern as in the United States of wetter and warmer climate in coastal ranges than in the Canadian Rocky Mountains. The southern region is wetter and warmer than the northern Canadian Rockies. This precipitation gradient is the reverse of that in the U.S. Rockies and is caused by the stronger presence of Pacific maritime air (and higher elevations) in the southern Canadian Rockies. The influence of the wintertime arctic front (southern limit of arctic continental air) is clearly defined as a strong gradient in mean January maximum temperature along the border between British Columbia and Yukon Territory (plate 4, right panel).
PLATE 4  High-Resolution Mean Climatology of Western Canada, 1961–1990. Left panel: Mean annual precipitation (cm). Right panel: January mean maximum temperature (°C). Spatial resolution is 2.5 minutes of latitude (4 km). Spatial analysis based on Daly et al. 2001, 2002; maps provided by Spatial Climate Analysis Service and used by permission.
Recent Climate History

We evaluated centennial trends and interannual variability in annual and seasonal precipitation, mean minimum and maximum temperature, and mean diurnal temperature range for the U.S. and Canadian Rockies. Trends in minimum and maximum temperatures reflect different climatic processes and have different effects on ecological dynamics and surface hydrology, so they are considered separately. Diurnal temperature range is a useful index that summarizes their different dynamics.

Historical Climate Trends

We evaluated centennial climate trends for the U.S. Rockies on the basis of longest-running station records selected from the United States Historical Climatology Network (HCN) data set (Kittel and Royle, unpublished analysis). The HCN monthly precipitation and mean minimum and maximum temperature data set includes stations that are long-term with relatively complete records that were corrected for time-of-observation differences, instrument changes, instrument moves, station relocations, and urbanization effects (Quinlan, Karl, and Williams 1987). Although there is a natural bias in longer records toward lower elevations (because early settlement was in mountain valleys and foothills), the stations span a wide range of elevations in each of the three U.S. regions (figure 4.1).

Annual and seasonal precipitation and mean minimum temperature increased significantly during the twentieth century. The magnitude and statistical significance of these trends varied by region and season. The strongest precipitation trends occurred in the northern and central U.S. Rockies in summer: 30% and 33% per century, respectively ($P < 0.05$) (figure 4.3a, b). Trends in annual mean minimum temperature were +0.7°C and +0.9°C per century, respectively ($P < 0.005$), for these regions. Annual and seasonal trends were not significant in the southern U.S. Rockies for all variables (figure 4.3c, f).

Mean diurnal temperature range decreased significantly. Trends were strongest in the northern and central U.S. Rockies (−0.6°C and −1.2°C/century, respectively, for summer; figure 4.3d, e). The decrease in mean diurnal temperature range came primarily from increasing minimum temperatures rather than from changes in maximum temperature. This pattern of decreasing diurnal range is consistent with trends observed throughout the United States, which are attributed to increased cloud cover. Greater cloud cover increases nighttime temperatures and decreases daytime temperatures (or at least causes smaller increases if there is a background increase in surface air temperatures) (Plantico et al. 1990; Dai, Trenberth, and Karl 1999). Observed increases in precipitation (with concurrent increases in cloud cover) are consistent with this explanation (figure 4.3a, b).
FIGURE 4.3 Climate Variability and Trends in the Northern, Central, and Southern U.S. Rockies, 1895–1993. Top panels, northern Rockies; middle panels, central Rockies; bottom panels, southern Rockies. (a–c): Summer total precipitation anomalies (cm). (d–f): Summer mean diurnal temperature range anomalies (°C). Regional anomalies are relative to record regional means and are based on simple averages of station values. Plots show yearly values (thin line), seven-year running mean (thicker solid line), and record trend line (dashed line). Summer is June–August. Where a station record was discontinuous or incomplete, we took advantage of local spatial autocorrelation structure to predict monthly values to create temporally complete climate series for the period 1895–1993 (Kittel et al. 1997).
In the southern Canadian Rockies, annual and seasonal mean minimum temperature increased significantly from 1900 to 1998 (+1.0°C to +2.5°C/century), and diurnal temperature range decreased (−0.5°C to −2.0°C/century; Zhang et al. 2000). For midlatitude Canada in general, the negative trend in diurnal temperature range was strongly correlated with increasing cloud cover, precipitation, and streamflow for this same period (Dai, Trenberth, and Karl 1999). In the southern Canadian Rockies, precipitation increased both annually and seasonally by +5% to +40% since 1900 (Zhang et al. 2000).

There were significant and comparable increases in maximum and minimum temperature (trends in annual means +1.5°C to +2.0°C per 50 years [yr.]) from 1950 to 1998 for both the southern and northern Canadian Rockies, with generally larger changes in the north (Zhang et al. 2000). This continues the positive poleward gradient in trends we found in the U.S. Rockies, with temperature changes for the second half of the twentieth century in the northern Canadian Rockies on the order of four times those farther south in the U.S. Rockies (Nicholls et al. 1996).

**Elevation Dependence in Trends**

The noted ability of mountains to “generate their own climate” comes from the strong modification of circulation, precipitation, and solar radiation regimes by mountain massifs. Although this is evident in long-term climatologies, how these processes affect climate variability and trends in mountains is not as well understood. Previous studies in the Rocky Mountains and elsewhere have pointed out significant differences in century-scale trends and interannual variability between mountain environments and adjacent lowlands (Greenland and Kittel 2002; Inouye et al. 2000; Beniston, Diaz, and Bradley 1997; Diaz and Bradley 1997). These differences can reflect an amplification of lower-elevation signals with elevation (Beniston and Rebetez 1996) or can reflect divergent signals, suggesting that lowland and highland climates are decoupled (Diaz and Bradley 1997, Greenland and Losleben 2001).

There are several explanations for an elevation effect. Higher reaches of mountain systems are in closer contact with the free troposphere and so respond rapidly to upper-air circulation changes. Higher mountains may also be less affected by ameliorating surface processes (e.g., irrigation, large lake effects, urbanization) (Greenland and Losleben 2001, Giorgi et al. 1997). In addition, upper air can override lower air (especially in the presence of inversions), decoupling high-elevation climates from convective mixing of the lower troposphere and development of valley fogs (Greenland and Losleben 2001, Diaz and Bradley 1997). A feedback between snow cover and surface albedo (fraction of solar radiation reflected at the surface) can also amplify a regional temperature signal with ele-
vation (Giorgi et al. 1997). Regional higher temperatures result in less snow cover, more absorption of solar radiation, and therefore greater local warming, enhancing the regional signal. The opposite is true in colder years.

Dependence of long-term temperature changes on elevation results in changes in lapse rate with consequences for atmospheric stability and generation of local circulation and precipitation. We found that changes in lapse rate varied by region and season. In the northern U.S. Rockies, daytime warming trends tended to be stronger with elevation, so lapse rates were generally reduced, resulting in a more stable atmosphere during the day. We found little elevation dependence in nighttime temperature trends or a weakening of these trends with elevation. These results contrast with those of Diaz and Bradley (1997), who found stronger warming at higher elevations due largely to increases in minimum temperature for midlatitude Northern Hemisphere mountains. However, their result depended on what elevation span was used and what time period was assessed (Diaz and Bradley 1997, Pepin 2000).

**Interannual and Interdecadal Climate Variability**

There is strong variation in monthly precipitation, monthly mean minimum and maximum temperature, and monthly mean diurnal temperature range at interannual and decadal scales (figure 4.3). Some of the variability is shared across the three regions, and some is distinct. Adjacent regions were more similar to one another than those farther apart. Correlation between adjacent regions in precipitation variability diminishes at time scales longer than five to ten years but is generally high for temperature at both shorter and longer time scales.

There were strong droughts in the late 1910s and early 1920s in the northern and central U.S. Rockies. The 1930s drought was strongest in the north, and the 1950s drought was most pronounced in the central and southern regions (figure 4.3a–c). Late-twentieth-century precipitation extremes associated with the midcontinental 1988 drought and 1993 floods extended into the northern and central U.S. Rockies.

Temperature minima and maxima had a general pattern of cooling until 1915, warming through the late 1930s, cooling or little trend into the 1970s, and then warming. This pattern agrees with the overall decadal pattern for the conterminous United States (Karl et al. 1996) and is reflected in the twentieth-century record for southern Canada (Zhang et al. 2000). The signal is related to shifts in Northern Hemisphere circulation patterns (Wallace, Zhang, and Bajuk 1996) and is possibly related to combined effects of solar output variation, volcanic eruptions, and enhanced greenhouse gas forcing (Houghton et al. 2001).

Although there is a hemispheric component of decadal variability in the Rocky Mountain signal, there is also a regional, elevation-dependent element.
There was a series of cold years in the early 1980s for high-elevation stations (higher than 3,000 m, or 9,800 ft.) in Colorado and Wyoming, but not for lower-elevation stations (Losleben 1997). We saw the same pattern in regional annual mean maximum temperature for all three regions of the U.S. Rockies. Although this is in some ways a local signal, isolated to higher elevations, regional consistency suggests broader-scale processes acting on the Rocky Mountains. The cold anomaly was related to extreme, near-synchronous events: particulates from volcanic eruptions reduced solar inputs, and a strong El Niño and changes in North Pacific atmospheric circulation patterns caused greater high-elevation increases in precipitation and cloud cover (Greenland and Losleben 2001).

Diurnal temperature range in the northern and central U.S. Rockies generally decreased until around 1910, increased through the mid-1930s, and decreased markedly since then (e.g., for summer; figure 4.3d, e). The midcentury shift from weakly increasing to strongly decreasing trends is similar to that for the conterminous United States as a whole, with some variation in timing of this break (Plantico et al. 1990). The shift in trends is in contrast to that in the southern Canadian Rockies, where there was a strong decline prior to the 1950s and weak trends since (Zhang et al. 2000). Increasing total cloud amounts in the Canadian mid-latitudes in the first half of the twentieth century are the probable cause (Zhang et al. 2000).

**Hemispheric and Global Teleconnections**

Interannual climate variability in the U.S. Rockies is linked to global and hemispheric circulation patterns. An analysis of such teleconnections (long-distance climate links) showed that a significant portion of interannual variation in temperature and precipitation can be explained by (1) the intensity of the Aleutian low-pressure center, which is inversely represented by the North Pacific Index (NPI) (Trenberth and Hurrell 1994), and (2) reversals in eastern Pacific sea surface temperature anomalies (with warm anomalies, or deviations from the average, identified as El Niño) and tropical Pacific atmospheric pressure anomalies (reflecting the Southern Oscillation) (Kittel and Royle, unpublished analysis).

Winter temperatures are negatively correlated with the NPI throughout the U.S. Rockies, with the relationship strongest in the central and northern regions ($P < 0.005$). A negative correlation indicates that warm winter temperatures occur when the Aleutian Low is most intense (deep low-pressure center) so that flow around the Low is more from the south, bringing warm air northward into the Rockies.

Winter precipitation is significantly positively correlated with the NPI in the central U.S. Rockies. This suggests that when the Aleutian Low is weaker, the
resulting westerly flow brings Pacific storms directly into the central Rockies. Winter precipitation and snow cover are also positively correlated with the occurrence of El Niño in the southern Rockies and are anticorrelated in the northern and central U.S. Rockies and the southern Canadian Rockies (Kittel and Royle, unpublished analysis; Groisman et al. 1994).

We found that summer precipitation in the northern U.S. Rockies was negatively correlated with the NPI and therefore positively linked to a deeper Aleutian Low. A stronger Low in summer directs storm tracks farther south than normal. Summer precipitation in the southern and central U.S. Rockies was not correlated with the NPI but was, on the other hand, strongly positively correlated with El Niño (Kittel and Royle, unpublished analysis; Dai and Wigley 2000).

Groisman et al. (1994) showed elevation dependence in the sign of snowpack response to El Niño across the Northern Hemisphere, including the Canadian and U.S. Rockies, supported by a regional analysis for the southern U.S. Rockies (M. Losleben, pers. comm.). Elevation-dependent teleconnections are found elsewhere, such as the Swiss Alps, where winter temperature responses to hemispheric circulation patterns are strongest at higher elevations (Giorgi et al. 1997).

**Sensitivity of Mountain Climates to Altered Forcing**

Terrestrial environments are experiencing a range of anthropogenic forcings that affect their current and future status. Some of these forcings act on climate; some are global in scope, and others are regionally generated. We consider two that directly influence climates of the Rocky Mountains: (1) changing character of the land surface at both regional and global scales and (2) changing composition of the atmosphere, especially with respect to radiatively active (greenhouse) gases. The potential effect of the resulting changes in climate will influence the function and structure of ecological and surface hydrologic systems and the human economies that depend on them.

**Changing Land Use—Regional to Global Forcing**

Although the Rockies deprive adjacent regions of precipitation by blocking and capturing moisture from maritime air masses, the cordillera is at the same time the source of great river systems on the Arctic, Pacific, and Gulf of Mexico sides of the Continental Divide—the Saskatchewan, Columbia, Peace-Slave-Athabaska, Colorado, Missouri, Arkansas, and Rio Grande, among others. Many of these rivers give rise to ribbons of moisture well out into the semi-arid Great Plains and the arid West, supplying irrigation flows throughout these regions. Development of irrigation systems and other land use changes in the Great
Plains have influenced local and regional climate (Pielke and Avissar 1990, Segal et al. 1998). Land use change in the Great Plains and in regions well removed (e.g., the Tropics) very likely has altered, and potentially can alter, climates of the Rocky Mountains.

Changing land cover has been demonstrated to affect regional and global climate significantly in modeling and observational studies at a variety of spatial scales (e.g., Betts et al. 1996). A change in land cover can have profound effects on the total energy absorbed by the land surface by altering surface albedo and on how that absorbed energy is returned to the atmosphere via latent and sensible heat fluxes. For example, energy in a densely vegetated region is returned to the atmosphere mostly by means of transpiration and evaporation (i.e., latent fluxes). This limits the direct (sensible) heating of the air, causing cooler, moister conditions. The opposite occurs in sparsely vegetated regions, where sensible heating dominates. Land cover–mediated changes in the surface energy budget affect surface hydrology, storm formation, regional circulations, and other quantities of meteorological and ecological interest.

Climate model simulations are used (almost exclusively) to address the influence of land cover changes on regional climate. Modeling studies of the effects of observed land cover changes across Canada and the conterminous United States found winter warming in the northern Canadian Rockies of as much as 2°C, significant springtime warming of more than 1°C and little July temperature response in the U.S. Rockies, and a significant increase in July precipitation in the southern and central U.S. Rockies (Copeland, Pielke, and Kittel 1996; Bonan 1997).

A high-resolution modeling study of the effect of change in agricultural land cover adjacent to the Colorado Front Range found that irrigated lands had a significant effect on the mountain-plains breeze circulation in summer and therefore affected weather patterns in the mountain regions as well as on the plains (Chase et al. 1999). The study suggested that land use change has resulted in reduced daytime upslope winds and cooling in the mountains in summer. Changes in summer upslope circulation patterns, as suggested by these simulations, would also significantly affect advection of pollution from nearby urban areas (e.g., Denver), altering nitrogen deposition rates in the Rocky Mountain alpine zone (chapter 9; Sievering 2001).

The potential effect of land cover change on the Rocky Mountains is not limited to local land cover changes. Climate model studies indicate that land cover changes across the globe, particularly in the Tropics, affect global-scale circulations and have the potential to affect areas far removed from the source of change (Chase et al. 1996, 2000; Zhao, Pitman, and Chase 2001). The exact nature of the effects is a function of model configuration and interacting envi-
ronmental influences and is therefore uncertain. However, the suggested climatic changes are of similar magnitude to those simulated for elevated greenhouse gases and therefore represent potentially significant factors in regional climate change (Chase et al. 2001, Pitman and Zhao 2000). Land cover–climate model simulations are incomplete for many reasons, but they do serve to illustrate processes and interactions at work. They emphasize that the atmosphere communicates environmental changes over large distances. No compilation of potential regional effects can focus solely on local and regional processes.

Changing Atmospheric Composition
Greenhouse gas concentrations and tropospheric aerosols (very fine suspended particles) have been increasing since the mid-nineteenth century, largely as a result of human activities. These increases have raised the concern that they have altered and will continue to alter surface climates (Houghton et al. 2001). We address the potential magnitude of the effects of rising greenhouse gases and aerosols on Rocky Mountain climates and how mountain regional processes might modify these global- to continental-scale forced changes. The effects must be considered in the context of other sources of climatic change, including land use change, natural internal oscillations in atmosphere-ocean-cryosphere (sea and land ice) dynamics, variation in solar output, and volcanic eruptions. Reports by the Intergovernmental Panel on Climate Change (Houghton et al. 2001, McCarthy et al. 2001) and the U.S. National Assessment Synthesis Team (NAST 2001, Wagner and Baron 1999) review scientific issues and assessments of these effects and their potential consequences for environmental and human systems.

For the U.S. Rockies, we evaluated climate sensitivity of coupled global climate model (GCM) experiments from (1) the Canadian Centre for Climate Modelling and Analysis and (2) the Hadley Centre for Climate Prediction and Research, United Kingdom, run under future emission-forcing scenarios of approximately 1%/yr. increases in greenhouse gases and sulfate aerosols through the twenty-first century (Boer, Flato, and Ramsden 2000; Mitchell et al. 1995). These simulations, referred to here as CCC and HAD, respectively, have a coarse resolution (grid intervals approximately 200–375 km, or 120–225 mi.), which was rescaled to a 0.5° latitude/longitude grid for the conterminous United States (Kittel et al. 2000). We looked at annual and seasonal changes in precipitation, mean minimum and maximum temperature, and two integrative variables: (1) Palmer Drought Severity Index (PDSI) and (2) length of the snow-free season (Kittel et al. 2000, Kittel and Thornton, unpublished analysis).

There was greater temperature sensitivity in the CCC experiment than in the HAD scenario (+5.6°C vs. +3.6°C/century for mean temperature) and mini-
mally higher precipitation sensitivity (+20% vs. +18%/century; e.g., figure 4.4a) in response to rising greenhouse gases and sulfate aerosols. These responses were larger than observed historical trends and varied by season. Both experiments showed generally greater changes in minimum temperature than in maximum temperature, resulting in a decrease in diurnal temperature range for most seasons (e.g., figure 4.4b; plate 5a, b).

Changes in PDSI were small in HAD compared with those in CCC (figure 4.4c; plate 5c, d). This was because the drying effect of increased temperatures (higher potential evapotranspiration) was roughly countered by increases in precipitation, whereas in CCC higher temperature sensitivity overwhelmed increased moisture inputs. The snow-free period increased in both scenarios, more strongly in CCC (by one or more months; see plate 5b). This change was largely in response to higher spring and early summer temperatures but possibly also because of less winter precipitation falling as snow in spite of strong increases in winter precipitation (30%–65%/century).

The CCC experiment showed temperature increases from 1971–1990 to the 2030s of 1°C–2°C in the southern Canadian Rockies, increasing to +2°C–3°C poleward in the northern Canadian Rockies (Boer, Flato, and Ramsden 2000). Modeled trends in precipitation in the Canadian Rockies were small and mostly positive; the signal was weaker than that simulated farther south in the U.S. Rockies (figure 4.4a).

Elevation dependence in climate change response to increasing greenhouse gases has been little explored but is suggested by elevation dependence in historical climatic trends, interannual variability, and teleconnections and by simulated dynamic interactions between mountain and adjacent lowland climates. In the CCC experiment, surface temperature response intensified with elevation over the Rockies (Fyfe and Flato 1999). The effect was restricted to winter and spring and was related to the snow-albedo feedback, wherein reduction of snowpack at mid- and higher elevations due to higher temperatures under elevated greenhouse gases resulted in greater absorption of solar radiation, further elevating surface air temperatures. Finer-resolution model simulations for the Swiss Alps found similar results and also showed more precipitation at higher elevations from winter through summer (Giorgi et al. 1997). With respect to hydrologic effects, changes in precipitation timing and type (e.g., snow vs. rain) and snowmelt timing strongly altered seasonal runoff patterns with elevation.

Although CCC results demonstrate elevation dependence in climate sensitivity at the broad scale across the Rocky Mountains, there is also a strong locally forced component that comes through in finer-resolution PDSI and snowpack dynamics (plate 5c–f). Responses in PDSI and snow-free period to the broad-scale GCM
PLATE 5  Sensitivity of U.S. Rocky Mountain Climates to Transient Greenhouse Gases and Sulfate Aerosol Forcing. Values are as simulated by two coupled global climate models (HAD and CCC; see text) for the ten-year period 2025–2034 relative to a 1961–1990 baseline. (a, b) Changes in winter mean diurnal temperature range (monthly mean maximum temperature–monthly mean minimum temperature, °C). (c, d) Changes in summer Palmer Drought Severity Index (PDSI). (e, f) Changes in snow-free period (months). Elevation is contoured as plotted in figure 4.1 (500 m interval). Winter is December–February; summer is June–August. Snow-free season was calculated with the terrestrial ecosystem model Biome-BGC (Running and Coughlan 1988) using daily precipitation, temperature, radiation, and humidity (from Kittel et al. 2000) to estimate snow accumulation and melt. The average number of snow-free months was calculated for each 0.5° grid cell for the baseline period (1961–1990) and for HAD and CCC future climate scenarios.
FIGURE 4.4 Sensitivity of U.S. Rocky Mountain Climates to Transient Greenhouse Gases and Sulfate Aerosol Forcing. Values for 1994–2100 are based on two coupled global climate model experiments (HAD and CCC; see text). Historical observed record (1895–1993) is also shown for reference. (a) Annual total precipitation (mm/yr.), northern U.S. Rockies. (b) Winter mean diurnal temperature range (°C), central U.S. Rockies. (c) Summer mean Palmer Drought Severity Index (PDSI), southern U.S. Rockies. Climate series are plotted as a thin solid line for historical period and HAD future climate scenario and a thick solid line for CCC scenario. Horizontal line is the 1960–1991 baseline climate mean. Regression (long-dash) lines are plotted if significant ($P < 0.05$); slopes for annual precipitation in (a) were $+11\%$ and $+25\%$/century for HAD and CCC scenarios, respectively; for winter mean diurnal temperature range in (b), $-2.0$ °C and $-5.7$°C/century for HAD and CCC; and for summer mean PDSI in (c), $-4.1$ units/century for CCC.
forcing are patchy by comparison, with much of this pattern correlated with finer-grid (0.5° lat.) elevations. Correlations were similar under the two climate scenarios but differed by region. Under CCC forcing, negative changes in PDSI were intensified at higher elevations in the northern U.S. Rockies but diminished with elevation in the central and southern regions (plate 5c). These patterns come from nonlinear relationships among elevation, climate, and snowpack, soil moisture dynamics, and interactions with vegetation and soil distribution. These results suggest that, as in the historical record, future climate sensitivity of the three U.S. Rocky Mountain regions will have distinct responses and that fine-scale processes will operate to enhance or diminish larger-scale changes within each of the regions.

Uncertainties and Approaches in Assessing Climate Sensitivity
Climate model results must be viewed with caution, especially when applied to mountainous terrain. They are limited by our understanding and model representation of key processes and forcings (Wigley and Raper 2001). A major limitation of GCMs is their coarse model resolution, which cannot adequately address regional processes such as convection or sufficiently resolve topography. Poor representation of land surface features and processes results in a high level of uncertainty associated with regional scenarios of climate change. In spite of these limitations, GCM experiments reveal general sensitivity of climate and potential magnitude of change. Regional climate models also have their limitations but are better suited to evaluate the sensitivity of mountain climates across highly heterogeneous terrain (Giorgi and Mearns 1991).

Although climate model results for the U.S. and Canadian Rocky Mountains do not constitute forecasts or even realistic projections of climate change that can be used as a basis for policy or land management decisions, they are a foundation for understanding potential magnitude for climatic shifts and for identifying potential vulnerability of social and environmental processes to anthropogenic climate change. Climate model output can be used to identify potential vulnerability of social and environmental processes to anthropogenic climate change. In spite of high uncertainty about the exact nature of possible future climate outcomes, regional and global climate modeling sensitivity studies serve as one component in the development of “least regrets” strategies for reducing vulnerability of human systems to rapid climate change, independent of any particular climate change scenario and in the context of multiple stressors (Sulzman, Poiani, and Kittel 1995; Pielke 2001).

Once and Future Climates
The Rocky Mountains are home to a wide variety of climates. Across their span, these climates exhibit behavior at interannual through centennial scales that
responds to hemispheric and global forcing but also is distinguished by region and elevation. These relationships carry through to our current insights as to the sensitivity of Rocky Mountain climates to land use change and altered radiative forcing from increasing levels of greenhouse gases and aerosols. Although we cannot foretell what future climate changes will bring to the region, we have a clear understanding that the Rocky Mountains will experience change in climates, ecosystems, and hydrology that reflects not only global forcing but also a substantial reworking of this forcing by the mountain system itself.

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