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2 Current Alpine Climate

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2.1 INTRODUCTION

In the past, the Alpine climate has exhibited large variations on a wide range of timescales. On the 100,000 year timescale, the advance and retreat of Alpine glaciers that accompanied the onset and decay of the Ice Ages left indelible marks on the landscape. On century and multidecadal timescales, significant climate variations have served, for example, to modify agricultural practice. Again, shorter-term climate variations have also had notable impacts. For instance "the year without a summer," 1816, resulted in a severe famine in the Alpine region. These climate signals experienced in the Alpine region were the local manifestation of three forms of natural changes in the planet's climate system linked respectively to (1) variations in the earth's orbit around the sun, (2) the redistribution of energy between the atmosphere and ocean, and (3) volcanic eruptions that substantially but transiently changed the atmosphere's aerosol composition.

Anthropogenic effects also contribute to global climate change. Incomplete observations and inadequate understanding of the climate system currently limit estimates of their impact, but their amplitude could rival that of natural variations. Recent assessments suggest that the increasing atmospheric concentration of specific long-lived trace gases (carbon dioxide, methane, nitrous oxide, and chlorofluorocarbons), although ameliorated somewhat by the countereffect of anthropogenic sulphate aerosols, could result in an unprecedentedly rapid (on the order of 0.3 K per decade) global mean temperature change. Such a change in the global mean would be expected to exhibit significant regional variations with accompanying ecological and socioeconomic effects.

This study focuses on aspects of Alpine climate and climate change. The Alps are a major mountain range that exerts a strong influence upon the in situ weather and climate. Their height is such that from the foothills to the Alpine crest the mean temperature decreases by typically some 15–20 K, with accompanying large spatial gradients in the soil type and cover (e.g., glaciers and snow cover) and in the types of ecosystems. The mountain range's scale and geometry enables it both to modify and to trigger weather systems and thereby to establish distinct climatic characteristics. Again the Alps demark the boundary between two major climate zones: the mid-latitude temperate and the Mediterranean type. The topographic factors also influence the nature of the Alpine response to global climate change and indeed suggest that the potential for complex (and significant) climate variations is enhanced in this region. Moreover, the resulting impacts could be substantial, since the population distribution, the broad range of ecosystems, agricultural activities and tourism are all shaped by, and adapted to, the physico-climatic topography of the region.

The foregoing remarks emphasize the need to consolidate our understanding of Alpine climate, to examine the nature and amplitude of possible climate change in the region, and to assess the form and extent of the impacts that could ensue from such a change. The undertaking of such a program is constrained, however, by the climate system's complexity and nonlinearity. The complexity is reflected in the myriad of interacting physical, chemical, and biological processes that render unattainable both precise specification and comprehensive understanding and explicit representation of the system. Likewise, the nonlinearity betokens an intrinsic limit to the system's predictability. These limitations to our observational base, scientific understanding, and ability to model, together with the inevitable uncertainty attached to any forecast of the climate's evolution, serve both to color the approach to and qualify the results of climate and climate change studies.

The present chapter is predominantly a synthesis of the extant Alpine climate studies, and our interrelated objectives are to describe the current Alpine climate, devoting particular attention to the climatic elements' geographical distribution and variability, and to outline the nature and physics of the processes that contribute decisively to establishing the particular features of the Alpine climate.

2.2 ALPINE CLIMATOLOGY: WHAT ARE THE GEOGRAPHICAL DISTRIBUTION AND VARIABILITY OF THE CLIMATIC ELEMENTS?

The Alps are an 800-kilometer, arc-shaped mountain range with a mean width of approximately 200 kilometers and an average ridge height of about 2.5 kilometers. Additional distinctive features of the range (see figure 2.1) are the major valleys that run predominantly north or south onto the foreland and several east-west inner-alpine valleys aligned along the main ridge.

The severity of Alpine weather and the occurrence of several distinctive orographically related atmospheric flow phenomena (e.g., the wind systems of the Bise, Bora, Föhn, and Mistral; Alpine lee cyclogenesis events; and orographic precipitation enhancement) has long attracted the interest of natural scientists and stimulated scientific investigation. Already in the eighteenth century Horace Bénédict de Saussure, with his combination of instrument design, field measurements, and physical considerations, laid the foundations for the discipline of mountain meteorology. This early interest has three sig-

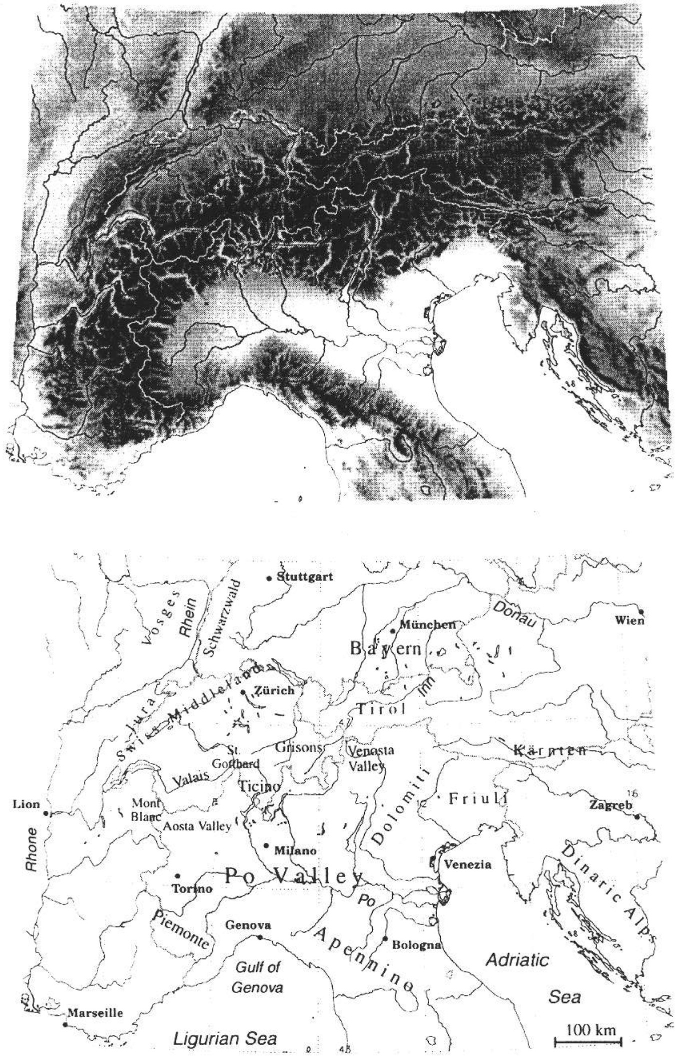

Figure 2.1 Topography and geography of the Alpine region. Included are the areal terms used in the text. (Courtesy of Dr. Daniel Lüthi, Atmospheric Science ETH.)

nificant repercussions for the present study. First, it led to the early establishment of meteorological observational stations, including several at high elevations. For example, of the principal mountain observatories Barry (1992) lists, almost one-third are Alpine stations. Observations at these stations represent an invaluable climate resource in the form of long time series for various climatic elements. Second, the systematic study of these data has provided the rudiments for an Alpine statistical climatology (for example, detailed documentation and cartographic displays of mean state climatic elements and their associated variance and trends), and the study of the spatial organization, dynamical characteristics and associated weather elements of the various Alpine-related flow systems forms the basis for a synoptic-dynamic climatology of the region (for example, the development and compilation of refined classifications of the prevailing weather types). Third, the early Alpine glaciologists provided both the first observational evidence for climate change in the form of ice ages (Agassiz 1840), and outlined the linkage of climate and its variation to the abundance of atmospheric greenhouse gases (Tyndall 1863).

In the present section we draw on and synthesize the available climatological data and the associated studies to provide an overview of Alpine climate in terms of the distribution of the climatic elements. In doing so, we restrict attention mostly to the past 100 years, for which modern instrumental records are available. For information on longer-term variability and paleoclimatic proxy data, the reader is referred to chapter 3.

2.2.1 Temperature

Our current knowledge of the climatic distribution of the air temperature is based on regular observations undertaken at the networks of surface stations and upper-air balloon soundings operated in the Alpine countries. In addition, special field campaigns during which observations were conducted at higher spatial and temporal resolution yielded information on the thermal conditions that prevail on scales ranging from that of the entire ridge down to that of specific valleys and slopes. In the following, we discuss the picture of the distribution that emerges from the analyses of these observations and proceed from the overall spatial distribution, including its seasonal and diurnal patterns, to the interannual variations since the beginning of regular observations.

Maps of the yearly and seasonal mean surface air temperature are available for individual Alpine countries and provide much local detail. An Alpine-wide overview with a coarser resolution is included in the Atlas of the World Meteorological Organization (WMO 1970). These maps' salient feature is a general temperature decrease with elevation, typically 0.65 K per 100 meters. This feature is a characteristic of the troposphere and is not directly related to specific topographic effects. Some of the topography-related flow systems, however, are linked to pronounced spatial, seasonal

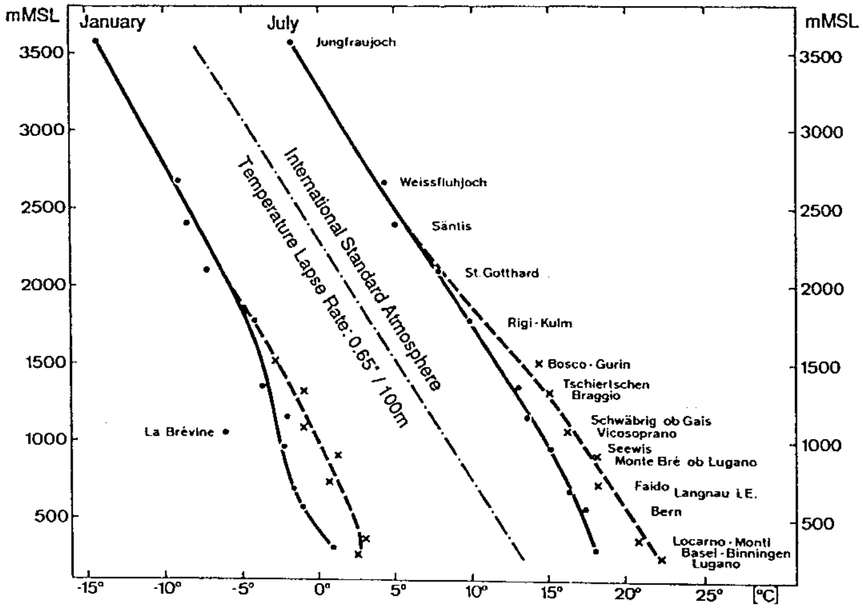

Figure 2.2 Vertical profiles of mean January and July temperature as given by surface observations in the Swiss Alps. Dots and crosses signify stations located to the north and south, respetively, of the main ridge crest. (After Schüepp et al. 1978; figure taken from Wanner 1991.)

and diurnal variations of the temperature and its vertical gradient. Effects of this kind are best illustrated by comparing observations on the same elevation, for example, between mountain tops or slopes and the free atmosphere, or between the valley floor and the adjacent flat land.

Figure 2.2 shows the vertical profile of mean January and July temperatures drawn from surface observations at various elevations in the Swiss Alps. In both seasons, distinct thermal regimes can be identified between the north and south sides of the main Alpine crest. Below 1500 m MSL, the southern Alpine area is about 2 to 4 K warmer than the north. This spatial variation over a horizontal distance of only about 200 kilomters exceeds the typical latitudinal temperature gradients in Central Europe (see, e.g., Schönwiese et al. 1993) and is indicative of topographic effects. The shielding of the south side against cold air advances from the north and northwest (cf. section 2.2.2) as well as influences on the distribution of cloud cover and hence solar insolation (cf. section 2.2.5) contribute to the regional enhancement of the temperature gradient across the Alps.

Horizontal temperature gradients are also observed between the Alpine ridge and the conditions in the free atmosphere over the adjacent foreland, both on seasonal and diurnal timescales. Based on frequent upper-air balloon soundings and summit observations during three summer months, Richner

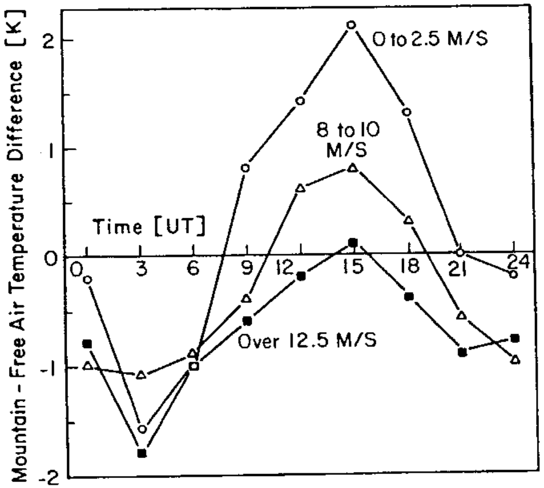

Figure 2.3 Mean summit minus free air temperature difference (K) in the Alps as a function of the time of day (universal time) and wind speed in summer. (After Richner and Phillips 1984; figure taken from Barry 1992.)

and Phillips (1984) describe the summit minus free-air temperature difference (measured at the same altitude) as a function of the time of day and wind speed (see figure 2.3). The thermal difference has a pronounced diurnal cycle, especially during calm situations, with relatively warmer (colder) summits during daytime (nighttime). During episodes of strong flow, mean summit air temperatures are colder than free-air temperatures at the same altitude, a property that derives from the adiabatic cooling associated with flow over the mountains (see section 2.3.2). Observations of the diurnal temperature variations during fair weather conditions (Wagner 1932; Phillips 1984) reveal a larger amplitude of the temperature cycle at Alpine valley floor stations as compared to the foreland regions, reflecting thermal gradients between the mountains and the foreland. Although such behavior can be expected primarily for summer high-pressure situations, it is also evident in the multiyear climatological mean (Schüepp and Schirmer 1977). Section 2.3.2.4 discusses some of the flow systems and meteorological phenomena associated with the inner-Alpine region's relative daytime warmth and nighttime coolness.

Motivated in part by the current discussion of climate change, a number of research groups have examined the interannual and longer-term temperature fluctuations in the Alpine region (von Rudloff 1971, 1986; Müller-Westermeier 1992; Pasquale, Flocchini, and Russo 1992; Sneyers, Böhm, and Vannitsem 1992; Auer, Böhm and Mohnl 1993; Böhm 1992, 1993; Weber, Talkner, and Stefanicki 1994; Beniston et al. 1994). Such investigations require a homogenization of the time series to correct for artificial variations introduced by frequent changes in observational techniques or in local envi-

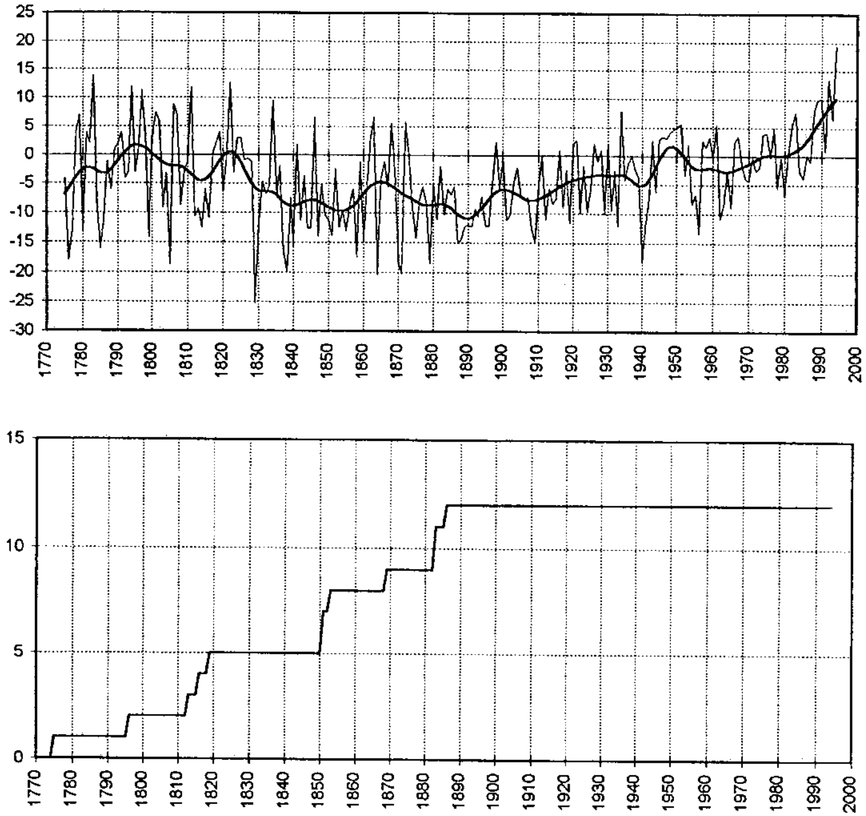

Figure 2.4 Surface air temperature in Austria 1775–1991. Top panel: Deviations of annual means from the thirty-year mean (1961–90) in units of 0.1 K, and low-pass filtered temperature (smooth curve; filter width twenty years). Bottom panel: Number of stations involved in the calculation of the annual means. (Courtesy of Dr. Reinhard Böhm, ZAMG Wien; see also Böhm 1993.)

ronmental conditions such as urbanization, agricultural changes, and deforestation. The reconstructed time series provide reliable records. For some of the Alpine stations, these date back to the beginning of the instrumental period in the late eighteenth century.

Figure 2.4 depicts the variations of annual mean temperature for Austria (Böhm 1993). The time series is based on a maximum of twelve single-station temperature records and is found to be representative for the northeastern Alpine region. A substantial year-to-year variability on the order of 1 K is evident. The longer-term variation, illustrated by the low-pass filtered (smooth) curve, shows warm periods at the beginning of the instrumental record until about 1820, and during the most recent fifty years. Between these two warm periods, the low-pass filtered temperature is substantially lower, with two major minima around 1850 and 1890. The warming since the beginning of this century was found to be stronger in autumn and winter than in spring and summer.

The salient characteristics of the temperature evolution in Austria are in qualitative agreement with other Alpine observations of similar duration (e.g., von Rudloff 1986; Müller-Westermeier 1992). Moreover, the temperature increase between the late nineteenth century and the 1950s is documented for many surface observing sites all over Central Europe (see, e.g., Schönwiese et al. 1993; Heinemann 1994). Concomitant with this warming, major recessions have been registered for many Alpine glaciers (cf. section 2.2.4).

The observed warming in the Alpine region during this century is in accord with (although slightly larger than) the increase in Northern Hemisphere temperatures (Houghton, Jenkins, and Ephraums 1990; Houghton et al. 1995) and accompanies the increase in atmospheric CO_2. It does not necessarily follow from these observations, however, that a causal relation exists between the regional warming and anthropogenic GHG emissions. For example, the warm phase around 1800 indicates that regional long-term temperature fluctuations on the order of 1 K can also result from natural variations of the climate system without simultaneous variations in CO_2. Nevertheless, the observed warming during this century is roughly consistent with the expected magnitude of GHG effects, and currently available climate change scenarios for the Alpine region suggest that an accelerated warming could take place in the near future (see chapter 4).

2.2.2 Pressure and the *Grosswetterlagen*

The pressure value at a particular location has comparatively little direct meaning for local weather and climate. The pressure's horizontal gradient, however, is closely related to the larger-scale horizontal flow field, and the three-dimensional pressure distribution is an indicator of the associated vertical velocity field. Thus the pressure pattern itself is related directly to the type of weather experienced in a region. Hence a customary approach to the study of weather and climate is to adopt a synoptic-climatological method and to characterize the weather in the European sector in terms of the occurrence of distinct weather regimes: the *Grosswetterlagen* types. Each regime corresponds to the persistence of a certain typical circulation pattern for a period comparable to or greater than the lifetime of individual synoptic systems.

Several such classification schemes have been developed, and among the most widely used are those of Lamb (1972) and Hess and Brezowsky (1952; see also references in Gerstengarbe and Werner 1993). These schemes categorize the pattern in terms of the ambient wind direction and introduce a further subdivision in terms of the dominant pattern of the pressure systems (high or low). For example, a southwesterly flow with a dominant low located west of Ireland with cyclonic curvature over central Europe (the *SWz*

Table 2.1 Relative frequencies of the Alpine weather types between 1945 and 1991

Weather type	Spring [%]	Summer [%]	Autumn [%]	Winter [%]	Year [%]
Convective					
Anticyclonic	12	20	18	14	16.0
Weak pressure gradients	30	36	28	16	27.5
Cyclonic	10	8	6	6	7.5
Advective					
West	10	10	10	14	11.0
North	16	12	16	24	17.0
East	6	2	4	8	5.0
South	12	6	12	10	10.0
Eddy over the Alps	6	6	6	6	6.0
Total					100.0

Note: Categorizations after Schüepp 1968.

category in the Hess and Brezowsky scheme) is consistent with cyclones tracking toward Scandinavia and the passage of fronts southward toward the Alps.

Further variants of the synoptic-climatological approach are worth noting. First, the method has been refined to apply specifically to the Alps by directly classifying weather types based on the pressure pattern in the immediate vicinity of the region itself. The two most frequently used schemes (cf. Wanner 1980; Fliri 1984) are those of Lauscher (1958) and Schüepp (1968). Table 2.1 illustrates a climatology in these terms and lists the relative frequencies of the Schüepp weather types in the seasons. Note, for instance, the predominence of northerly flows in winter and the two maxima of southerly flows (which include Föhn) during autumn and spring. Second, in a seminal paper, Kirchhofer (1974) examined the statistical relationship between European-scale weather types and the meteorological parameters at specific Alpine locations and in so doing pioneered the present downscaling studies (see chapter 4). Third, modern statistical techniques are increasingly being used to deduce fully objective synoptic classification schemes (see, e.g., Cacciamani, Nanni, and Tibaldi 1995).

It follows that distinctive aspects of the climate and its variability can be inferred from the statistics for the occurrence, persistence, and alternation of the contrasting weather regimes. The statistics for the last decades show several significant features (Bardossy and Caspary 1990; Gerstengarbe and Werner 1993; Murray 1993). For the annual mean, Bardossy and Caspary detected an increase in the frequency of southerly circulation types (from approximately 5 percent to approximately 10 percent) between 1880 and 1990. The most substantial changes in the last few decades occurred in winter, when the frequency of zonal (westerly) circulation patterns increased substantially at the expense of easterly and northerly configurations. These trends are consistent with the increased frequency of wet and warm winters during the last decades. We return to the link between the foregoing European-scale variations and larger-scale features of the atmospheric circulation in section 2.3.1.

2.2.3 Precipitation

Precipitation in mountainous regions often differs considerably from that in the surrounding lowlands, and section 2.3.2 dicusses some of the meteorological mechanisms responsible for these differences. The Alpine precipitation signal reveals a great spatial variability from the scale of the whole Alps to that of single slopes, and this concerns both the long-term mean as well as the occurrence of strong precipitation. To describe the main features of the Alpine precipitation climate, we proceed by discussing in turn the long-term mean distribution, the seasonal variability, and finally the long-term changes.

Today there are approximately 8,000 rain gauges in use in the Alps belonging to different national networks, and only a small fraction of their observations are distributed internationally. Collection and homogenization of the data is a demanding logistical and statistical task, and a large number of studies are based merely on national rather than Alpine-wide data sets. Precipitation analysis is particularly hampered by observational errors that can attain 15 percent for rain and up to 50 percent for snow (Sevruk 1985). The principal error source is induced by wind field deformations above the gauge, and the high and exposed Alpine stations are especially subject to this problem. Measurements are sensitive to the ambient wind speed, and this introduces an additional error source associated with changes in the surroundings of the gauges, such as tree growth, erection of buildings, and relocation of the gauges. These factors add to data inhomogeneity.

Long-term means based on national data sets can be found in Steinhauser 1953 for Austria, in the *Atlas Climatique de la France* (see Direction de la Meteorologie Nationale 1988, 1989) for France, Schirmer and Vent-Schmidt 1979 for Germany, Touring Club Italiano 1989 for Italy, and for Switzerland in Uttinger 1949 or in the Hydrological Atlas (Landeshydrologie und Geologie 1992). Long-term means of precipitation for the whole Alpine region are provided by Fliri (1974); Baumgartner Reichel, and Weber (1983); and Frei and Schär (1998). Figure 2.5 shows the yearly precipitation totals of the period 1931–60. The map is based on data from about 1,000 rain gauges, most of which are located at altitudes lower than 2000 meters. A relatively dry zone can be noted along the main crest of the Alps surrounded by wet zones to the north and south, and only at a few locations, for example, the Mont Blanc or St. Gotthard regions, do the two wet regions merge.

Not apparent in the smoothed distribution of figure 2.5 are the local extremes found at the driest spots located in the Aosta Valley, with annual

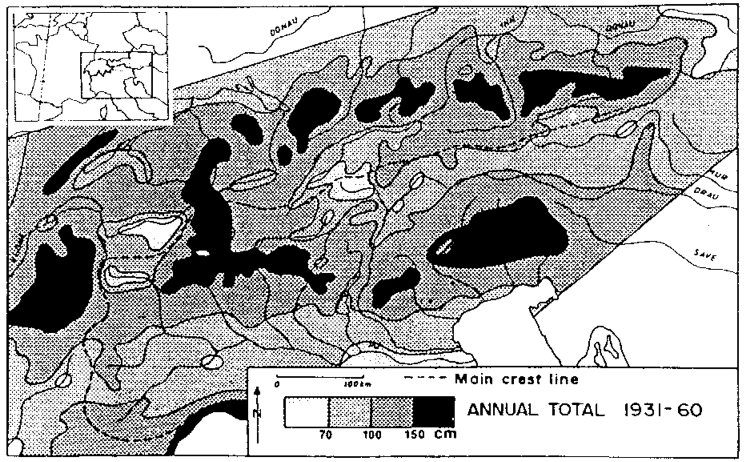

Figure 2.5 Mean annual precipitation sums, 1931–60. (From Fliri 1974.)

means below 550 millimeters per year; in the Venosta Valley and the Valais, with values around 600 millimeters per year; and in the dry northern Alpine valleys of the Grisons and northern Tirol. In the northern wet zone, annual means exceed 2,000 millimeters per year at several locations; in the southern wet zone in the Friuli region, precipitation values reach 3,000 millimeters per year at a few spots, whereas about 2,500 millimeters per year is observed in the Centovalli valley in Ticino. In general, the precipitation totals increase with altitude, but this relation shows strong variations with respect to location and season (Uttinger 1951; Lauscher 1976a; Lang 1985; Blumer and Spiess 1990; Frei and Schär 1998). Figure 2.6 displays an indication of the seasonal cycle of precipitation, based upon the same data as figure 2.5. In the northeastern Alps there is one maximum in summer, whereas the northern and north-western Alps show a second, somewhat weaker maximum in winter. In the southern regions, two maxima arise in spring and autumn.

Particular interest in Alpine precipitation also arises from the frequent occurrence of strong precipitation events in the region and their attendant adverse effects. Either directly, by flooding populated valley floors, or indirectly, by triggering landslides and avalanches, severe precipitation can cause catastrophic damage to agriculture and human infrastructure. A sequence of flooding events during the last few years (Piedmont, November 5–7, 1994; Brig, September 23, 1993; Vaison-la-Romaine, September 23, 1992), some involving loss of human lives, has tragically demonstrated the threat of extreme precipitation and the vulnerability of the Alpine region. In addition, during winter and spring, rapid snow melt during episodes of sudden warming can enhance normal runoff, and the resulting water discharge through the major rivers can affect areas far remote from the Alps.

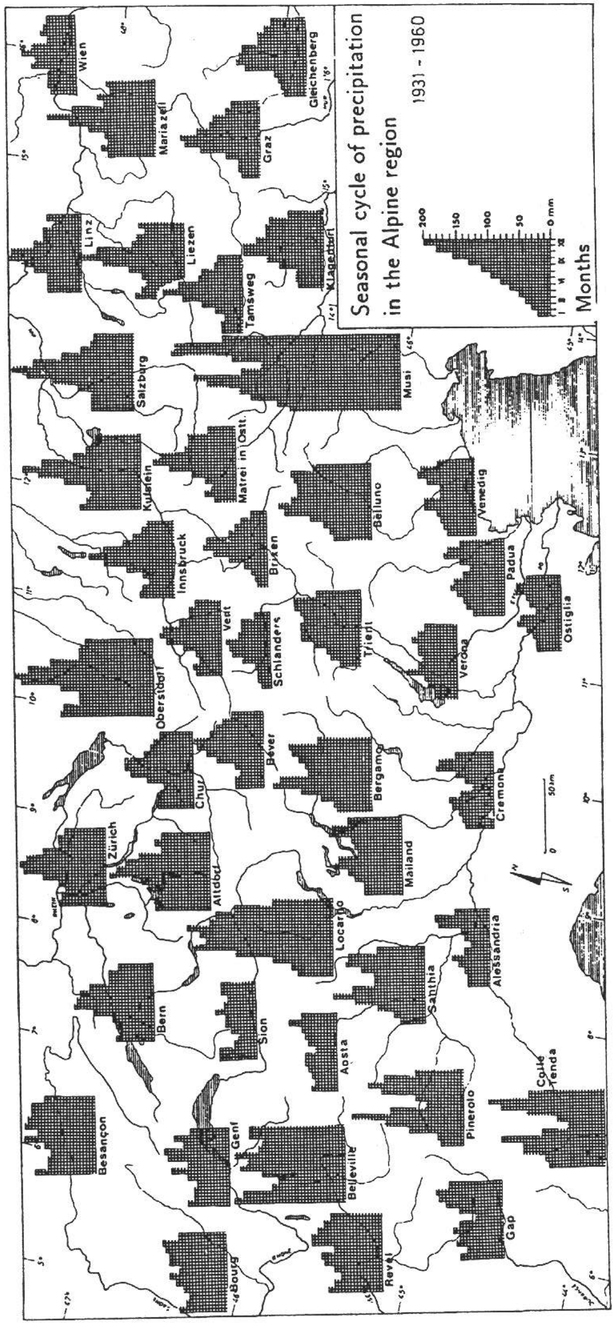

Figure 2.6 Mean monthly precipitation, 1931–60, at selected sites. (From Fliri 1974.)

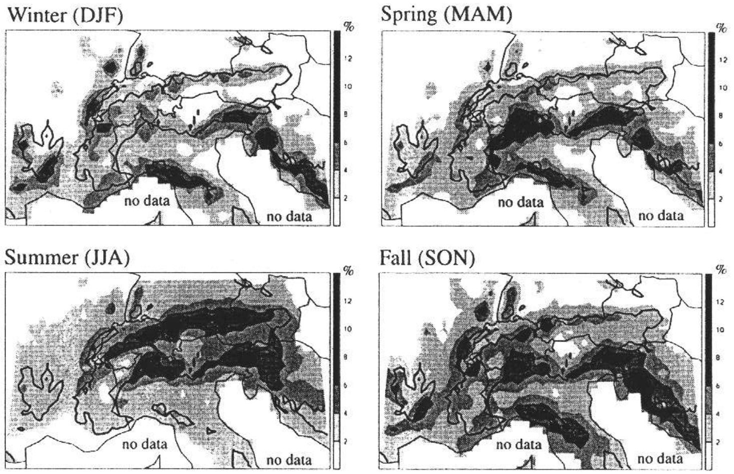

Figure 2.7 Frequency (in percent) of days with strong precipitation (daily total ≥ 20 millimeters) in the Alpine region for the seasons of the year. The bold contour indicates the topography at a height of 800 meters. (Based on data described in Frei and Schär 1998.)

The significance of strong precipitation in the Alpine region also arises from the fact that relatively rare intense events contribute a substantial amount to the long-term mean. For instance, precipitation during the wettest 4 percent of days contributes about 40 percent to Alpine precipitation totals. Statistical analyses of extreme precipitation in some areas of the Alpine region can be found, for example, in Geiger, Zeller, and Röthlisberger 1991; Nobilis, Haiden, and Kerschbaum 1991; and Bonelli and Pelosini 1992. Figure 2.7 displays an Alpine-wide analysis of the frequency of strong (but not necessarily extreme) cases based on twenty years of data at about 3,000 rain gauge stations. Regions with frequent occurrence of strong precipitation (daily values of at least 20 millimeters) can be identified along the northern and southern rim of the Alpine ridge, whereas inner-alpine valleys and the adjacent flatland areas appear less affected. Strong precipitation is most frequent during summer, and during this season it mainly originates from severe thunderstorms (heavy, organized, convective systems; cf. section 2.3.2.5). In spring and fall, high activity is found in particular to the south of the main Alpine crest, that is, in the northwestern Po valley and the Ticino and Friuli areas. Here, strong precipitation usually relates to the advection of warm and moist air from the Mediterranean toward the Alps, induced by the approach of a cold front from the west or the development of mesoscale cyclones over the western Mediterranean (cf. section 2.3.2.2). Finally, during the winter season the frequency of strong precipitation is generally lower.

Figure 2.8 shows the time series for the annual precipitation totals and the number of wet days (with at least 0.1 millimeter) at the Sonnblick Observatory (3106 meters MSL) in Austria. This remote site, unaffected by direct

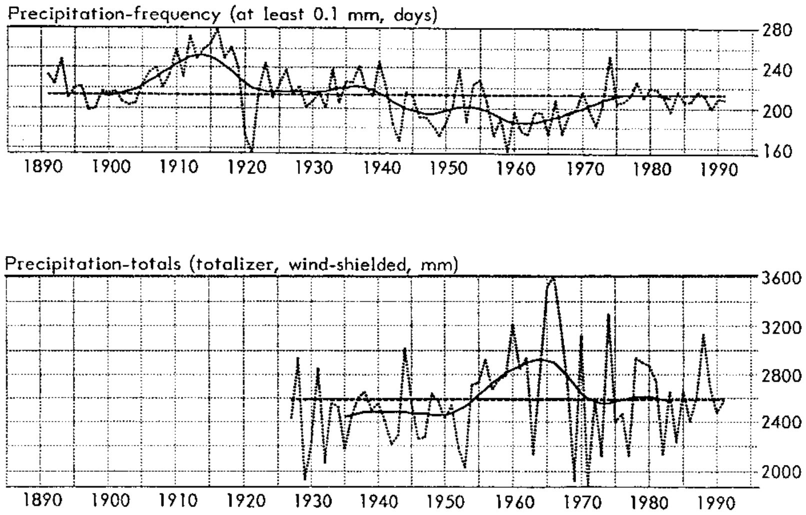

Figure 2.8 Annual precipitation frequency and annual precipitation sums at Sonnblick Observatory. Smoothed curves obtained by twenty-year Gaussian low pass filtering. (From Auer, Böhm, and Mohnl 1993.)

human activities, provides a very homogeneous data set appropriate for the study of interannual variability. Both series reveal large year-to-year variability. Furthermore, the twenty-year low pass filtered data show fluctuations on longer time scales, like the precipitation maximum in the 1960s or the decrease in the frequency of wet days from 1915 to 1960. The low correlation between the amount of precipitation and the frequency of wet days indicates the complexity of precipitation processes. Since the maximum in the totals coincides with the frequency minimum, the increase in the precipitation totals during the 1960s is attributable to an increase in intermediate- and high-intensity events.

The most comprehensive study of precipitation trends in Europe is the *Klimatrend-Atlas Europa 1891–1990* (Schönwiese et al. 1993, 1994). The study revealed a long-term increase of winter precipitation over the whole Alpine region, in its western part by up to 20 percent, during the last 100 years. This trend is consistent with a recent trend analysis conducted for Switzerland based on a relatively dense precipitation network of 113 rain gauge stations with daily resolution (Widmann and Schär 1997). This study detected an increase of winter precipitation in the northern and western parts of Switzerland by as much as 30 percent, and smaller but still positive trends for the southeastern parts of the country. The statistical significance of the trend at single stations using the Mann-Kendall test (Sneyers 1990; Denhard and Schönwiese 1992) is rather low, and the significance level is about 70 percent (Schönwiese et al. 1993). However, the notion of significance in climatological applications is dependent on the spatial resolution. If a region with a common long-term trend is considered, the investigation of the area mean rather than single station records can substantially increase the significance level. Using such (weighted) area means, the increase of winter precipitation during this century can be demonstrated to be statistically significant at a level of up to 90 percent and furthermore to possess some regional variations of similar significance (Widmann and Schär 1997). On the other hand, the decrease of autumn precipitation during this century by 20 percent reported by Schönwiese et al. is of substantially lower statistical significance over Switzerland.

In the European-scale analysis of Schönwiese et al., trends have also been detected for the shorter period 1961–90. An increase of spring precipitation by 30 percent is observed in the southwestern regions of the Alps, whereas the northeastern regions show a decrease of 25 percent. During summer, no significant trends can be detected. The northern and southern regions of the Alps behave differently in the autumn season, with an increase (up to 25 percent) to the north and a 25 percent decrease to the south. As for the centennial period, an Alpine-wide increase of precipitation (25 percent) occurs during winter (Schönwiese et al. 1993, 1994; Widmann and Schär 1997).

The occurrence of warm winters with a substantial surplus of precipitation represents an anomaly on a timescale of many hundreds of years. Already Pfister (1984; see also Pfister 1992) has noted in his analysis of proxy and early instrumental records starting in the sixteenth century that the winters in 1965–80 were warmer (by 1.3 K) and moister (by 25 percent) than the 1901–60 mean, and that such pronounced deviations had never occurred in the previous 500 years. As indicated above, this trend toward warmer, wetter winters has continued into the 1990s, and may constitute one of the first signals of global climate change in the Alpine region.

2.2.4 Snow and Ice

The seasonal snow cover, permanent high-Alpine snow fields, glaciers, and permafrost regions are characteristic features of the Alpine landscape and constitute an integral part of the regional climate system. The significance of the Alpine cryosphere components relates to their sensitivity to the distribution and variation of primary climatic elements (temperature and precipitation), to their effects on atmospheric climate processes, and to their relevance for ecosystems and human infrastructure in the region (agriculture, tourism, water resources, hydropower, etc.). In this section, we discuss the spatial and temporal variations in the Alpine snow cover and report on signatures of past climate variations inferred from glacier observations during the last few hundred years. For longer-term variations, including the ice ages, see chapter 3.

The duration, depth, and water equivalent of the snow cover in the Alps depends primarily on altitude (Schüepp, Gensler, and Bouët 1980; Witmer et

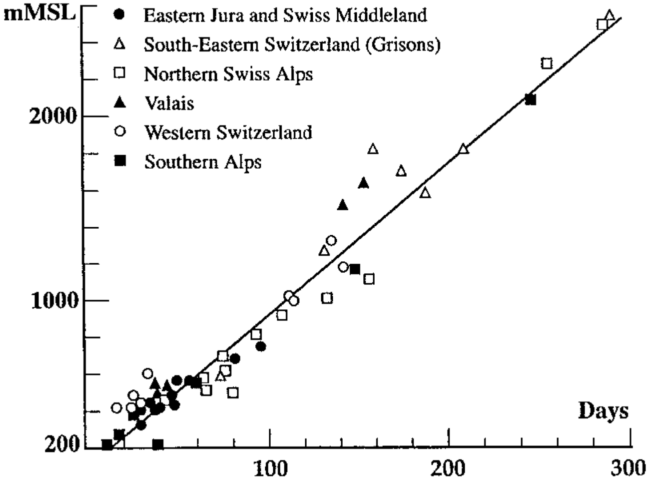

Figure 2.9 Altitude dependence of the average number of days with snow cover at Swiss observing stations (1959/60–1978/79). Distinct symbols are used for various subregions. (After Schüepp, Gensler, and Bouet 1980.)

al. 1986; Fliri 1991) and is controlled by the temperature decrease and precipitation increase with altitude (cf. sections 2.1 and 2.3). Figure 2.9, based on Swiss observations, shows the average number of days with snow cover per year as a function of elevation. For the northern and southern alpine foreland (200–500 meters MSL), continuous snow cover is found rarely for more than two consecutive weeks. At these elevations, snow depths during snowfall episodes typically reach only several to a few tens of centimeters, and intermittent melting periods often lead to widespread disappearance of the snow even deep in the winter. For Alpine areas at elevations between 1,000 and 3,000 meters MSL, however, the snow cover is predominantly seasonal. It accumulates during winter (typically starting in November) and then melts off during spring and early summer (see figure 2.10). Finally, above the climatological snowline, which varies from 2,400 to 3,400 meters MSL (Stone 1992), extended snow fields remain throughout the year. At these elevations, mean summer temperatures are still near freezing, and a substantial fraction of summer precipitation falls as snow (Lauscher 1976b).

Consider now further temporal and spatial features of the snow cover. Its seasonal character, with accumulation of winter snow falls, and its reduction during the melting period are responsible for the summer maximum in the annual runoff pattern of major Alpine rivers (see, e.g., Martinec 1987). Spatial variations also modulate the cover's primary altitude dependence. On the scale of the entire ridge, observations indicate a larger snow-water equivalent, a longer-lasting seasonal cover, and a lower elevation of the snowline along the northern and southern flanks of the Alps as opposed to the inner-

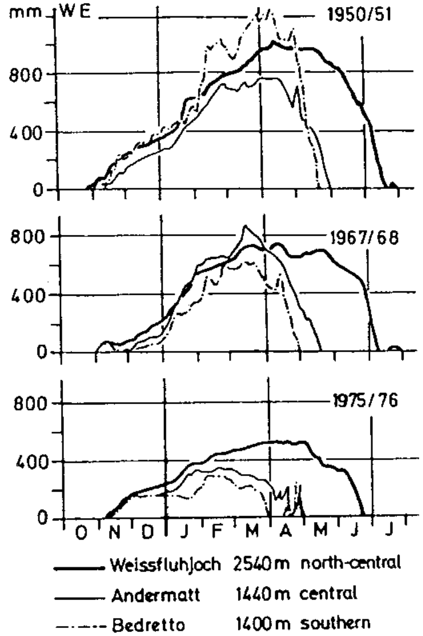

Figure 2.10 Time trace of the accumulation and ablation of the seasonal snow cover water equivalent (WE) at a north central, a central, and a southern site in the Swiss Alps. The three panels represent winter periods with a relatively thick, an average, and a relatively thin snow cover, respectively. (From Lang and Rohrer 1987.)

Alpine region (Lang and Rohrer 1987; Fliri 1991; Stone 1992). This feature is attributed to the partial shielding of deep central Alpine valleys during major snowfall events (Lang and Rohrer 1987) and correlates with the regional distribution of mean winter precipitation (Fliri 1974; see also section 2.3).

The amount, extent, and distribution of the snow cover in the Alps is subject to large year-to-year variations. Conditions between extreme winters can vary by more than a factor of 2 for the maximum snow-water equivalent and by one month for the spring termination of the snow cover (see figure 2.10). Föhn (1990) discusses longer-term variations and trends inferred from snow cover observations in the Swiss Alps. For the last fifty years, no significant trends are found for deep-winter snow conditions (i.e., snowfall, depths, profiles and avalanche activity; see also De Quervain and Meister 1987). There are some indications, however, of recent changes in the early winter situation, characterized by the snow depth at January 1. Since about 1982, a pronounced decrease is evident at a number of western and southern alpine sites (Föhn and Plüss 1989). As a result of the pronounced warming observed in the Alpine region (see section 2.2.1), winter precipitation is increasingly falling as rain rather than snow. Simple estimates taking into account this and other factors suggest that a warming of 1 K can lead to a rise in the snowline of 100 to 200 m (see, e.g., Abegg and Froesch 1994).

The large year-to-year fluctuations of primary climatic elements (cf. sections 2.2.1 and 2.2.3 and the discussion earlier in this section) often make it difficult to discern slowly varying climate signals such as long-term trends. In this respect the behavior of mountain glaciers and their long-term monitoring in the Alps offer valuable additional information. Mountain glaciers are sensitive to changes in external climatic conditions. In contrast to the seasonal response, which is more immediate, glaciers react with some delay to climatic variations on annual or secular timescales (Patzelt and Aellen 1990) and exhibit a favorable "signal-to-noise ratio" for the observation of slowly varying climate signals.

The growth and retreat of glaciers is associated with snow accumulation and with ice ablation by melt and sublimation. Whereas snowfall during the winter half year predominantly controls accumulation, ice ablation is particularly sensitive to summer temperature and sunshine and also depends on the deposition of pollutants and dust, which influences the radiative characteristic of the glacier surface. An imbalance between accumulation and ablation is accompanied by a reaction or adaptation of the glacier's ice flow characteristics and typically results in an advance (retreat) of the glacier terminus in case of a mass increase (decrease). The type, magnitude, and lapse time of this reaction can vary greatly, however, between individual glaciers (see, e.g., Kuhn 1990) and makes it necessary to consider an ensemble of glaciers for an indirect assessment of climate variations.

Patzelt and Aellen (1990) discuss the evolution of Alpine glaciers since the modern glacial maximum around 1850. Analyses of available observations indicate a major reduction of glacier area by 46 percent in the Austrian Alps and by about 30 percent in the Swiss Alps. The general tendency is clearly evident in the evolution of length changes (see figure 2.11). A majority of Swiss glaciers have been retreating during most of the period. In accord with similar observations for the Austrian Alps, the retreat is interrupted by short episodes of recovery, which are attributed to cool, cloudy summers (in the 1890s; see also section 2.2.1) and a number of winters with heavy snowfall (1915–20). Increased temperatures and sunshine duration as well as reduced precipitation in the high-Alpine area were found to accompany the major retreat phase between 1930 and 1950 (see also von Rudloff 1962). The brief readvances during the late 1970s and early 1980s, which affected predominantly smaller glaciers, largely ceased and the general retreat of Alpine glaciers continued during the last decade of this century.

2.2.5 Surface Energy Balance

Classical climatological parameters like temperature measure the underlying physical processes indirectly and provide little indication of why a particular climate prevails at a specific location. The local climate depends both on

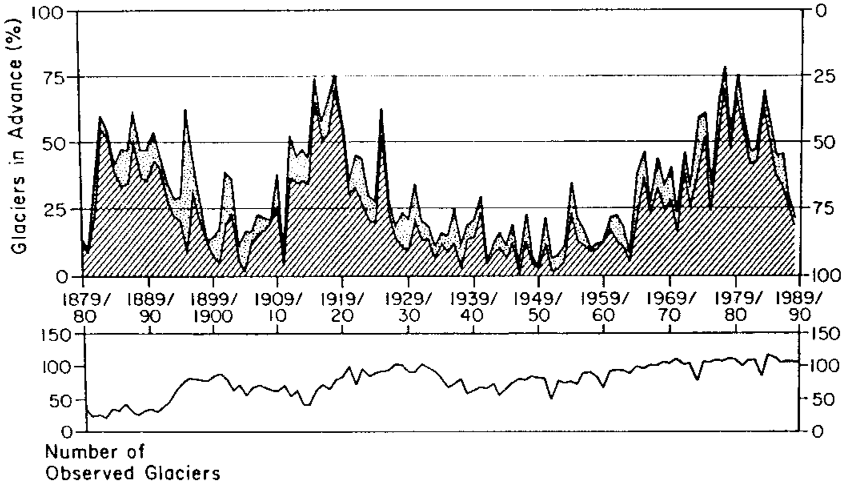

Figure 2.11 Percentage of Swiss glaciers showing advance (hatched) or retreat (white) of the terminus. The stippled portion denotes stationary glaciers. The bottom panel gives the number of observed glaciers in each year. Data from the World Glacier Monitoring Service, Zürich. (See also Aellen 1994; figure taken from Barry 1992.)

regional processes (like the synoptic setting; see section 2.2.2) and also on point processes (discussed in this section). Energy transformations at the earth's surface heavily affect the evolution of the near-surface temperature and humidity field. The associated processes (see figure 2.12) include the absorption of solar direct and diffusive radiation (together referred to as global radiation), the emission and absorption of longwave infrared radiation, and the turbulent exchange of energy in the atmosphere's surface layer through vertical fluxes of sensible and latent heat.

The net radiative energy flux (defined as the surplus of absorption over emission) drives the near-surface processes, and the "surface energy balance" describes how this energy input is redistributed into the turbulent fluxes of sensible and latent heat, the heating of the soil, and the melting of snow. The relative importance of the sensible and latent heat fluxes depends on the properties of the underlying surface and the local atmospheric conditions. Over dry soil, evaporation plays a minor role and the net radiative flux is primarily converted into turbulent fluxes of sensible heat and the heating or cooling of the soil. In contrast, over wet soils and the sea, evaporation and transpiration can dominate the turbulent energy flux by transporting the latent (condensational) heat. As with sensible energy fluxes, its effect is to cool the surface (by evaporation) and ultimately heat the atmosphere (by condensation).

Mountains are instrumental in affecting the shortwave radiation budget. Analyses of clear-sky mountain observatories' data in the Alps show that

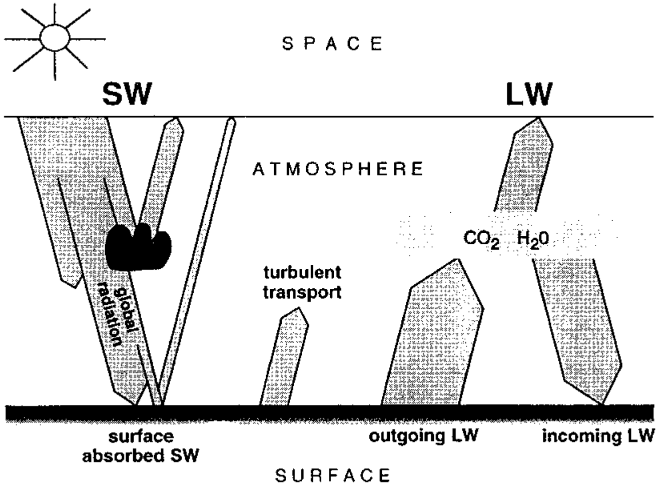

Figure 2.12 The atmosphere's global energy balance and its major contributions: shortwave (visible) radiative fluxes, longwave (infrared) radiative fluxes, and turbulent atmospheric transport of sensible and latent heat.

the global radiation increases in tandem with the altitude by approximately 0.8 Wm^{-2} per 100 meters (Müller 1984, Barry 1992). This effect is related to the overlying atmosphere's optical depth, which determines the amount of shortwave absorption and scattering. In the climatological mean (which includes clear-sky and overcast conditions), the vertical gradient of the global radiation is less firmly established. Data from sites at different altitudes in Switzerland (Ohmura et al. 1990) indicate an increase of insolation with altitude during winter and a decrease during summer. The decrease in summer is related to the shading associated with topographically generated convective clouds, whereas in winter lowland areas are frequently beneath a stratiform cloud cover. Observations of global radiation furthermore indicate that sites on the Alps' southern slope receive substantially more insolation than those on the northern slope (see figure 2.13), an effect associated with the smaller mean cloud amounts to the south of the Alps.

The albedo of the underlying surface controls the absorption of incoming global radiation. The persistence of snow covers in elevated areas substantially increases the albedo and leads to increased reflection and reduced absorption of global radiation (Barry 1992). On the scale of individual valleys, the valley slopes' orientation with respect to the sun as well as shading effects by neighboring mountains also control the absorption of shortwave radiation. Over complex terrain, the resulting surface temperature contrast drives a range of mountain wind systems (see section 2.3.2.4).

The incoming longwave radiation emitted from the atmosphere and the outgoing longwave emission from the surface determine the longwave-

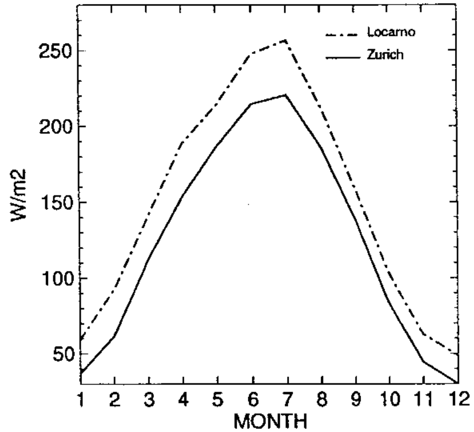

Figure 2.13 Mean annual cycles of global radiation at Zurich (north of the Alps) and Locarno (south of the Alps) in W/m^2. (From Wild et al. 1995, based on data from Ohmura, Gilgen, and Wild 1989.)

radiation balance at the surface. Both the incoming and outgoing fluxes decrease with altitude because of the decreasing water vapor content of the overlying air, and the decreasing soil temperature, respectively. The observed vertical gradient of incoming longwave radiation at different altitudes in the Alps is estimated to be close to -3 Wm^{-2} per 100 meters, whereas the vertical gradient of outgoing longwave radiation is estimated to be close to -2 Wm^{-2} per 100 meters. This yields a slight increase of the net longwave cooling with altitude.

The sum of the absorbed short- and longwave radiative exchanges between the surface and the atmosphere form the surface net radiation. This quantity determines the amount of energy available for the nonradiative components of the surface energy balance. The mean net radiation tends to decrease with altitude because of the longer duration of snow cover and the increase in net longwave cooling (Barry 1992). Accordingly, turbulent fluxes transfer less energy from the surface into the atmosphere at higher elevations. Quantitative estimates of the reduction of evapotranspiration with height show a large spread from -7 to -36 mm y^{-1} per 100 meters, as reviewed by Lang (1981).

Time series that allow an estimation of the temporal variability of the components of the surface energy balance over many years are very limited and largely restricted to global radiation. Figure 2.14 shows time series of annual and seasonal global radiation for the region of Zurich (average of three stations). In this area, the global radiation decreased from 1959 to the end of the 1970s by as much as 20 percent, and tended to recover thereafter, a trend found in winter and summer as well as in the annual mean. A similar trend was detected at other monitoring stations in the interior of Europe (including Salzburg and Potsdam). A substantial number of these decreases

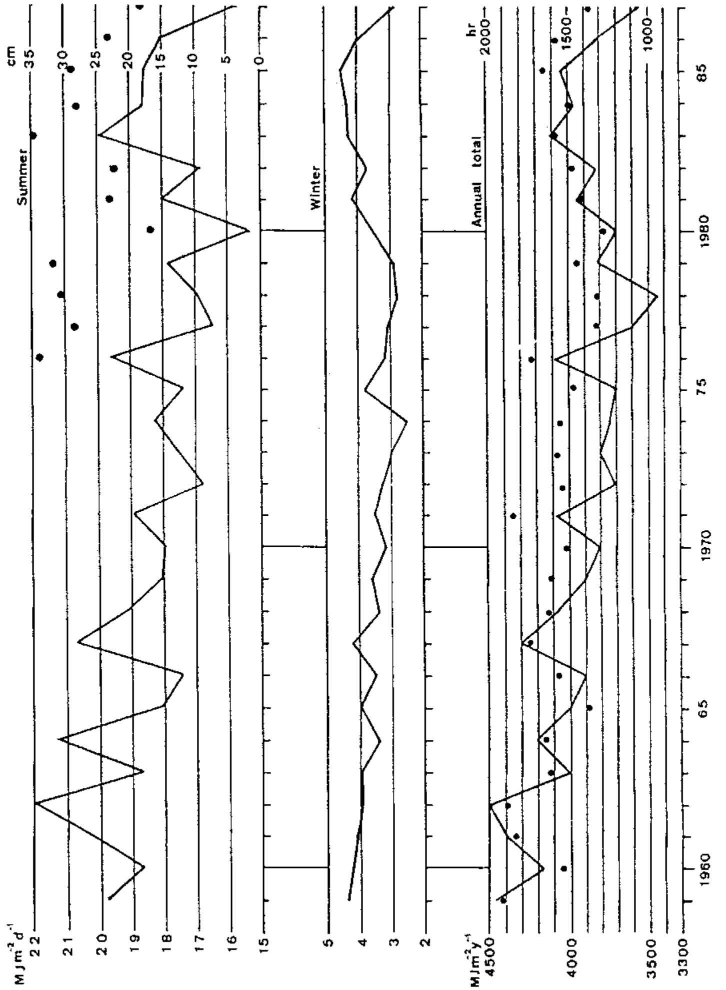

Figure 2.14 Time series of annual and seasonal (summer, winter) global radiation for Zurich (1 MJ m^{-2} d^{-1} = 11.6 W/m^2; 1000 MJ m^{-2} $year^{-1}$ = 31.7 W/m^2). Dots in the upper right are summer evapotranspiration. Dots in the bottom panel are annual total sunshine duration for Zurich. (From Ohmura and Lang 1989.)

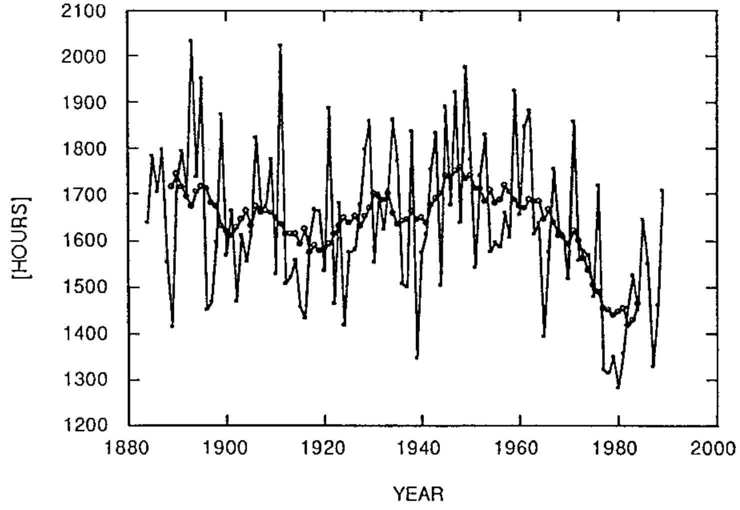

Figure 2.15 Annual total sunshine duration hours for 1884–1985 with the eleven-year running means for Zurich. (From Ohmura and Lang 1989.)

resulted from changes in cloud cover. Global radiation's variability significantly affects terrestrial processes and is considered the primary reason for the fluctuations in the time series of evapotranspiration measured at the experimental hydrological basin Rietholzbach, 30 kilometers east of Zurich (see figure 2.14).

The monthly global radiation correlates well with the monthly sunshine duration, as can be seen from figure 2.14. The analysis of the global radiation's variability can thus be extended back to the time when sunshine monitoring was initiated. Figure 2.15 shows such a 102-year time series of annual total sunshine duration measured at Zurich. A power spectral analysis of this time series (see Ohmura and Lang 1989) suggests that the global radiation has changed with a periodicity of twelve years and an overlying decreasing trend.

Changes in the radiative fluxes are an important aspect of climatic changes. The most direct effect of an increased concentration of carbon dioxide in the atmosphere is an increase of the longwave atmospheric radiation directed to the surface. It is a fundamental question in climate change studies how this additional energy will be redistributed within the components of the surface energy balance.

2.3 CLIMATE PROCESSES: WHAT DETERMINES THE ALPINE CLIMATE?

The climate experienced in the Alpine region is linked to a long chain of processes that operate on a wide range of spatial scales. On the global scale, the incident solar radiation and the atmosphere's composition are the main contributors to setting the global mean temperature and establishing the (seasonally) varying pole-to-equator temperature differences. Again, the longitudinal variations in the planet's distribution of oceans and continents and the presence of major mountain ranges contribute to establishing the atmosphere's continental-scale time mean atmospheric circulation pattern. Within this large-scale pattern, smaller (synoptic-scale), transient mid-latitude weather systems form, propagate, and decay, and to a great measure the passage of these systems—fronts, cyclones and anticyclones—determines the day-to-day weather variations in the extratropics. In turn, Alpine climate is a register of the ensemble effect of the synoptic-scale weather systems, including their orographic modification.

Four aspects of the foregoing chain of spatial scales merit further comment. First, the influence is not merely down-scale. For example, the genesis, track, and strength of the synoptic-scale weather systems not only is strongly dependent upon, but also significantly influences the pole-to-equator thermal difference and the large-scale circulation pattern. Second, each of these scales shows significant temporal variability. For example, the planetary waves exhibit considerable fluctuations with respect to the long-term time mean and thereby induce important anomalies in the occurrence of the different weather types (*Grosswetterlagen*) on time periods of between a week and a season. Third, any change in the global climate setting is also transmitted along this chain. In particular, because Alpine climate is strongly influenced by the eastward passage of fronts and cyclones from their genesis region in the Atlantic, changes in the strength, frequency, location and track length of Atlantic cyclones substantially influence the region's climate variations. Fourth, phenomena resulting from the Alps' modifications of weather systems have considerable influence at large distances downstream along the (variable) storm track, affecting at times practically the entire eastern Mediterranean region. A program to establish the causal connection between global and alpine climate change therefore needs to examine the possible multiple linkage(s) in the aforementioned chain.

In this section we consider, in the context of climate, the form of the large- and synoptic-scale circulation patterns in the European sector and some of their principal variations; the dynamics of the Alpine modification of the incident flow; and the form of the subalpine-scale systems.

2.3.1 Large-Scale Setting and Variability

2.3.1.1 The Large-Scale Mean Flow and the Atlantic Storm Track Figure 2.16 shows the winter and summer seasonal mean sea-level pressure (SLP) pattern and depicts a measure of the day-to-day variability of the mid-tropospheric flow in the east Atlantic–European sector. In winter, (see panel (a) of the figure) the Alps are at the center of the deformation-like pattern formed by the Icelandic Low to the northwest, the Azores High to the

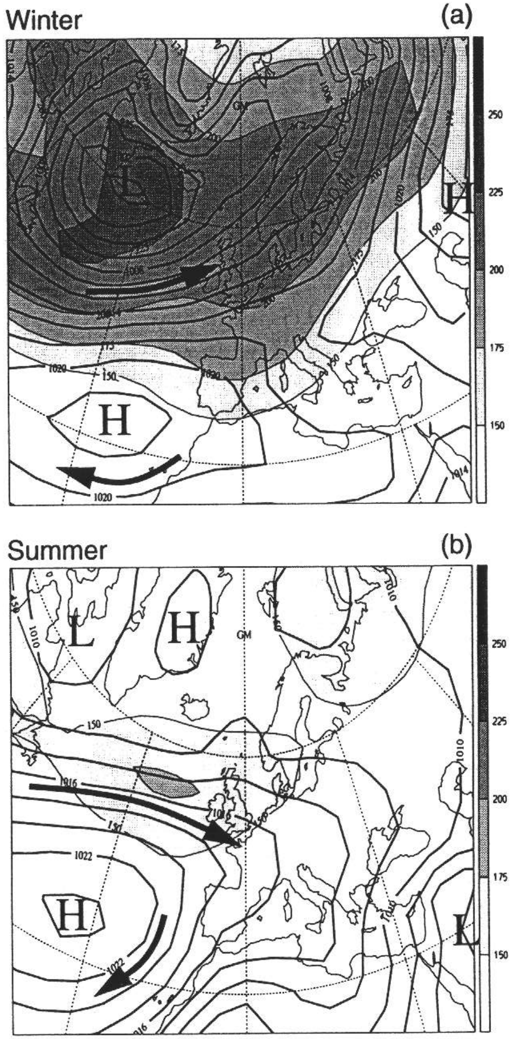

Figure 2.16 The time mean sea-level pressure pattern (contour interval 2 hPa) over the eastern Atlantic and the European sector in (a) winter and (b) summer, shown by the contours. The shading depicts a measure of the day-to-day variability of the upper tropospheric flow (the standard deviation of the 250 hPa surface in meters). (Based on data from Hoskins et al. 1989.)

southwest, the cold continental Siberian High to the east, and a weak Mediterranean Low feature to the south. In effect, the Alps are embedded in one of the slackest pressure gradient fields of the extratropical northern hemisphere. In summer (panel (b) of the figure), the Azores High extends northeastward to become the major pressure influence in the vicinity of the Alps.

The shaded patterns in figure 2.16 indicate the mean storm track. In winter, the majority of cyclones that influence Europe traverse the western Atlantic into northeastern Europe. Thus, the Alps are located south of the tail end of this storm track, and for the most part the weather is linked at the surface only to the cyclones' southernmost frontal features. The southward passage of these fronts around the Alpine massif itself is often a precursor of cyclogenesis to the lee, and thereby the central Mediterranean marks the entrance region of a weaker secondary storm track (Whittaker and Horn 1984; Buzzi and Tosi 1989). These cyclones' low-level orographic initiation results in their vertical structure's differing radically from that they exhibit elsewhere (von Ficker 1920; Tibaldi, Buzzi, and Speranza 1990). In summer, the Alps are further removed from the weaker and shortened Atlantic storm track, and in consequence fronts and their associated rain bands pass less frequently, and lee cyclogenesis is less frequent. Note also that changes in the incident fronts' frequency and strength would influence Alpine lee cyclogenesis and thereby exert a significant influence downstream affecting the entire Mediterranean region of southeastern Europe.

2.3.1.2 Variations on the Large-Scale Pattern

North Atlantic Patterns The North Atlantic Oscillation (NAO) relates to changes in the strength and location of the Icelandic Low and the Azores High on interannual and multidecadal time scales. One of the NAO's central features is that the changes in the two systems' intensity tend to be anticorrelated. A crude, one-parameter index for the change is the normalized mean surface pressure difference between Ponta Delgadas in the Azores and Akureyri in Iceland (see Rogers 1984), and its time series for the winter period shows an amplitude increase from 1890 to 1920, a subsequent decrease to around 1963, and thereafter a strong positive trend (see figure 2.17). Likewise, the Alpine surface pressure signal and that for the Icelandic region show a distinct inverse correspondence (see, e.g., Exner 1913, 1924).

Statistical studies examining the spatial structure of the anomalies in the large-scale sea-level pressure distribution indicate that the wintertime NAO signal is one of the two most recurrent low-frequency teleconnections—simultaneous variations at geographically separate regions—in the Northern Hemisphere. The NAO's north-south dipolar signal is present in the seasonal mean SLP at all seasons (Glowienka-Hense 1990; Rogers 1990) but it undergoes a seasonal variation in intensity, phase, and alignment. Other patterns identified applying the empirical orthogonal functions (EOF) technique to the surface pressure signal of the Atlantic-European sector include an east-

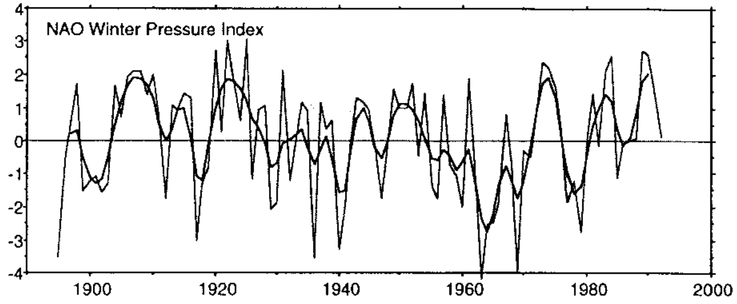

Figure 2.17 Winter index of the North Atlantic Oscillation (1895–1992) based on the mean normalized pressure difference between Ponta Delgadas, the Azores, and Akureyri, Iceland. The solid line represents smoothed data using a 1-4-6-4-1 low-pass filter. (Based on Rogers 1984; Bresch 1995; and data from Hoskins et al. 1989.)

west dipole and a dipole with centers located east of Iceland and in southern Europe (Rogers 1990). The NAO signal's amplitude in the Alpine region is rather weak, whereas those of the other two patterns are, respectively, about 4hPa and 8 hPa. These latter two patterns also resemble those derived using data for the eastern Atlantic and Europe (Glowienka-Hense 1990; Hagen and Schmager 1991; Fraedrich, Bantzer, and Burckardt 1993).

Consider now the climate imprint of the weather associated with the prevalence of these three surface signatures. For the NAO, it has been shown that high index values, and hence strong mean surface westerlies in the central Atlantic, tend to be accompanied by storm tracks extending further into northern Europe and cold (warm) thermal anomalies over Greenland (Scandinavia) and vice versa. Again an extreme, but not uncommon, flow setting is a strong negative anomaly of the NAO index associated with a persistent (over, say, a few weeks) blocking event in the North Atlantic resulting in a reversal of the latitudinal surface pressure gradient and an anomalous north-south high-low dipole. In winter, notable below normal temperatures in Scandinavia and enhanced precipitation over the Mediterranean usually accompany such an event (cf. Rex 1950, 1951; Moses et al. 1987; Lamb and Peppler 1987; Hurrell 1995). In contrast, the persistence of the other two patterns is not necessarily reflected strongly in the NAO index and yet can exert a significant impact on the in situ weather. Positive values of these signatures combine to yield stronger southwesterlies over northeast Europe, higher pressure over central Europe, a concomitant northward shift of the cyclone track, and a prediction of drier conditions. Contrary negative values would herald more westerly wind regimes penetrating into central Europe, a higher frequency of cyclone and frontal passages, and increased precipitation.

In terms of their cause and dynamics these variations of the atmospheric SLP signal have been linked to changes in sea surface temperature (SST) distribution of the North Atlantic (Deser and Blackmon 1993; Kushnir 1994).

On the interannual scale, the SST anomaly patterns are consistent with wind-driven atmospheric forcing. There is also some evidence, however, that SST anomalies in the western North Atlantic influence European weather on monthly timescales (Ratcliffe and Murray 1970; Palmer and Sun 1985). On decadal and longer timescales, there is a geographical and scale mismatch between the SST and SLP patterns. It has therefore been argued that the longer-term variations indicate a basinwide dynamical interaction between the atmosphere and ocean, with the ocean providing the principal forcing. On these long timescales, the SST signal has undergone a marked fluctuation during the last century, with negative anomalies in the 1920s, positive anomalies in the 1960s, and a return to colder conditions thereafter. The positive trend in the NAO index is consistent with the increase in the frequency of weather types with a southerly flow component, as noted in section 2.2.2. Again, on the associated timescales, an integral feature of the flow in the Atlantic is the overturning component, with dense water formation in the northern Atlantic being fed at the surface by water from the other oceanic basins and disgorging of the North Atlantic deep water to those basins at depth (see, e.g., Held 1993).

Clearly there is a complex chain of linkages to consider in this case. In effect the sequence takes the form global climate change ⇔ the interocean basin exchange ⇔ the Atlantic SST pattern ⇔ the atmospheric mean pattern in the North Atlantic ⇔ the characteristics of the in situ storm track ⇔ the frequency of the *Grosswetterlagen* ⇔ the response in the Alps. For instance, the positive trend of the winter NAO index noted above is consistent with the increased frequency of westerly circulation patterns (cf. section 2.2.2) and the increased frequency of warm and wet winters (cf. sections 2.2.1 and 2.2.3).

The El Niño/Southern Oscillation The El Niño/Southern Oscillation (ENSO) phenomenon is the dominant mode of interannual variability of the tropical ocean–atmosphere system. It is linked with anomalous SST patterns over a substantial portion of the tropical Pacific and with a shift in the main region of deep cloud-diabatic heating of the tropical atmosphere. ENSO induces strong, well-defined SLP and SST signals and anomalous weather across a broad belt of the tropics (Ropelewski and Halpert 1987; Halpert and Ropelewski 1992), and in addition, a teleconnection pattern extends as an arc-shaped, quasi stationary wave train over North America that tapers out over the southeastern United States.

The Atlantic-European sector is somewhat removed from both the forcing region for the El Niño and the tail end of the aforementioned wave train. However, perturbations of the large-scale pattern in the genesis region of the Atlantic storm track could, albeit intricately, influence the track's downstream features and thereby European weather. The nature of the influence(s) is related to the characteristics of the cyclones comprising the storm track and can vary with the strength and location of the tropical SST anomaly, the intensity of the midlatitude westerlies, and the prevailing settings of the SST and NAO in the Atlantic. Thus detecting an ENSO-European link and determining its dynamical nature is a challenging task. Nevertheless, there are indications of such a link (Van Loon and Madden 1981; Hamilton 1988; Kiladis and Diaz 1989; Fraedrich 1990; Fraedrich and Müller 1992; Wilby 1993; see also the overviews of Palmer and Anderson 1994; Fraedrich 1994). The effects include the forcing of anomalous winter weather patterns; modification in the frequency of cyclonic and anticyclonic *Grosswetterlagen*; signatures in the surface pressure (~ 1 hPa), temperature ($\sim 0.2^0$K) and precipitation (~ 10 mm); and changes in the intensity and location of the Atlantic storm track.

2.3.2 Alpine Effects on Weather and Climate

Alpine topography influences the atmospheric circulation by deflecting the flow horizontally and vertically, by introducing elevated sources and sinks of sensible and latent heat, and by inducing waves that propagate into the free atmosphere. These processes are mediated on a wide range of scales. On the meso-α-scale (horizontal scales in the range 200–2,000 kilometers), they include the retardation and modification of approaching synoptic systems and the formation of lee cyclones. On the meso-β-scale (20–200 kilometers), many of the flow phenomena (such as Föhn and valley winds) are associated with the generation and propagation of so-called internal gravity waves—which in turn are associated with the restoring effects of the atmosphere's stable density stratification. On the meso-γ-scale (2–20 kilometers), topographic effects include stationary mountain lee waves, deep convective clouds, and planetary boundary layer effects.

This section discusses some of the Alpine effects by first giving consideration to archetypal flow configurations, then in turn to the processes on the synoptic scale, the Alpine scale, the valley scale, and finally to aspects of orographic precipitation.

2.3.2.1 Archetypal Flow Configurations Although the real atmosphere is rarely so simple, consider a steady and uniform flow incident on an isolated topographic obstacle. This idealized setting illustrates some key aspects of flow past topography. For mountains with horizontal scales in the range of a few to about fifty kilometers, effects associated with the atmosphere's stable stratification dominate the dynamical response. The stratification tends to inhibit vertical displacement and thereby reduces the air parcels' ability to flow over the mountain. This feature gives rise to two distinct flow regimes, which figure 2.18 depicts. In the first regime (panels (a) and (b)), the flow is over the obstacle, and a pattern of vertically propagating gravity waves is established (see Smith 1979; Durran 1990). Even at upper levels, air parcels experience vertical excursions over a depth that is comparable to the mountain height.

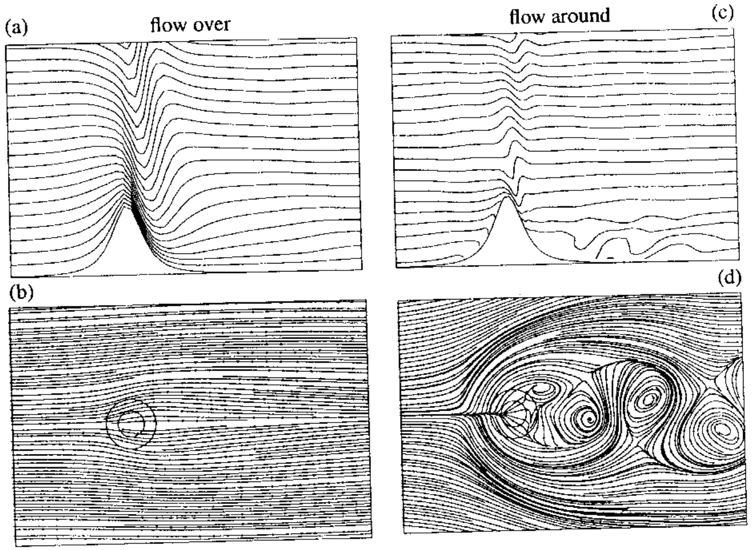

Figure 2.18 Nonrotating flow past an idealized isolated mountain. The flow is from left to right, and the figure shows both the flow-over regime (panels (a) and (b)) and the flow-around regime (panels (c) and (d)). Panels (a) and (c) show the instantaneous streamlines in a vertical section across the mountain top, and panels (b) and (d) show the same features on the surface. (From Schär and Durran 1997.)

If stratification effects become stronger, or if the upstream wind speed is smaller, the air parcels are unable to surmount the orography (panels (c) and (d) of figure 2.18). The flow splits upstream, and the air parcels make broad, quasi horizontal excursions as they tend to circumscribe the mountain rather than traverse it. For this configuration, there is still some gravity wave activity in the upper portion of the obstacle, but the vertical extent of the associated oscillations is significantly reduced compared to those of the flow-over regime. To the lee, a wake is established that can become unstable and undergo a transition toward vortex shedding (Schär and Smith 1993; Schär and Durran 1997). In the case of isolated topographic obstacles (such as mountainous islands), these effects can imprint on satellite pictures as spectacular Kármán vortex streets (Etling 1989). In the Alpine context, vortex shedding is not observed, but edge vortices have been identified during periods of strong flow, both in observations and numerical experiments (Steinacker 1984b; Thorpe, Volkert, and Heimann 1993; Aebischer and Schär 1998). There are also indications that these features can contribute to the formation of Alpine lee cyclones.

The transition between the two archetypal flow regimes shows some of the characteristics of a bifurcation (Smith and Grønås 1993; Smolarkiewicz and Rotunno 1989) and can result from small changes in the upstream conditions. Here the governing control parameter is related to the mountain

height and geometry, the static stability, and the approaching airstream's speed. On the scale of the Alpine massif, the effects of the earth's rotation must in addition be accounted for. Numerical experiments have revealed that rotational effects relevant for large obstacles tend to favor the flow-over regime (Trüb and Davies 1995). The flow past the Alps is hence less likely to be around than the flow past a smaller-scale topographic feature of the same height. It then follows that the flow past the Alps can be over on the alpine scale but at the same time around on the scale of individual massifs and mountains.

An evaluation based on typical Alpine upstream parameters also indicates, in agreement with observational studies (Binder, Davies, and Horn 1989; Chen and Smith 1987), that both the flow regimes of figure 2.18 can occur on the Alpine scale. It is important to represent this feature in studying the Alpine climate, since flow-over and flow-around result in very different distributions of key climatic variables such as temperature and precipitation. In this regard, it should be noted that the implicit smoothing of the topography in low-resolution numerical models (such as global climate models) strongly shifts the response toward the flow-over regime.

2.3.2.2 Effects on Approaching Synoptic Systems and Lee Cyclogenesis The Alps are most effective in modifying the ambient circulation when the flow is roughly perpendicular to the main Alpine ridge, that is, from the north or south. These flow situations usually occur in conjunction with the approach of a trough or low-pressure system from the west or northwest, toward central Europe or Scandinavia, a configuration that is particularly common during autumn, winter, and spring (cf. section 2.3.1). Many of the approaching systems are close enough that their fronts come into direct contact with the Alps. When a frontal system impinges on the Alps, the dynamical processes can no longer be viewed as the modification of a quasi steady airstream, but rather represent a complicated three-dimensional and time-dependent flow evolution (for a concise review, see Egger and Hoinka 1992), which often induces transitions from flow-over to flow-around and vice versa. Changes in the wind speed and stratification of the incoming air mass trigger such transitions, consistent with the discussion of the archetypal flow regimes in the previous section.

Ahead of approaching cold fronts there is a strong southwesterly flow toward the Alps (see figure 2.19a). This airstream consists of warm, moist air, is comparatively weakly stratified, and has a high wind speed. It is thus able to pass over the Alpine ridge, leading to South Föhn in the north. In contrast, the cold air behind the front is stably stratified and often capped by a pronounced inversion. It is thus unable to climb the Alps but rather deflected laterally. In effect this leads to retardation and deformation of the incident low-level cold front, and the lateral deflection of the northerly flow induces Mistral and a pronounced cold-air outbreak to the west (figure 2.19b) and Bora to the east of the Alps (figure 2.19c). At the same time, a lee cyclone

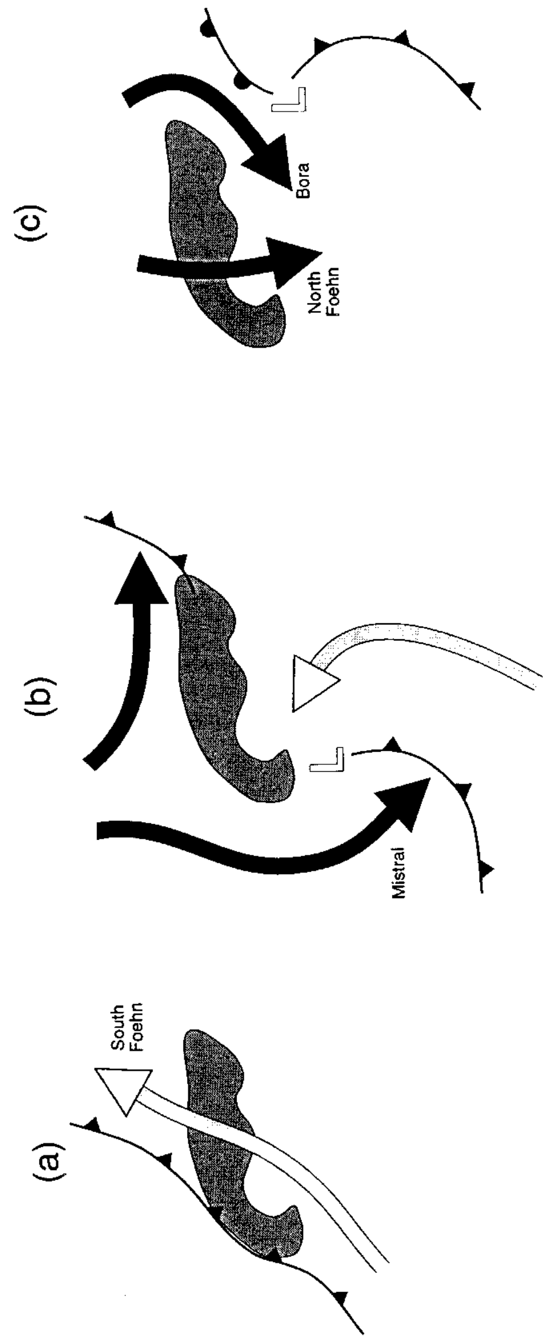

Figure 2.19 Schematics of the interception of a cold front by the Alps. The three phases correspond to (a) deformation of cold front and onset of South Föhn, (b) cold-air outbreak into the western Mediterranean (Mistral) and formation of lee cyclone, (c) eastward progression of lee cyclone and onset of Bora and North Föhn (adapted from Smith 1986.)

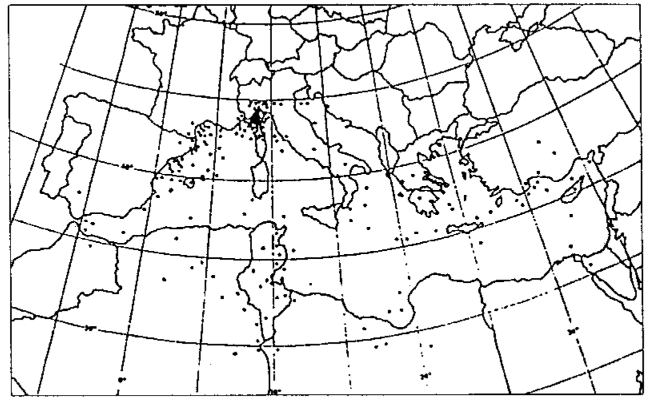

Figure 2.20 Climatological distribution of winter cyclogenesis in the Mediterranean region. (From Reiter 1975.)

may form over the Gulf of Genoa. Finally, as cold air piles up to the north, the associated cold-frontal inversion is lifted, and the cold air underneath spills over the mountain passes, inducing North Föhn. The circulation associated with the eastwards propagating lee cyclone also supports the formation of North Föhn and Bora winds.

Synoptic conditions are most suitable for lee cyclogenesis when a deep upper-level trough and its associated surface cold-front enter the Alpine region (Bleck and Mattocks 1984; Tafferner 1990; Tibaldi, Buzzi, and Speranza 1990). On the average, about thirty Alpine lee cyclones form per year, most frequently during the spring and the autumn. Figure 2.20 shows the climatological distribution of lee cyclogenesis. The concentration of events over the Gulf of Genoa near the southwestern tip of the Alps betokens an orographic influence. Once a lee cyclone has formed, it becomes an important governor of the regional weather and climate. Lee cyclones often attain maximum strength over northern Italy, and the associated strong southerly flow ahead of the cyclone can induce storm surges in the Adriatic sea (which occasionally threaten Venice) and advect moist Mediterranean air toward the Alps (which substantially contributes to the annual precipitation totals in northern Italy and the eastern Alps). Further downstream, lee cyclones propagate toward the east, following one of two major tracks (Whittaker and Horn 1984): One leads across the eastern Alps into eastern Europe, the other follows the northern border of the Mediterranean Sea. As a result, Alpine lee cyclones are crucial to the precipitation in much of the eastern Mediterranean region. Since lee cyclogenesis is sensitive to the meridional location of the midlatitude storm track (as well as to other factors), Mediterranean precipitation in the event of global climate change is thus difficult to assess.

2.3.2.3 Regional Wind Systems Several of the regional Alpine wind systems possess some characteristics of the archetypal flow configurations (figure 2.18) and occur transiently during episodes of frontal passage (figure 2.19). Key aspects of North and South Föhn (cf. Seibert 1990) are evident in the stream crossing the mountains. The densely packed isentropes in the vertical sections of figure 2.18a are indicative of the Föhn region, which is associated with low pressure and increased wind velocities. A lee-side warming results from latent heat release along the parcel's trajectories, and—presumably more importantly—from adiabatic warming through net descent if the parcels at the lee slopes stem from some higher level upstream. Downslope wind storms develop when the gravity wave response undergoes a nonlinear amplification involving the transition to supercritical-like flow (Smith 1985; Durran 1986), gravity-wave breaking (Clark and Peltier 1977), or both. The latter process is associated with the overturning of density surface, yields a configuration with dense fluid above lighter fluids, and induces clear air turbulence through the breakdown of this unstable configuration. The elongated shape of the Alps significantly facilitates the development of Föhn-like flows and makes them more two-dimensional both in cases of northerly (North Föhn) and southerly (South Föhn) flow. The upstream deceleration can then be interpreted as the blocking of the incident upstream flow (Pierrehumbert and Wyman 1985). The deep inner-alpine valleys experience particularly strong Föhn flows. To some extent this results from gap and channelling effects (cf. Wippermann 1984), but it is presumably also related to the reduced horizontal scale on the valley transects, which makes the flow more susceptible to gravity-wave effects as opposed to the effects of Earth rotation.

Mistral, Bora, and the Bise (for a description of these flows, see respectively Pettre 1982; Smith 1987; and Wanner and Furger 1990) can be associated with streams of air deflected horizontally by the Alpine massif (panels (c) and (d) of figure 2.18). In these cases, some of the stratification is often concentrated in a inversion layer that separates the cold air below (which flows around the Alps) from the potentially warmer air aloft (which flows over the Alps). Mistral and Bora usually occur when the cold air behind a cold-frontal passage is horizontally deflected around the western and eastern portions of the Alps, respectively (figure 2.19). The Bise results from the channeling of north-easterly flow between the Alps and the Jura mountains. The specifics of the local topography as well as other factors modify each of these wind systems. For instance, the Bora is on the one hand associated with a broad excursion of cold air around the Alps and yet flows itself over the northern portion of the Dinaric Alps and thereby induces high winds. The predominant flow configurations determine the major thermal anomalies of the Alpine climatology (see section 2.2.1). In the major Föhn valleys to the north of the Alps, South Föhn occurs on as many as 15 percent of the days in the month, with maximum frequency during the spring and autumn. During Föhns, the local surface air temperatures may increase by more than

10 K, and the combined effect of this Föhn warming on the longer-term local climatological mean is pronounced. The most significant factor on the regional scale, however, is related to the shielding of the lee against upstream influence. If the flow is around the Alps, it leaves a pool of almost stagnant air in its wake that greatly reduces the importance of advective effects. The prolonged residence time of air parcels in the Alpine wake then allows local boundary layer and radiative processes to determine the lower atmosphere's thermal structure. These processes, which are also affected by the orographically controlled distribution of cloud cover, lead to a pronounced climatological temperature contrast across the Alpine main crest and also make the often stagnant air mass in the Po valley susceptible to air pollution.

2.3.2.4 Thermally Driven Mountain Circulations In addition to influencing the structure and development of preexisting flow systems that impinge upon the Alpine topography, the terrain is also associated with and responsible for the in situ excitation of daily periodic circulations on scales varying from individual slopes and valleys to the massif itself.

Figure 2.21 sketches the daytime and nighttime patterns of the slope and valley circulations as typically observed in the Alpine region during clear-sky conditions (see, e.g., Urfer-Henneberger 1970; Oke 1987). Over individual slopes, the response to diurnal radiative variation takes the form of upslope winds during the day and downslope winds at night (see, for example, Defant 1949; Egger 1990). These slope wind circulations are accompanied by and coupled with low-level airflow up and down the valleys, referred to respectively as valley and mountain winds (see, e.g., Wagner 1938). The driving mechanism for these flows is directly related to the cycle of daytime solar heating and nighttime radiative cooling of near-surface air layers.

Again on the scale of the main Alpine ridge itself, there is some indication of a diurnal circulation between the ridge and the adjacent foreland regions (Kleinschmidt 1922; Burger and Eckhart 1937), with low-level inflow (drainage) during the day (night) and a weak return flow above crest height. A larger diurnal temperature cycle in the inner Alpine region as compared to that over the flatland (cf. section 2.2.1) and periodic, Alpine-confined pressure anomalies (Scherhag 1966; Hafner et al. 1987) during fair weather support the existence of such a circulation. These circulation systems have been attributed to two effects, either singly or in combination: The terrain provides an elevated heat source (sink) effect during the day (night) (see, for example, Flohn 1953), and the more rapid heating (cooling) of the inner-Alpine region associated with the lesser volume of air per unit horizontal area in the major Alpine valleys compared with the surrounding plain (the so-called volume effect; see, e.g., Steinacker 1984a; Whiteman 1990) enhances thermal response in the region.

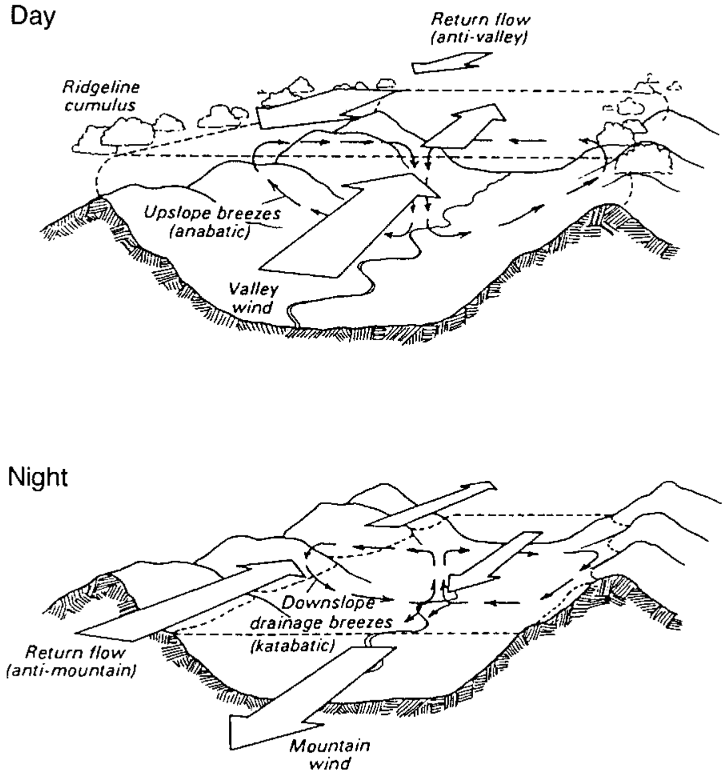

Figure 2.21 Schematics of the thermally driven circulations in a mountain valley during the day and at night. The view is up valley. (From Oke 1987.)

The mountain-foreland circulations can trigger or contribute to various flow phenomena characteristic of Alpine climate. For example, during the warm season, they contribute to establishing the preferred regions for convective activity and cloud formation (cf. Barry 1992; Banta 1990). Likewise in the cold season, the nighttime cold air drainage contributes to the buildup and maintenance of the persistent inversion layers and fog frequently experienced in the Swiss middle-land (see, for example, Wanner 1979).

2.3.2.5 Orographic Precipitation and Convection Seasonal or yearly precipitation totals generally tend to increase with height. Section 2.2.3 reviewed some of the corresponding literature for the Alpine region, and Smith (1979) and Kahlig (1986) offer comparative reviews covering a range of topographic obstacles. The precipitation-height relationship is in general far from linear, and hills as small as fifty meters can enhance precipitation by as much as 100 percent (Bergeron 1968; see also Browning 1980). The

seeder-feeder mechanism (Browning 1985) is generally accepted to be of prime importance in this context. At low levels, topographic lifting leads to condensation and the formation of small, usually nonprecipitable water droplets. Appreciable precipitation can result when falling precipitation from aloft, induced either by orographic lifting or a preexisting synoptic-scale disturbance, collects these droplets. With the help of the seeder-feeder mechanism, the orographic enhancement of precipitation pertains to a deep layer (on the order of several kilometers) that can far exceed the mountain height.

The discussion of airflow regimes in section 2.3.2.1 suggests a significant sensitivity of the rainfall distribution and intensity to the predominant flow regime. Smith (1989) has documented such a sensitivity for the islands of Hawaii, but we are unaware of a corresponding systematic study for the Alpine region. If the flow is over the Alps, upslope precipitation occurs, and the low-level moisture is essentially extracted from the ambient flow. If the flow is instead around, low-level moisture is deflected laterally. Both flow regimes result in a shielding of the lee from precipitation, but the situation upstream is highly dependent upon the flow regime. Some of the case-to-case variability in upstream precipitation is likely a result of this effect. Precipitation signals for frontal passages along the Gotthard section (Phillips 1984) include both cases with light (for example, March 4, 1982, with precipitation totals of a few millimeters) and very heavy precipitation (such as April 29–30, 1982, with totals of about 100 millimeters), and the corresponding flows have been classified as around and over, respectively (Chen and Smith 1987).

An important aspect of Alpine precipitation is the comparatively frequent occurrence of widespread heavy and long-lasting precipitation and flooding events to the south of the Alps. Such events are most frequent during autumn, when there is an ample supply of warm, moist air over the Mediterranean Sea. The typical synoptic setting is associated with moist and warm southerly flow ahead of a cold front, deep trough, or lee cyclone. Since the incident air mass is weakly stratified, this southerly airstream is able to flow over the main alpine crest (cf. figure 2.19a). To the north, Föhn-like effects result, whereas lifting and adiabatic cooling to the south result in condensation and heavy precipitation. If such a synoptic situation persists for several days, devastating flooding can result. The improved understanding and forecasting of heavy precipitation in the Alpine region is currently one of the primary objectives of a large-scale international field campaign (cf. Binder and Schär 1996), and an assessment of the associated flooding potential is a vital aspect of climate change research in the Alpine region. Although some promising numerical results are available at intermediate computational resolution (e.g., Buzzi et al. 1995; Binder and Rossa 1995), the detailed simulation of flooding episodes might require very high horizontal and vertical resolution, because the precipitation processes might rely on the interaction of the topographic gravity-wave signal with the embedded convection.

The discussion above pertains mostly to stably stratified weather regimes. Even these regimes often have embedded convection, related either to frontal activity or to the generation of instability through ascent (see Smith 1979). The horizontal scale of convective updraughts is only a few kilometers and eschews the laminar description of the flow as given in section 2.3.2.1. Moist convection is particularly important in summer, when it provides the dominant contribution to the rainfall totals. The Alps can influence the flow in such situations either by inducing low-level convergence of moist air or by conditioning the local profiles of temperature, humidity, and wind. The latter effects include processes associated with the Alpine thermal anomaly as well as wind channeling effects along the Alpine foothills (cf. Binder, Davies, and Horn 1989; Houze et al. 1993). Though the Alps' thermal anomaly can itself be instrumental in inducing convection (e.g., air mass thunderstorms), most severe weather occurs when an active triggering takes place, usually in conjunction with an approaching synoptic system. Observational evidence for this type of triggering is available for convective systems both to the south (Buzzi and Alberoni 1992; Cacciamani, et al. 1995) and north of the Alps (Schiesser, Houze, and Huntrieser 1995; Huntrieser et al. 1997). The associated synoptic systems are important in both providing moisture supply and inducing ascent.

Precipitation in the Alpine region is a particularly important climate parameter, and not only for its relevance to ecological and economical systems. Because precipitation results from a chain of complex and highly nonlinear processes, it is very sensitive to external parameters and to the large-scale flow. Global warming could affect substantially the frequency and distribution of precipitation. Even a modest warming will significantly increase the air's potential for transporting water vapor. Current GCM numerical simulations of the 2 × CO_2 climate indeed suggest the global concentration of water vapor has increased on average by about 10 to 30 percent (Houghton, Jenkins, and Ephraums 1990), and regional climate simulations indicate that this could significantly increase precipitation, particularly in mountainous and coastal regions and during the synoptically active seasons (Schär et al. 1996; Frei et al. 1997).

2.3.3 Feedbacks to the Larger-Scale Flow

Several of the mesoscale processes referred to in the previous sections can feed back through nonlinear processes to the larger-scale flow and thereby affect the atmosphere's synoptic and finally planetary-scale circulation. Little is known about the Alps' relative contribution, but there can be no doubt that the earth's topography as a whole is key in defining the climatic zones' geographic distribution. Charney and Eliassen (1949) mooted that the stationary planetary-scale waves are "anchored" to the major topographic obstacles, but the extent to which topography controls this process has only

recently become evident from general circulation numerical experiments. Figure 2.22 reproduces a spectacular example from such an experiment by Broccoli and Manabe (1992). The figure shows the results of a pair of global climate simulations. In both experiments, the geographical distribution of land and sea is prescribed, but the topography has been removed in the simulation shown in the right-hand panels. The bottom panels depict the simulated geographical distribution of wet and dry climatic zones on the Northern Hemisphere spring. The local effects of topography comprise precipitation anomalies (e.g., in the Alpine region) and shadow effects (which may be inaccurately represented because of the employed model's low resolution). In addition, there are remarkable effects far removed from mountains. For instance, large parts of Siberia are classified as dry in the topography run (in agreement with the observed climatology) but would experience substantially larger rainfall amounts in the absence of topography. The top panels in figure 2.22 imply that the control topography exerts is indeed through planetary-scale standing wave patterns and their embedded storm tracks.

The results of Broccoli and Manabe, which are supported by other numerical experiments, entail substantial up scale contributions, meaning that the response's horizontal scale significantly exceeds that of the topography. Such effects rely on the generation of flow anomalies through irreversible processes that can in turn quasi permanently alter the downstream conditions. The overall scale of the response is then governed by advective and dissipative processes downstream rather than by the scale of the topography itself. Dynamical instabilities, in the atmosphere primarily due to barotropic and baroclinic processes, can in addition amplify and affect the structure of orographic wakes.

There are several orographic mesoscale processes of the above category, including the dissipation of internal gravity waves at upper levels (gravity-wave drag), topographically induced condensation (orographic precipitation), processes associated with flow splitting and surface friction (formation of shear lines and vortices to the lee), and the formation and subsequent decay of an orographic cyclone (lee cyclogenesis). General circulation models cannot resolve several of these processes, which must be parameterized instead. The most important example is gravity-wave drag, considered an absolutely essential contributor to the planetary-scale circulation (Davies 1986; Palmer, Shutts, and Swinbank 1986; McFarlane 1987) as well as one of the key factors in forcing the planetary-scale quasi-stationary waves (figure 2.22, top diagrams). These waves not only control the distribution of climatic zones but are in addition highly sensitive and responsible for a considerable portion of the interseasonal and interannual variability in the Northern Hemisphere (e.g., Wallace, Zhang, and Lau 1993; Lau, Sheu, and Kang 1994). Currently the limited knowledge about these waves' behavior in a changed climate is of some concern (cf. Held 1993).

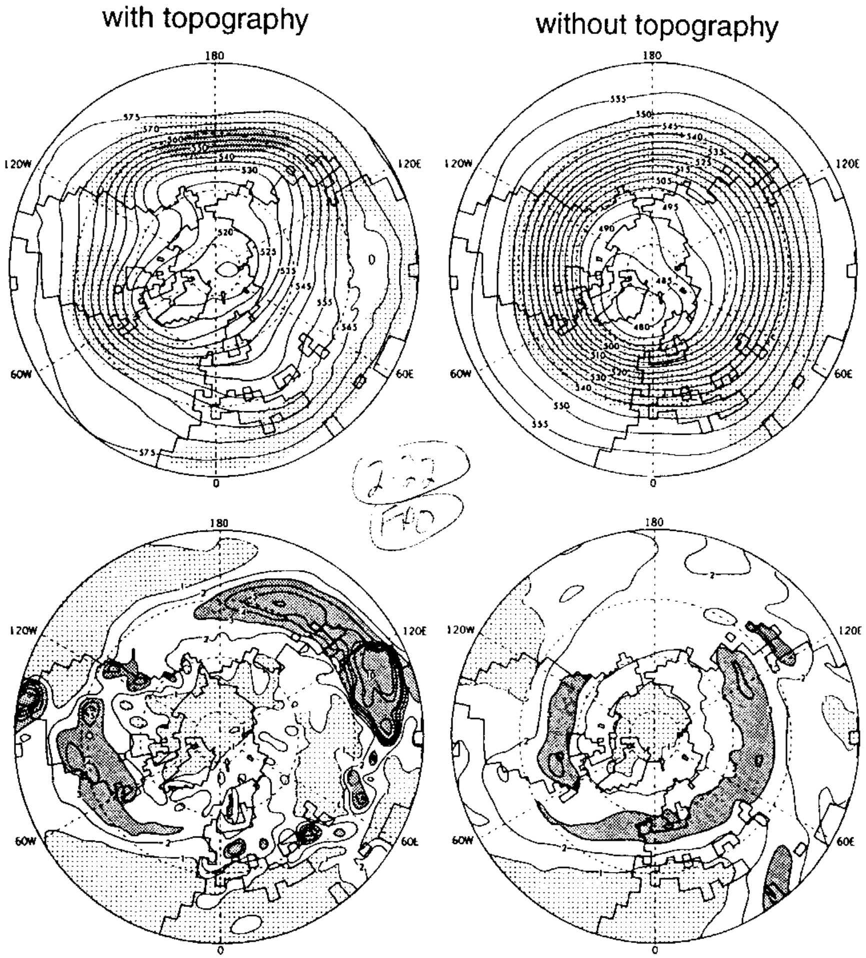

Figure 2.22 Mean spring circulation and precipitation from global numerical simulations with topography (left) and without topography (right). Top panels show the 500 hPa geopotential height in decameters. Light and dense stippling indicates winds greater than 12 ms^{-1} and greater than 24 ms^{-1}, respectively. Bottom panels show precipitation rates [mm d^{-1}]. Contours are given at 1, 2, 3, 4, 5, 6, 8, 10, 15, 20 and 30 mm d^{-1}. Light stippling indicates dry regions with precipitation less than 1 mm d^{-1}; dense stippling indicates wet regions with precipitation greater than 3 mm d^{-1}. (From Broccoli and Manabe 1992.)

2.4 FURTHER REMARKS

Details of the present-day climate were summarized using the available observational data and discussed in terms of the underlying dynamical processes. This approach highlighted the Alps' influence on the region's distinctive weather phenomena and climate. It also served to pinpoint both the incompleteness of our climatological database and the shortfalls in our understanding. From the observational standpoint, there is a pressing need to develop from the data already existing more comprehensive, accurate, homogenized, Alpine-wide data sets of the basic climatic elements. Likewise, the spatial structure and temporal development of many key Alpine-related flow systems have yet to be described adequately, and rectifying this will require specifically designed field experiments. From the standpoint of understanding it was emphasized that the geometry and height of the Alps rendered the dynamics of the orographically induced phenomena complex, nonlinear, and often multiscaled. This makes their study particularly challenging, but nevertheless recent theoretical developments and the availability of powerful computers has opened the way for significant progress.

The understanding of current Alpine climate forms a sound foundation for studies both of the Alpine region's paleoclimatic history and its fate in the event of a global climate change. Subsequent chapters will consider these aspects.

GLOSSARY

This section is partly based on the definitions given in the glossary produced by the United Kingdom Meteorological Office (Meteorological Office 1991).

adiabatic: An adiabatic (thermodynamic) process is one in which heat does not enter or leave the system. The adiabatic transport of an air parcel can nevertheless lead to temperature changes as a result of expansion and compression. An example of a nonadiabatic (diabatic) atmospheric process is latent heating through condensation of water vapor.

advection: Transfer of mass (or of an air mass property) by horizontal or vertical winds.

albedo: The fraction of incoming solar radiation reflected by the Earth's surface or the cloud cover.

baroclinic: An atmosphere that entails quasi-horizontal temperature gradients is said to be baroclinic. A process that depends on this property is also said to be baroclinic.

Bise: A cold, dry northeasterly wind to the north of the Alps.

blocking: In a blocked midlatitude westerly flow, a quasi-stationary high-pressure system interrupts cyclones' usual eastward progression. In a blocked circulation, the upper-level westerly flow is split into two branches.

Bora: A cold northeasterly wind that blows down from the Dinaric Alps onto the eastern Adriatic coast.

climatological mean: The average distribution of some climatic element (for example, temperature) over a climatological period (say, thirty years).

convection: Vertical heat transfer by rising warm and sinking cold air. Convection in the atmosphere often occurs in the form of moist convection, which involves condensation and convective cloud formation in the ascending branches.

downscaling: To infer the regional-scale climate from larger-scale information. The term is often used in conjunction with climate scenarios, where it relates to the construction of regional or local climate scenarios from larger-scale information.

empirical orthogonal function (EOF): The variability of some field can always be expanded as a superposition of orthogonal base functions. Empirical orthogonal functions are a special type of orthogonal base functions objectively constructed so as to describe the observed variability with as few base functions as possible.

Föhn: A warm, dry wind to the lee of a mountain range. The term originated from the Alpine South Föhn but is used as a general term of this type of winds.

gradient: The slope of a function or field. In the case of a topographic field, the gradient always points in the direction of the steepest uphill slope. In general, the gradient is a vector.

gravity wave: A type of wave that depends for its existence on the restoring force buoyancy provides. In a stably stratified fluid such as the atmosphere, the vertical displacement of fluid parcels (e.g., by topographic lifting) generates internal gravity waves.

inviscid: Not affected by viscosity.

isentropes: Surfaces of constant potential temperature. The adiabatic motion of an air parcel always follows isentropic surfaces.

Kelvin [K]: A temperature scale that starts at absolute zero (−273.15 °C). To convert temperatures from degrees Celsius to Kelvin, add 273.15.

latent heat: The quantity of heat absorbed or emitted during a change of state. In the atmosphere, latent heating associated with condensation (evaporation) and freezing (melting) of water is important for cloud formation and for the transport of energy.

lee cyclogenesis: The formation of a cyclone (lee cyclone) to the lee of a large-scale mountain. Lee cyclogenesis to the lee of the Alps is often in the region of the Gulf of Genoa.

low-pass filter: A mathematical operation (or electronic circuit) that filters a time series so as to remove its high-frequency components.

mesoscale: The horizontal scale appropriate to atmospheric systems of a size between that of individual cumulus clouds (approximately 2 kilometers) on the one hand and that of major depressions and anticyclones (approximately 2,000 kilometers) on the other.

Mistral: A northerly wind that blows along the lower Rhone valley and offshore along the Mediterranean coast.

North Föhn: A dry northerly wind to the south of the Alps.

orography: A term used in meteorology to signify the variation of the height of the ground above sea level. It is synonymous with topographic height.

planetary scale: The scale appropriate to atmospheric features with horizontal scales comparable to that of the planet (several thousand kilometers).

planetary waves: Meandering waves of the midlatitude westerly flow with horizontal wave lengths typically of about 5,000 kilometers.

potential temperature: The temperature an air parcel would possess if it were adiabatically compressed (expanded) to the standard pressure, 1,000 hPa. Unlike temperature, potential temperature increases with height in most of the atmosphere.

storm track: A region with a high frequency of cyclonic activity, located along the typical tracks of low-pressure systems. The major storm tracks of the Northern Hemisphere are the Atlantic and Pacific storm tracks.

stratification: In an unstratified fluid (neutral stratification) the density is constant. In a stratified fluid (stable stratification), the density decreases with height. Most layers of the atmosphere are stably stratified.

stratosphere: The region of the atmosphere above the troposphere and below the mesosphere. It extends from the tropopause (approximately eleven kilometers in midlatitudes) to the stratopause (approximately fifty kilometers).

synoptic scale: The scale appropriate to atmospheric features related to low- and high-pressure systems (approximately 1,000 to 4,000 kilometers).

teleconnection: A connection (relation) between climate variations or anomalies at locations separated by distances greatly in excess of the normal synoptic scale.

troposphere: The lower region of the atmosphere, extending to about sixteen kilometers near the equator, eleven kilometers in midlatitudes, and nine kilometers near the pole. The troposphere is capped by the tropopause. Most clouds are located within the troposphere.

trough: A trough (of low pressure) is a pressure feature in the synoptic weather chart. It is characterized by isobars shaped like a trough.

Universal Time (UT): Western European time zone, formerly Greenwich mean time (GMT).

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