CLIMATOLOGY OF THE INTERIOR COLUMBIA RIVER BASIN

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INTRODUCTION

The Columbia River Basin is in a transition-type climate zone. It is influenced by three distinct air masses: 1) moist, marine air from the west that moderates seasonal temperatures; 2) continental air from the east and south, which is dry and cold in winter and hot with convective precipitation and lightning in summer, and 3) dry, Arctic air from the north that brings cold air to the Basin in winter and helps cool the Basin in summer.

The timing and extent of competing air masses is controlled largely by synoptic weather patterns and local terrain features that vary across the Basin. Prolonged periods of drought occur when Pacific storms are deflected around the region, preventing the intrusion of moist, marine air. At these times dry, continental conditions prevail. Damaging frosts and freezing conditions commonly occur when Arctic air invades the Basin before winter hardening in autumn or after budbreak in spring. Cold damage also may occur in winter if a warm, marine intrusion is followed by a sweep of Arctic air. In addition, the unique interplay between air mass types result in dramatic changes during transition. The most unique of these transitions is rain-on-snow flooding that occurs when warm, wet marine air displaces cold, Arctic conditions in winter. Lightning and gusty winds also occur during transitions between continental and marine air masses, mainly in spring and summer.

Section C of this chapter discusses extreme climate patterns that cause ecological disturbances in detail. This section concentrates more on averages and trends of climate in the Basin. In addition, some circumstances that control regional climate patterns are discussed.

PREVIOUS CLIMATE STUDIES

Every decade new normals (30 year averages) from thousands of weather observing stations throughout the U.S. are computed as part of an international effort led by the World Meteorological Organization. These normals are summarized in several documents available from the United States Department of Commerce, National Climatic Data Center (NCDC) in Asheville, North Carolina. The summaries list normals, means, and extremes for National Weather Service first-order and Cooperative observation stations that have long-term weather records. Also, they include a narrative describing important features of distinct climate zones and summaries of climate divisions within each state. In addition to these official summaries, others have summarized various aspects of weather and climate in the Columbia River Basin and western U.S. (for example, Rumney 1968; Pacific Northwest River Basins Commission, Meteorology Committee 1968, 1969a, 1969b; Phillips 1962).

An interesting analysis of equivalent potential temperature found two major frontal zones in the Columbia River Basin (Mitchell 1976). A Pacific air mass boundary, which dominates during summer, extends diagonally across the Basin from northwest California to northwestern

Montana. Relatively moist, marine air exists north of the boundary and drier, continental air is common south of the boundary. Mitchell noted that the Pacific boundary coincides with the eastward extension of coastal vegetation found in northern Idaho and northwestern Montana. A westerlies-anticyclone boundary, which dominates during the winter, was found to stretch east-west along the Oregon-Nevada border and across northern Utah. It marks the boundary between a region of prevailing westerly winds in the north and southerly winds in the south, caused by circulation around a persistent center of high pressure (anticyclone) over southern Nevada. This frontal zone also coincides with the northern or southern extent of several tree species.

Bryson and Hare (1974) noted that precipitation variability in the western United States was caused by numerous small-scale climatic controls (primarily the result of complex topography) embedded within large-scale climate controls. Mock (in press) was able to determine the hierarchy of climatic controls that operate at different spatial scales by analyzing month-to-month precipitation patterns. He found that while large-scale climate controls (like the polar jet stream, Pacific subtropical high, and subtropical ridge) play important roles in precipitation variability, small-scale climate controls (like complex topography, thermal troughs, confined mixing heights, and convective systems) can dominate. In particular, winter precipitation is dominated by orographic lifting of strong westerlies associated with the mean position of the polar jet stream over Oregon and Washington. Also, marine air flows through the Columbia gorge into the Snake River valley and other valleys adjacent to and oriented toward the oncoming Pacific air mass. Some winter precipitation in the northern Rocky mountains of Montana also may be upslope induced by polar anticyclones (Boatman and Reinking 1984). During spring a thermal trough migrates north from California. When combined with a marine push, convective precipitation and thunderstorms are possible, especially in Oregon and southern Idaho. In Washington, northern Idaho, and northern Montana, Pacific storms remain influential during spring. During summer the polar jet stream usually is too far north to bring many storms into the Basin. The thermal trough and intermittent marine pushes continue to play a role. Strong convection in southeastern Idaho and western Montana continue through the summer but lack moisture because the prevailing northwesterly flow aloft is relatively dry.

DATA

Observation Data

Recorded weather observations began in the western United States during the late 1800s. Because population was relatively sparse at the time, only 8 stations within the Interior Columbia River Basin have records approaching 100 years. These are Spokane WSO, Washington; Dufur, Oregon; Fortine, Kalispell WSO, and Haugan, Montana; and Priest River Experiment Station, Caldwell, and Saint Ignatius in Idaho. Daily summaries of precipitation and temperature from these stations have been adjusted for station relocations, changes in instrument heights and types, changes in observing times, and increases in urbanization as part of the Historical Climatology Network (HCN, Karl and others 1990, Hughes and others 1992).

Another source of daily precipitation and temperature measurements is maintained by the National Weather Service, Cooperative Observer Network (COOP, National Climatic Data Center 1991). Data from COOP stations provide the highest spatial resolution of daily measurements. Because many observers in the COOP network are volunteers, however, consistency and quality can be somewhat lacking. These data have undergone some quality control, but not as completely as HCN data. The data need sorting in order to locate those sites with reliable records.

High quality measurements of multiple parameters are available from the National Weather Service first-order stations that are operated by trained observers. Hourly temperature, dew point, relative humidity, wind, precipitation, and radiation data from these sites are available through the Solar and Meteorological Surface Observation Network (SAMSON) data set developed by the National Renewable Energy Laboratory (1992). In addition, the hourly data are summarized each day and added to the COOP database.

The HCN and SAMSON observation sites usually are located near population centers or airports away from the wild land areas of forests and mountains. The COOP sites have more diverse locations but most are near homes or businesses that commonly exist in valleys or away from wild land areas.

Significant additions to high elevation data occurred in the mid 1930s and late 1970s. The U.S. Department of Agriculture, Soil Conservation Service (now Natural Resources Conservation Service) added many new snow course observations in the mid 1930s (at the height of a nation wide drought) to measure snow depth and snow water equivalent near the headwaters of major river basins (USDA, Soil Conservation Service 1988). Snow course measurements are acquired about every month during winter and spring. In the late 1970s an automatic system, SNOTEL, was added. SNOTEL sites transmit current temperature, total precipitation, and total snow water equivalent approximately once each day (Barton 1977). SNOTEL sites commonly are placed near the head of drainage basins. In the late 1970s remote automated weather systems (RAWS) also were being installed by the USDA Forest Service and Bureau of Land Management (Redmond 1991; U.S. Department of the Interior, Bureau of Land Management 1995). These stations are designed to support fire weather forecasting so they operate mainly during summer and are located in forest clearings on hill slopes and ridges. RAWS stations transmit hourly information on air temperature, precipitation, fuel temperature, relative humidity, and wind. There are about 200 RAWS, 200 SNOTEL, and 200 snow course sites in the Interior Columbia River Basin.

A northwest cooperative agricultural weather network (AgriMet) is maintained by the Bureau of Reclamation in Boise, Idaho as part of the

Pacific Northwest Hydrometeorological Network for river and reservoir management.[1] Historical data since 1983 include daily soil moisture, soil and air temperatures, crop water use, and evapotranspiration.

A consistent network of radiosonde observations (RAOBs) began in the mid 1930s (U.S. Department of Commerce 1964) with electronic files available since 1946. RAOB sensors measure wind, temperature, dew point, and height at mandatory atmospheric pressure levels (surface, 850 mb, 700 mb, 500 mb, etc.) and other significant levels twice a day, at 0 Greenwich Mean Time (GMT) and 12 GMT. Only two stations within the Basin regularly report upper-air data: Spokane, Washington and Boise, Idaho.

[1] Monte McVay, Bureau of Reclamation, 1150 N. Curtis Road, Boise, ID 83706.

A summary of surface and upper-air climate data, which are in accessible electronic formats, is shown in Table 1. Other local and regional meteorological observations exist within the Columbia River Basin but are not as easily accessed as the above data. For example, the Hanford meteorological station, which includes a 125 meter (410 feet) tower, has been recording observations on a plateau in southwestern Washington since 1945 (Stone and others 1983). The Columbia River Operational Hydromet System (CHROMS) organizes data from the Bureau of Reclamation, Geological Survey, Natural Resources and Conservation Service, Forest Service, and British Columbia Hydro and Power Authority.[2] State highways began installing automated weather systems in the early 1990s. Also, regional avalanche centers and ski areas maintain automated weather stations (for example, Ferguson and others 1990). Other meteorological data, like the forest fire lookout observations, remain in paper form but are no less valuable.

Simulated Data

[2] Contact the individual agency for historical records.

Over 600 mountain weather and snowpack observations in the Interior Columbia River Basin would seem like enough to describe mountain climate. These many stations, however, are spaced too far apart (50 to 150 km) to represent small-scale climate caused by complex topography in the region. At least 10 km horizontal resolution is needed to resolve many of the mesoscale features that influence climate in the Pacific Northwest (Doran and Skyllingstad 1992). In addition, the limited amount and quality of data from each station make analysis rather cumbersome. It is no wonder then, that model-generated data play an increasing role in climate analyses of the mountainous west. Three sets of model-generated data were available for this study: 1) historical means of monthly and annual precipitation at 2.5 minute (about 5 km) and 5 minute (about 10 km) latitude/longitude spatial resolution from the PRISM model (Daly and others 1994); 2) daily temperature and precipitation for three characteristic years (1982, 1988, and 1989) at 2 km resolution from the MTCLIM-3D model (Thornton and Running 1996); 3) and monthly mean winds at 5 minute latitude/longitude resolution from the WINFLO model[3]. Mapped output from MTCLIM-3D is shown in Appendix Q-1, 2, and 3; from PRISM in Appendix Q-4, and from WINFLO in Appendix Q-5, 6, 7,and 8.

REGIONAL CLIMATE PATTERNS

Zone Summaries

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[3] Ferguson, S.A., M.R. Peterson, P.S. Hayes, and T. Akram. [In Preparation]. Surface wind patterns in the Pacific Northwest.

The Interior Columbia River Basin has been divided into three broad regions: 1) Eastern Cascades, 2) Northern Rockies, and 3) Central Columbia and Snake River Plateaus (Figure 1). The broad regions have been divided further into 13 distinct Ecological Reporting Units (ERUs). Because the ERUs do not exactly match National Weather Service climate division boundaries for which 30 year averages are calculated, new summaries were calculated using available COOP and HCN data within each ERU (Tables 2 to 14). In order to increase representation of variability caused by terrain in the climate, all data with records longer than 10 years were used. This means that the ERU summaries may not match NCDC calculated normals (30 year averages) for the same area. Whenever possible, ERU summary data were compared with climate division summary data to ensure that magnitudes and trends of ERU calculations were within reason. Because the spatial density of stations increase over time, ERU summary data may be slightly biased toward the most recent ten years.

Despite the increased spatial density of weather observing stations, they still do not represent the variability of climate in highly complex terrain. In order to quantify this concern, 30-second digital elevation model (DEM) data were analyzed for each ERU (Figure 2). Only ERUs with reasonably accessible terrain (for example, those dominated by rolling hills and plateaus as in 3, 4, 5, and 10) have average weather station elevations in the midrange of terrain elevations. Inaccessible mountain terrain in all other ERUs cause most weather stations to be sited at elevations well below the mean terrain elevation. This can make it difficult to describe climate influences on vegetation within the Basin because the elevational extent of many species (Franklin and Dyrness 1988; Volland 1976; Fowells 1965; Johnson and Simon 1987; Daubenmire and Daubenmire 1968; Shantz and Zon 1924) is well above the average elevation of weather observing stations.

The elevational distribution of weather observations would improve by including data from snow course, SNOTEL, and RAWS stations. Unfortunately, the different reporting formats, suspicious quality control, and difficulty in obtaining records covering a long period from these observing networks prevented their use in ERU summary calculations. Data from these stations were used, however, to help verify model results and improve qualitative descriptions.

Elevation-regression models like PRISM (Daly and others 1994) and MTCLIM-3D (Thornton and Running 1996) improve representation of high-elevation climate by interpreting available observations using Atopographically intelligent@ techniques. Both methods are relatively consistent, differing only in local detail.

Eastern Cascades

The Cascade Mountains block most marine air from entering the Basin. The eastern slopes of the mountains (ERUs 1, 2, and 3) lie in a rain shadow of oncoming Pacific storms. During winter, however, when westerly winds are strongest, enough moisture spills over the crest to cause this region to remain wetter than other parts of the Basin (Figure 3). This region also receives quantities of snowfall. Seasonal totals typically range from about 200 cm in the south (ERU 03) to over 300 cm in the north (ERU 01), with greater amounts at higher elevations (over 2000 cm of snowfall has been recorded at Crater Lake, Oregon). During summer, when westerly winds are weak, the rain shadow effect of the Cascades is most apparent and this region becomes the driest in the Basin (Figure 3).

Chinook winds can cause occasional warm, dry, and windy conditions that rapidly can melt snow or initiate blow-down. Strong winds also are common within mountain gaps as air flow is channeled from both east and west directions. The westerly gap winds, most common in summer, are strongest as they flow into the Eastern Cascade region. The principal mountain gap is the Columbia River gorge, just east of Portland, Oregon. Tornadoes and funnel clouds have been observed near the outflow of the gorge's westerly winds. In addition, the persistence of its winds allow the gorge to be one of the nation's principal wind surfing recreation areas. Strong southerly winds also are common, mainly during winter, on several east-west oriented ridges that protrude into the Basin from the Cascade crest.

Although this region often is under continental-type climate conditions (with cool, dry winters and hot, dry summers), marine air spilling over the mountain crests and through mountain gaps moderates both summer and winter temperatures. In addition, Arctic air often pools in the Basin and is pulled against the Cascades, causing a persistent temperature inversion to about 1200 meters, especially in winter. Sharp contrasts in temperature occur when Arctic air is displaced by marine air and rain-on-snow flooding is common at all elevations.

Northern Rockies

The Canadian Rockies and British Columbia Mountains block most Arctic air from entering the Basin. The deep Okanogan, Columbia, and Pend Oreille valleys, however, can funnel the dry Arctic air into the Basin where it often stagnates, especially during winter when it is cold and dense.

The Rocky Mountains on the eastern border of the Basin intercept continental air masses that rise over the imposing mountains, favoring thunderstorm development around the edge of the Basin, especially in western Montana and southern Idaho. A thermal trough that migrates northward during spring and summer also can cause thunderstorms, mainly in central Oregon around the Blue Mountains (ERU 06). The convection causes an increase in precipitation during spring, with 24 hour accumulations often greater than 25 mm. Drier lightning is more common during summer and fall. Most of the convection and lightning occur in the east and southeast units of this region (ERUs 9 and 12) nearest the Continental Divide.

Blow-down in this region is common. Strong down burst winds associated with convective cells can cause blow-down, most commonly during spring. Blow-down also occurs in the north-south elongated valleys that channel strong southerly storm winds, mainly during winter. The high and contrasting topography also favors accelerated storm winds near ridgetops during winter and persistent slope winds during summer.

The northern Rockies are the coldest part of the Basin with mean winter temperatures between minus 4.2°C and minus 10.2°C. Snowfall amounts vary from about 200 cm to over 300 cm, with greater amounts at higher elevations. Note that although winter precipitation in the northern Rockies is less than the eastern Cascades, snowfall amounts are comparable. There are two reasons for this: 1) cold winter temperatures cause relatively low density snow to fall in the northern Rockies so snowfall amounts appear greater even though snow water equivalents may be less than the eastern Cascades, and 2) colder spring and fall temperatures in the northern Rockies allow more snowfall during those seasons than in the warmer Cascades.

Despite cold winter temperatures in the northern Rockies, occasional marine intrusions can cause sudden warming and initiate rain-on-snow floods, mainly at lower elevations and those places exposed to flow from the Columbia gorge. In summer the marine intrusions moderate summer high temperatures and add moisture to the convective cycle, increasing the chance that lightning is accompanied by precipitation.

Central Columbia and Snake River Plateaus

This is the driest part of the Basin (Figure 3). The Columbia Plateau and Snake River Valley, however, are susceptible to marine intrusions that can relieve summer hot spells and winter cold spells. The accompanying moisture can put out summer wildfires or cause winter rain-on-snow floods. Although rain-on-snow floods are rare in this region, when they do occur they are more destructive and of much greater magnitude than spring floods. Typical seasonal snowfall totals range from 40 to 80 cm.

The upper plateaus (ERUs 4, 10, and 11) experience a moderate spring cycle of convective precipitation with lightning most common in ERU 11. The convection can be caused by the northward migration of a thermal low, especially in ERU 05, ERU 04, and in the western part of ERU 10. Also, hot unstable air from the Great Salt Lake region can increase thunderstorm and lightning development over ERU 11 and eastern ERU 10.

Strongest winds in this region occur during the summer from the west at the east outflow of the Columbia gorge (ERU 05) and during the winter from the south and west along ridgetops. Although climate in this region is marked by few extremes, long periods of stagnation occur during winter in the central Columbia Basin (ERU 05), in the Snake River Valley (ERU 10), and in high, isolated basins (ERUs 4, 10 and 11). The stagnation events cause this region to be the most susceptible to air pollution concerns.

TRENDS IN REGIONAL CLIMATE PATTERNS

Prior to about 1900, climatic trends in the Interior Columbia River Basin can only be determined using proxy data like that evident from tree rings, glacier fluctuations, ice cores, deep sea sediments, lake levels, and fossil pollens. Several major epochs in modern[4] climate have been manifested globally (Intergovernmental Panel on Climate Change 1990). A thermal maximum occurred between about 9000 and 5000 years ago (Holocene period) when temperatures were 1° to 2°C greater than today. Another warm period occurred between AD 1000 and 1250, known as the Medieval Climatic Optimum, with temperatures about 0.5°C warmer than today. A ALittle Ice Age@ between AD 1550 and 1850 caused low snow levels and advancing mountain glaciers.

Regional trends in ancient climate also are noteworthy. For example, fossil-pollen data show variable responses to alternating warm, dry and warm, wet climates in the Holocene period (NOAA 1992 and 1993; Whitlock and Bartlein 1993); with the spatial heterogeneity controlled primarily by topography (Davis and others 1986; Whitlock and Bartlein 1993). In

[4] Many consider modern climate to have begun after the last major ice age, about 10,000 years ago.

more recent times, local tree-ring chronologies (for example, Briffa and others 1992) suggest periodic warm periods in 1630s, between 1640 to 1660, 1790s, and the 1920s, with temperatures about 0.5℃ greater than the 1881 to 1982 average. Alternating cold periods also are apparent in the chronologies, with the most significant from 1870 to 1900 (about the time of the Little Ice Age) having temperatures 0.2 to 0.5℃ less than the 1881 to 1982 average.

Historical weather observations within the Columbia River Basin began in the late 1800s around the time that the Little Ice Age ended. Since then, mountain glaciers have generally retreated (Meier 1984; Ferguson 1992), a sign of general warming. A large number of weather stations were added during the 1930s drought to help monitor mountain snowpack and agricultural water conditions. Therefore, many trends seen in available data could show slight cooling from that drought period. Drought returned in the 1950s and 1980s but cooling comparable to the late 1800s has not occurred. There appear to be decadal trends in regional climate that could be related to the El Niño Southern Oscillation (ENSO) and Pacific North American (PNA) indices (Redmond and Koch 1991; Cayan and Peterson 1989; Yarnal and Diaz 1986; Ropelewski and Halpert 1986). These trends are most obvious in streamflow fluctuations, which aggregate precipitation and temperature signals over large areas of the catchment and drainage basin of the stream.

To analyze climatic trends in each ERU of the Interior Columbia River Basin, representative stations were selected that have weather records longer than 40 years and that lie near the midrange of station elevations for that unit. A simple ten year moving average of monthly mean values was used to evaluate trends and to help smooth out year-to-year variability. Historical trends in precipitation are apparent, while trends in temperature are not obvious.

A notable maximum in snowfall and winter precipitation occurred in all areas during the mid 1970s (Figure 4). This coincides with a step-wise change in climate observed in other environmental variables during 1976 (Ebbesmeyer and others 1991). Several new ski areas developed and existing areas expanded during that period. Previous maximums in winter precipitation occurred in the mid 1950s (when most alpine ski areas in the western U.S. began), early 1940s, and mid 1930s, but these were not as pervasive throughout the Basin as the mid 1970s maximum.

Winter precipitation since the mid 1970s decreased by 30 to 80 percent in all areas to a level comparable to historical means, except in ERUs 3, 8, 9, and 10 whose current winter precipitation trend is slightly lower than any in previous history (Figure 5).

In contrast to winter trends, all areas showed an increasing trend (30 to 80 percent) in summer precipitation during the last 30 years (Figure 6). Comparable highs in summer precipitation occurred around 1910. Since the mid 1980s, about half the Basin (ERUs 3, 6, 10, 12, and 13) has shown a 20 to 70 percent decrease in summer precipitation, but remain 40 to 80 percent above the 1960s levels. Periods of drought throughout the Basin since 1988 appear to be related to the overall decrease in winter precipitation and beginning signs of decreasing summer precipitation (Chapter 17c).

Trends in annual precipitation over the last 100 years show significant variability around the Basin (Karl and others 1996). The central Basin (Washington, northeastern Oregon, and eastern Idaho) showed 5 to 10 percent increases. Elsewhere, annual precipitation decreased 5 to 10 percent, with ERU 03 showing a 20 percent decrease. It is difficult to compare the 100 year trends with those calculated from stations within each ERU that have records of only 50 to 70 years. Also, Mock (in press) illustrated the highly seasonal nature of precipitation in the western U.S. Therefore, changes in spring and autumn months that influence annual trends are not shown in the January and July records analyzed for each ERU.

Although there has been almost no change in measured winter temperatures in the Eastern Cascades, there has been a notable increase in the percentage of water in the winter precipitation (Figure 7). This could indicate somewhat higher snow levels with less snow accumulating at lower elevations. In the same region a 1 to 2°C decrease in the diurnal range of summer temperatures may suggest slightly increasing summer cloudiness (Figure 8). Other regions have less obvious changes but the overall trends appear to be toward slightly cooler summers and slightly warmer winters.

Mean temperature has increased 1 to 2°C throughout the Basin over the last 100 years (Karl and others 1990). Note again that many of the stations used to analyze ERU trends have records begin during the 1930s drought era. Therefore, their trends are smaller than the 100 year trends, and even slightly negative.

The persistent drought throughout the Basin in the 1930s may relate to warm winter temperatures and associated high snow levels. Many areas also experienced a decrease in summer precipitation during the 1930s, adding to severe drought conditions. Periods of significant drought in a few areas of the central Basin and plateau regions during the 1950s appear driven more by warmer than normal temperatures rather than changes in precipitation.

SUMMARY

Climate of the Interior Columbia River Basin can be characterized by 3 distinct air mass types that interact with each other in a region of complex topography. Most precipitation accumulates during winter (75 to 125 cm in the Eastern Cascades, 25 to 95 cm in the Northern Rockies, and 20 to 40 cm in the Central Columbia and Snake River Plateaus). The mountain snowpack acts like a natural reservoir and supplies the Basin with most of its useable water. Only the east and south ERUs of the Basin have summer maximum precipitation, which is associated with significant thunderstorm activity. Summer precipitation throughout the Basin ranges from about 20 to 50 cm. Trends in the last 50 to 100 years indicate a general decrease in winter precipitation and increase in summer precipitation.

Temperatures are generally mild in the Basin because of periodic influxes of moderating Pacific moisture. Winter mean monthly temperatures range from minus 10 to minus 3°C while summer temperatures range from 10 to 15°C. Trends in the last 50 to 100 years indicate a slight increase in winter temperatures and slight decrease in summer temperatures.

More effective analysis of the Columbia River Basin climate would include high elevation data from snow course, SNOTEL, and RAWS sites. To do so requires a significant amount of data processing to check and adjust for quality. In addition, outfitting RAWS sites with winter durable sensors would improve their reliability as climate indicators. Developing a unified database from all available data sources also would be a great asset. Exploring new methods of analyzing spatial and temporal variability would significantly improve our understanding of climate patterns in regions of complex topography like the Columbia River Basin.

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FIGURES CAPTIONS

Figure 1. The Interior Columbia River Basin showing broad zones, boundaries of Ecological Reporting Units, and climate division boundaries.

Figure 2. Terrain distribution and weather station elevations in each Ecological Reporting Unit (ERU) within the Interior Columbia River Basin. Terrain elevations were divided into 500 meter increments, beginning with 0 to 500 meters and ending with 3500 to 4000 meters. The percentage of terrain in each elevation increment is shown by the height of bar. The terrain increment, which includes the average station elevation for COOP and HCN weather observation stations, is marked with an arrow.

Figure 3. Seasonal values of temperature and precipitation for each ERU. The solid line shows July average temperatures and the dashed line shows January average temperatures. The hatched bar shows January precipitation and the hollow bar shows July precipitation.

Figure 4. January precipitation and snowfall data from Moscow, Idaho that represent winter climate trends in ERU 05. Shaded bars show monthly precipitation totals in millimeters. Hollow bars show monthly snowfall totals in centimeters. The thick line shows the moving ten-year averages of monthly precipitation and the thin line shows the moving ten-year averages of snowfall.

Figure 5. January precipitation and snowfall data from Lake View, Oregon that represent winter climate trends in ERU 03. Same as in Figure 4.

Figure 6. July precipitation data from McCall, Idaho that represent summer climate trends in ERU 13. Same as in Figure 4 except no snowfall.

Figure 7. January precipitation and snowfall data from Bend, Oregon that represent winter climate trends in ERU 02. Same as in Figure 4.

Figure 8. July temperature data from Bend, Oregon that represent summer climate trends in ERU 02. Shown are monthly average temperatures (•), monthly average range between daily maximum and minimum temperatures (•), with moving ten-year averages of each (thick and thin lines, respectively).

TABLES CAPTIONS

Table 1. Summary of climate data sets. See text for an explanation of data set names. T = dry-bulb temperature, Td = dew point temperature, Tf = fuel temperature, RH = relative humidity, ppt = precipitation, SWE = snow water equivalent, W = wind, P = atmospheric pressure, Q = solar radiation, H = snow depth.

Table 2a. Climate summary for ERU 01 (26 weather stations). The range of station elevations (meters) are shown. Data from available HCN and COOP weather observation stations having records of 10 years or more were used to calculate monthly mean values (°C) of maximum daily temperature [T(max)], minimum daily temperature, daily range in temperature [del T], average daily temperature [T(avg)], and monthly total values of daily precipitation [PPT] in millimeters and snowfall in centimeters. In addition, the ratio of snowfall to precipitation [percent water] is given to help show water content of snow.

Table 2b. ERU 01 seasonal trends in daily temperature range (delT), average temperature (Tavg) and precipitation (PPT), as approximated by data from Cle Elum, Washington (588 meters elevation and 64 years of record). Significant trends in winter precipitation after about 1975 and summer precipitation after about 1960 and 1985 also are shown.

Table 3a. Climate summary for ERU 02 (18 weather stations). Same as in Table 2a except for ERU 02.

Table 3b. ERU 02 seasonal trends. Same as in Table 2b except from Bend, Oregon (1116 meters elevation and 67 years of record).

Table 4a. Climate summary for ERU 03 (12 weather stations). Same as in Table 2a except for ERU 03.

Table 4b. ERU 03 seasonal trends. Same as in Table 2b except from Lake View, Oregon (1457 meters elevation and 67 years of record).

Table 5a. Climate summary for ERU 04 (22 weather stations). Same as in Table 2a except for ERU 04.

Table 5b. ERU 04 seasonal trends. Same as in Table 2b except from Squaw, Oregon (1420 meters elevation and 58 years of record).

Table 6a. Climate summary for ERU 05 (87 weather stations). Same as in Table 2a except for ERU 05.

Table 6b. ERU 05 seasonal trends. Same as in Table 2b except from Moscow, Idaho (810 meters elevation and 95 years of record).

Table 7a. Climate summary for ERU 06 (34 weather stations). Same as in Table 2a except for ERU 02.

Table 7b. ERU 06 seasonal trends. Same as in Table 2b except from Union, Oregon (844 meters elevation and 67 years of record).

Table 8a. Climate summary for ERU 07 (52 weather stations). Same as in Table 2a except for ERU 07.

Table 8b. ERU 07 seasonal trends. Same as in Table 2b except from Fortine, Montana (914 meters elevation and 80 years of record).

Table 9a. Climate summary for ERU 08 (29 weather stations). Same as in Table 2a except for ERU 08.

Table 9b. ERU 08 seasonal trends. Same as in Table 2b except from Haugan, Montana (944 meters elevation and 80 years of record).

Table 10a. Climate summary for ERU 09 (21 weather stations). Same as in Table 2a except for ERU 09.

Table 10b. ERU 09 seasonal trends. Same as in Table 2b except from Butte, Montana (1689 meters elevation, 95 years of record).

Table 11a. Climate summary for ERU 10 (62 weather stations). Same as in Table 2a except for ERU 10.

Table 11b. ERU 10 seasonal trends. Same as in Table 2b except from Caldwell, Idaho (722 meters elevation and 90 years of record).

Table 12a. Climate summary for ERU 11 (22 weather stations). Same as in Table 2a except for ERU 11.

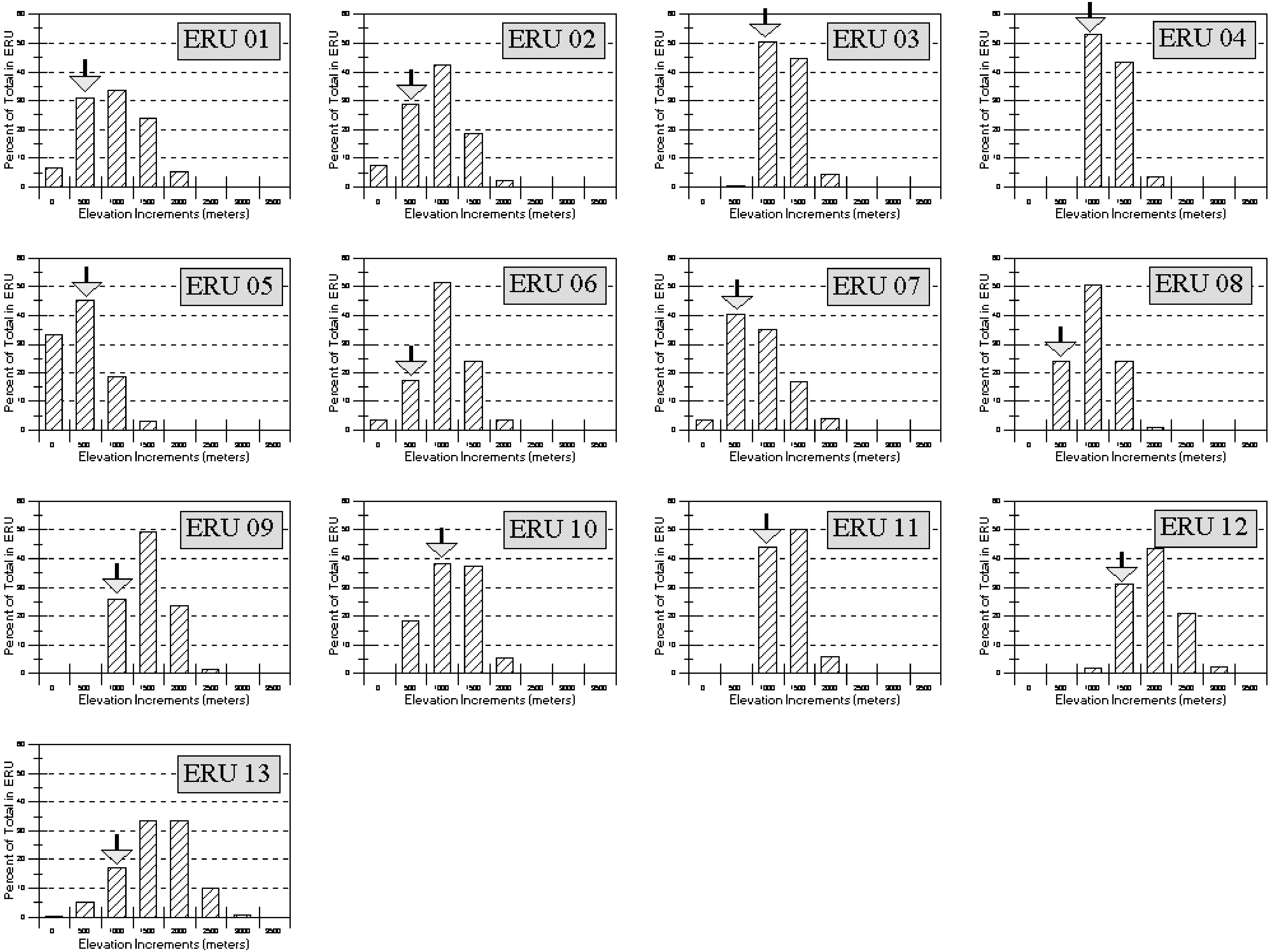
Table 12b. ERU 11 seasonal trends. Same as in Table 2b except from Aberdeen, Idaho (1344 meters elevation and 81 years of record).

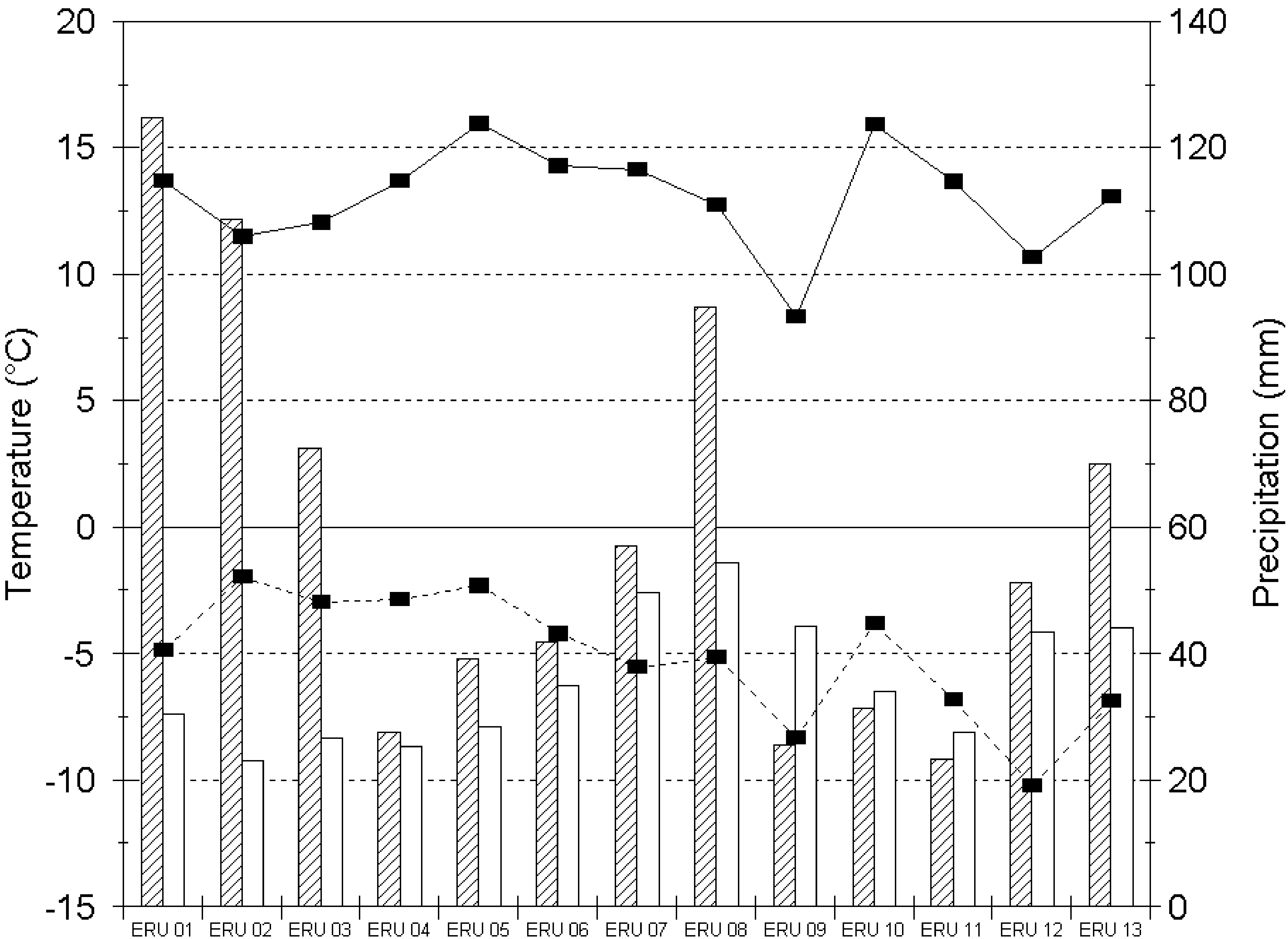
Table 13a. Climate summary for ERU 12 (12 weather stations). Same as in Table 2a except for ERU 12.

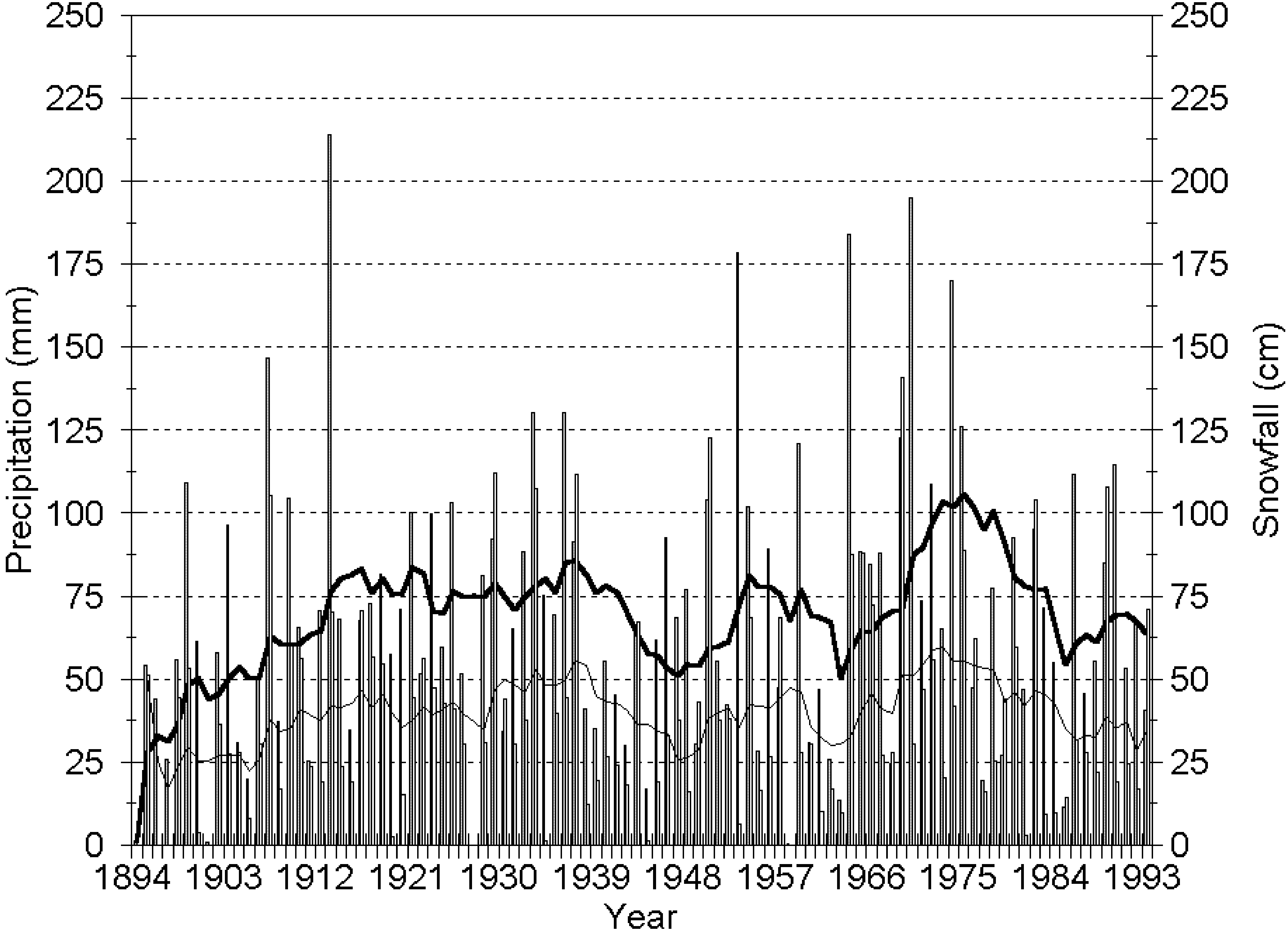
Table 13b. ERU 12 seasonal trends. Same as in Table 2b except from Ashton, Idaho (1603 meters elevation and 47 years of record).

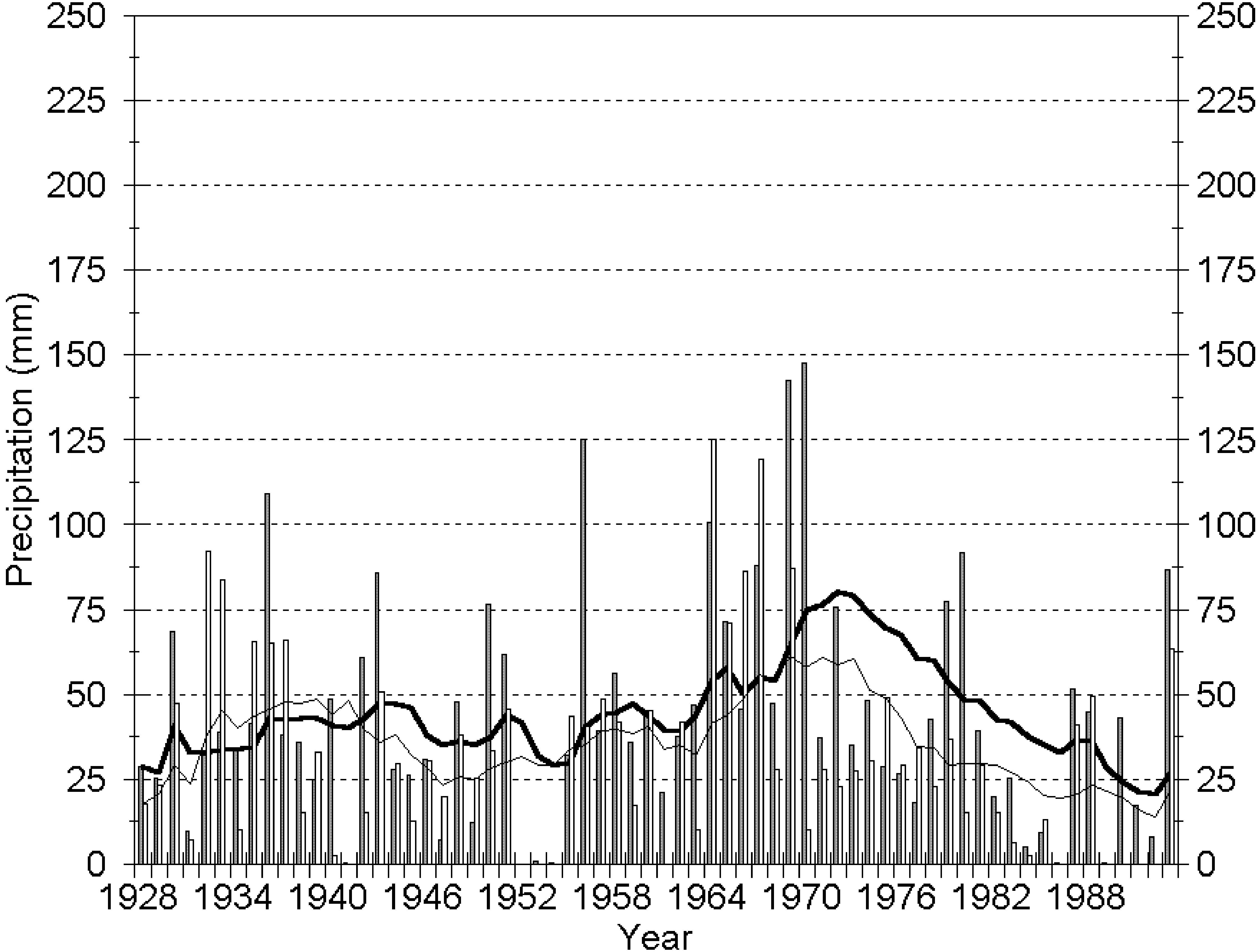
Table 14a. Climate summary for ERU 13 (51 weather stations). Same as in Table 2a except for ERU 13.

Table 14b. ERU 13 seasonal trends. Same as in Table 2b except from McCall, Idaho (1533 meters elevation and 65 years of record).

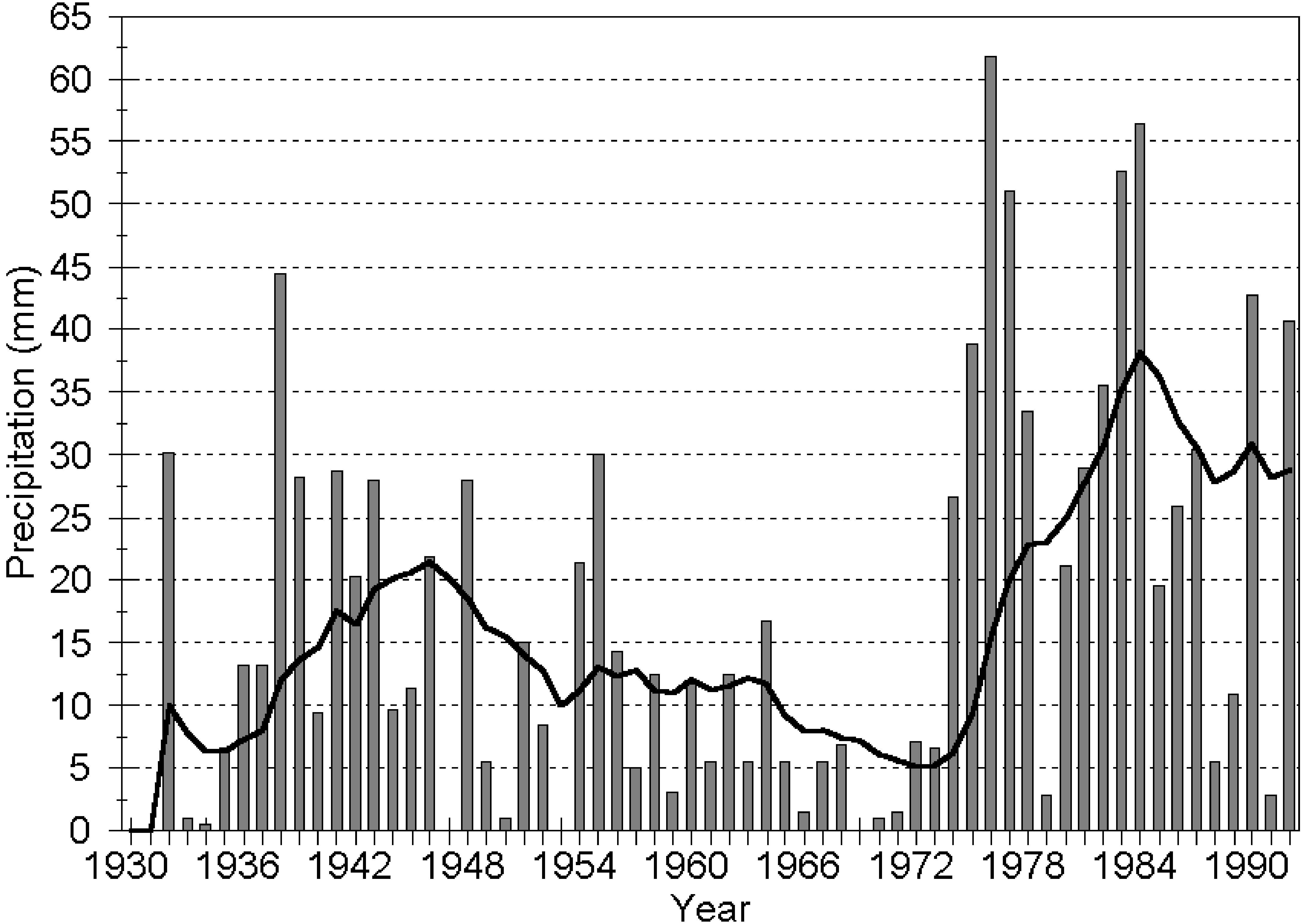

ERU 01
Percent of Total in ERU
Elevation Increments (meters)
ERU 02
Percent of Total in ERU
Elevation Increments (meters)
ERU 03
Percent of Total in ERU
Elevation Increments (meters)
ERU 04
Percent of Total in ERU
Elevation Increments (meters)
ERU 05
Percent of Total in ERU
Elevation Increments (meters)
ERU 06
Percent of Total in ERU
Elevation Increments (meters)
ERU 07
Percent of Total in ERU
Elevation Increments (meters)
ERU 08
Percent of Total in ERU
Elevation Increments (meters)
ERU 09
Percent of Total in ERU
Elevation Increments (meters)
ERU 10
Percent of Total in ERU
Elevation Increments (meters)
ERU 11
Percent of Total in ERU
Elevation Increments (meters)
ERU 12
Percent of Total in ERU
Elevation Increments (meters)
ERU 13
Percent of Total in ERU
Elevation Increments (meters)

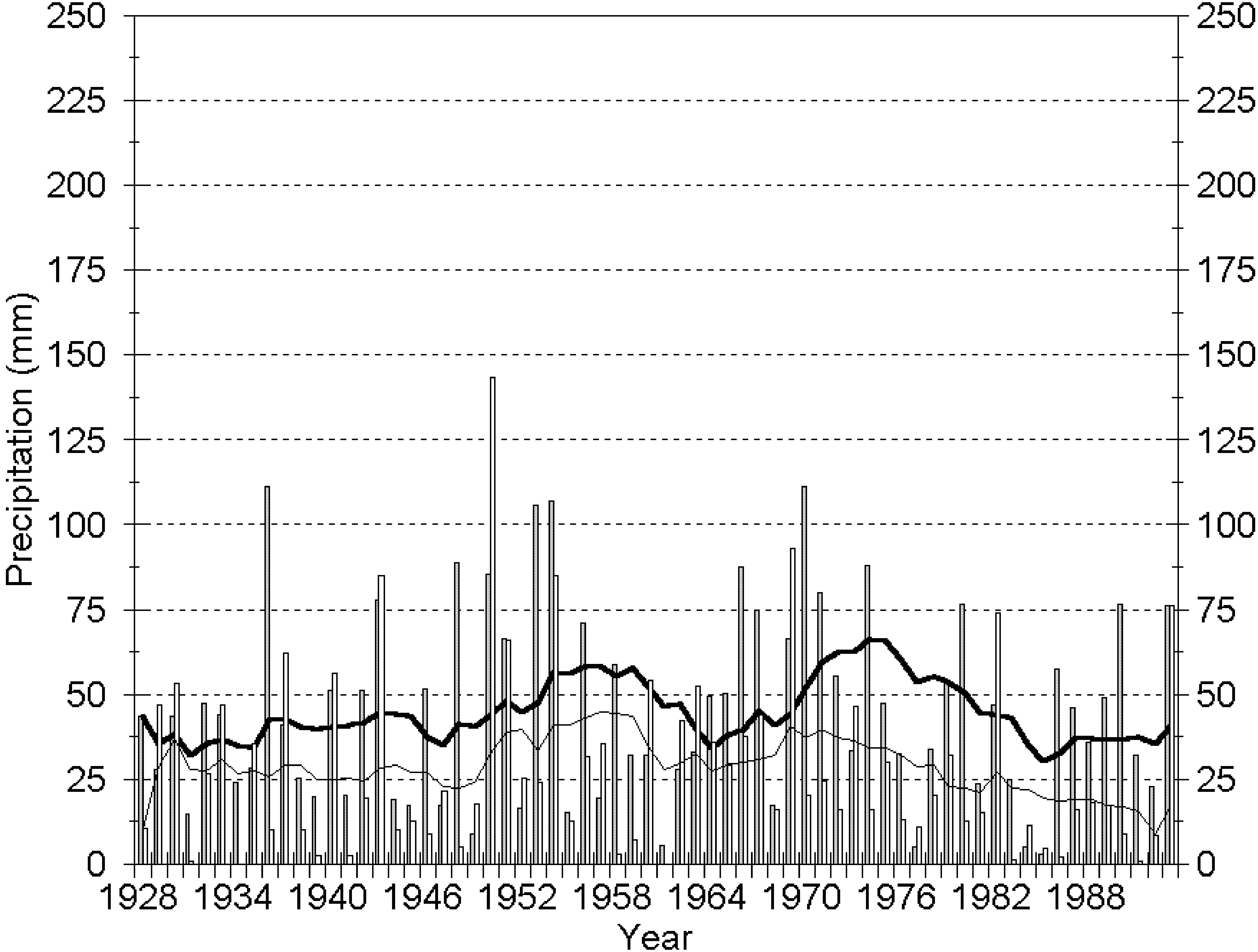

20
15
10
5
0
-5
-10
-15
Temperature (°C)
140
120
100
80
60
40
20
0
Precipitation (mm)
ERU 01
ERU 02
ERU 03
ERU 04
ERU 05
ERU 06
ERU 07
ERU 08
ERU 09
ERU 10
ERU 11
ERU 12
ERU 13


Precipitation (mm)
Snowfall (cm)
Year
250
225
200
175
150
125
100
75
50
25
0
1894
1903
1912
1921
1930
1939
1948
1957
1966
1975
1984
1993

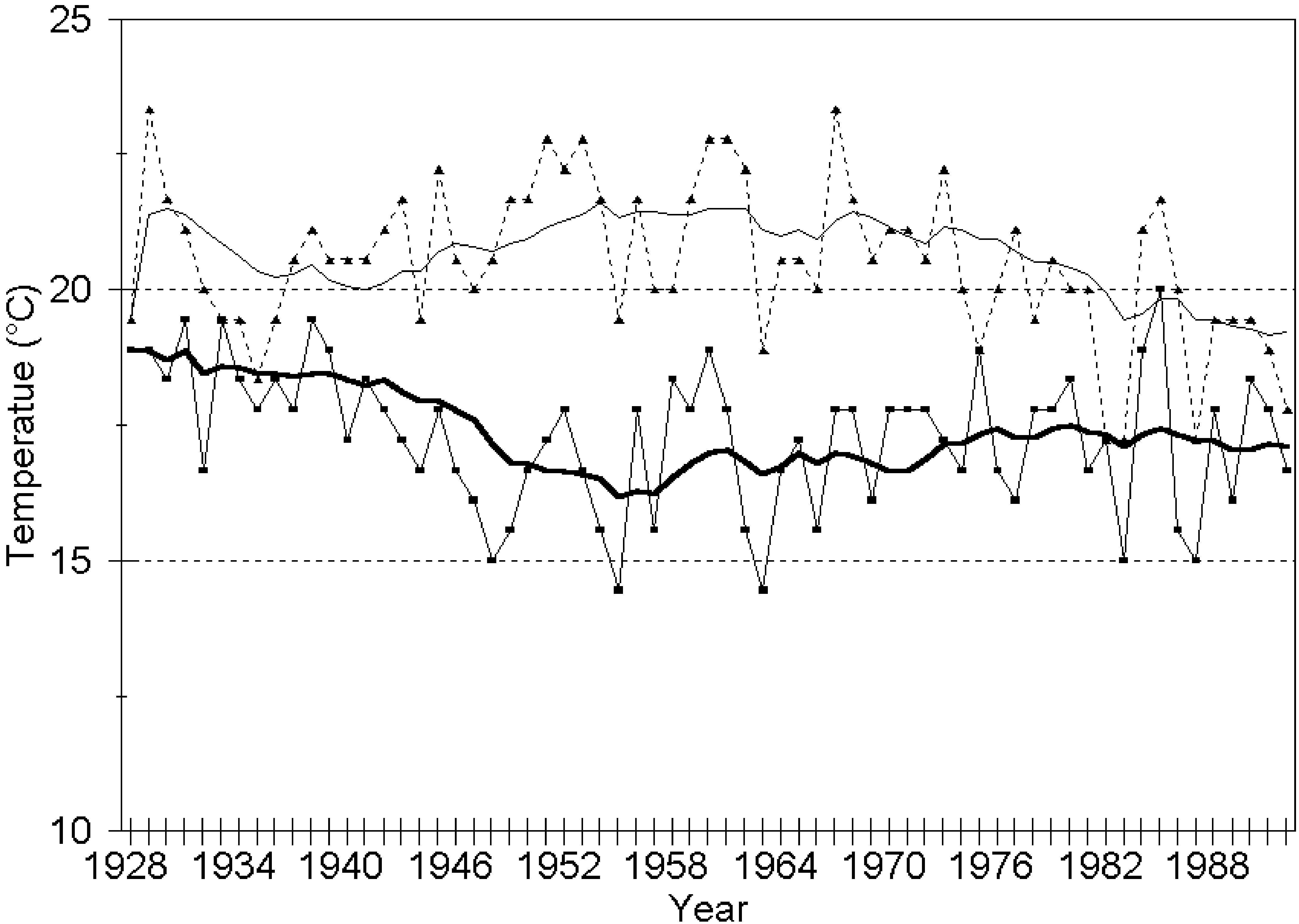


Precipitation (mm)
Snowfall (cm)
Year
0
25
50
75
100
125
150
175
200
225
250
1928
1934
1940
1946
1952
1958
1964
1970
1976
1982
1988


65
60
55
50
45
40
35
30
25
20
15
10
5
0
Precipitation (mm)
1930
1936
1942
1948
1954
1960
1966
1972
1978
1984
1990
Year



250
225
200
175
150
125
100
75
50
25
0
Precipitation (mm)
Snowfall (cm)
1928
1934
1940
1946
1952
1958
1964
1970
1976
1982
1988
Year


25
20
15
10
Temperatue (°C)
1928
1934
1940
1946
1952
1958
1964
1970
1976
1982
1988
Year