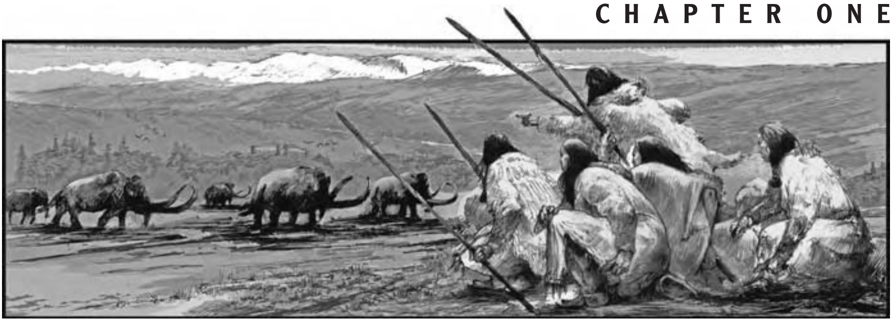
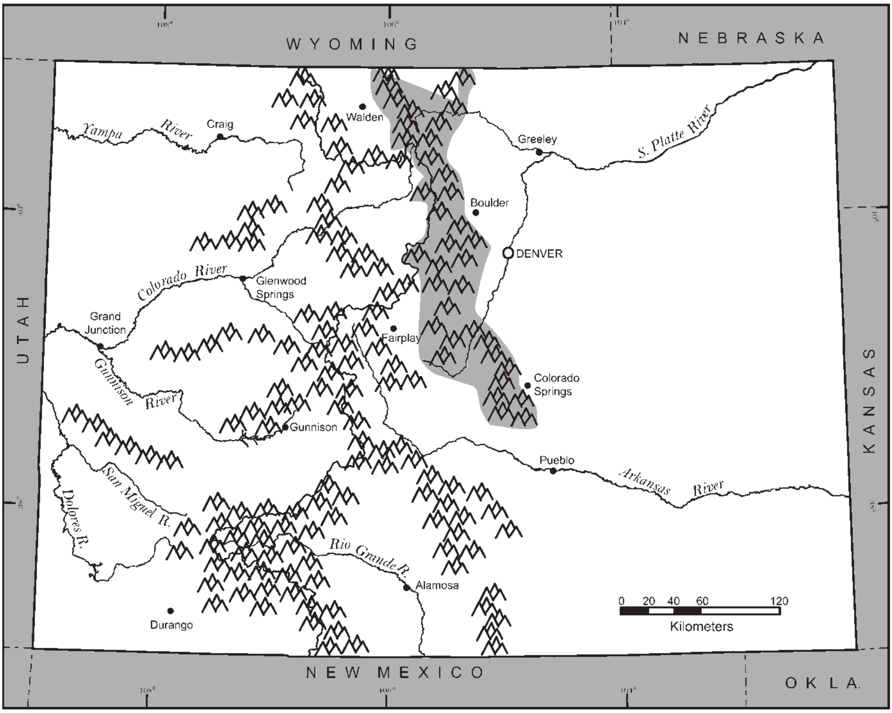
CHAPTER ONE



James P. Doerner

Late Quaternary Prehistoric Environments of the Colorado Front Range

This chapter examines the prehistoric environments of the Colorado Front Range during the past 25,000 years, the interval encompassing the most recent glacial-interglacial cycle (Porter 1983). This interval is generally referred to as the late Quaternary Period and is a critical time in both human and earth history. It was during this time that humans first arrived in North America, large-scale extinctions of Pleistocene mammals occurred, and boreal vegetation consisting mainly of tundra plants and spruce woodlands covered large areas of North America south of the continental ice sheets (Holloway and Bryant 1985). The interval also includes the termination of the Pleistocene and establishment of a new post-glacial macroclimate during the Holocene. These new environmental conditions meant that old plant associations and distributions had to adjust to remain in equilibrium with the changing environment (Baker 1983). As glacial



1.1. Map showing major physical features of Colorado. Shaded area shows the generalized location of the Colorado Front Range in north-central Colorado.

conditions ameliorated and alpine glaciers melted away, vegetation zones apparently responded by advancing upslope from the plains and the lower montane zones to reclaim the higher elevations (Elias 1988).

Environmental changes during the late Quaternary have been substantial. Climate is probably the most important environmental factor, controlling growth and decay of ice sheets, distributions of flora and fauna (including human populations), development of soils, and erosion and sedimentation processes on the landscape (Wright 1983). The reconstruction of past environments is essential for interpreting past human occupations and cultural landscape modeling.

The remainder of this chapter is divided into four sections: (1) a description of the regional setting, (2) a review of climate reconstructions based on glacial chronologies, (3) an overview of paleoecological studies derived from pollen and fossil beetle investigations, and (4) a summary of late Quaternary climate change in the Colorado Front Range.

REGIONAL SETTING

The Colorado Front Range (Figure 1.1) is made up of the easternmost mountain ranges that comprise the Southern Rocky Mountain physiographic province

1.2. Map showing major air mass trajectories into the Colorado Front Range.

(Hunt 1967). The Front Range is about 300 km long, extending from the Arkansas River in the south to the Colorado-Wyoming border in the north (Short 1985). The eastern margin of the Front Range is clearly defined by steeply dipping sedimentary layers that mark the contact with the Great Plains. Along the western margin, the Front Range is bordered by other mountain ranges and intermontane basins (Marr 1967). The topography throughout most of the area is rugged and steep, with broad glaciated valleys prominent above 2,800 m (Veblen and Lorenz 1986).

The modern climate is classified as highland continental, with short cool summers, long cold winters, and relatively dry conditions throughout the year (Griffiths and Rubright 1983). Dramatic changes in average temperature and precipitation occur with increased elevation (Barry 1972). Marr (1967) notes that the area is subject to extreme changes in atmospheric conditions, which can occur from hour to hour, day to day, season to season, and year to year. Climate is also strongly influenced by its position relative to prevailing air mass trajectories (Figure 1.2).

During winter months, westerly circulation brings cool, moist Pacific air masses to the mountains. This can produce heavy snowfalls for the mountains west of the Continental Divide, while the east slope of the divide experiences relatively little precipitation (Veblen and Lorenz 1991). The western slope has a wet season during winter months when the westerlies are the strongest, while the eastern slope is relatively dry during winter (Rink and Kiladis 1986). The Front Range receives most of its annual precipitation from storm systems originating in the Gulf of Mexico (Marr 1967). Maximum precipitation on the east slope usually occurs during the spring (Rink and Kaladis 1986). A secondary peak

Table 1.1. Climatic Parameters and Ecosystem Vegetation for the East Slope of the Colorado Front Range (Barry 1972; Marr 1967; Veblen and Lorenz 1991).

Elevational Range (m)	*Mean Annual Precipitation (mm)*	*Mean Annual Temperature (°C)*	*Vegetation Zone*
1,800 to 2,350	590	8.3	Lower montane
2,350 to 2,800	590	5.6	Upper montane
2,800 to 3,350	770	1.6	Subalpine
> 3,550	1,025	–3.3	Alpine

occurs in July and August, reflecting the influence of summer convective storms (Barry 1972). Elevation can also influence the timing of the precipitation peak for east slope locations in the Front Range (Barry 1973). Elevations at 3,000 m have a spring precipitation maximum, whereas areas above 3,750 m have a winter precipitation peak (Barry 1973).

Elevations in the Front Range vary from about 1,700 m at the base of the mountains to over 4,300 m on the highest peaks. This elevational gradient creates distinct climatic and vegetation zones, which roughly correspond to altitude (Table 1.1). In general, with an increase in elevation, there is a decrease in temperature and an increase in precipitation (Barry 1973, 1981). This provides more available moisture but shorter growing seasons in the mountains. As a result, the mountains have extensive forests compared with the grasslands on the surrounding plains and intermontane basins (Rink and Kiladis 1986).

In Colorado's mountains, grasses and herbs dominate vegetation at the highest and lowest elevations (Mutel and Emerick 1992). Forests in the Front Range are restricted to elevations between 1,800 and 3,350 m. This results in an upper and a lower tree line. At the highest elevations, trees are limited by cold, drought, and wind. For elevations below 1,800 m, trees are limited by drought. Needle-leafed conifers dominate the Front Range forests. Except for areas where disturbance has allowed aspen to invade, broad-leafed deciduous trees are limited to riparian habitats.

Forests in the Rocky Mountains exhibit altitudinal zonation, which can be seen in the vegetation of the Front Range. Marr (1967) divided the vegetation of the eastern slope of the Front Range into four major zones: (1) the lower montane zone, (2) the upper montane zone, (3) the subalpine zone, and (4) the alpine zone. Each of these zones is characterized according to a given range of temperature, humidity, type and amount of precipitation, growing season length, wind, and soil conditions. Boundaries between zones are not sharply defined; instead, there are ecotonal transitions between zones in which plant species from adjacent zones are intermixed.

The Lower Montane Zone

The vegetation in the lower montane zone is an open forest of broad-crowned evergreen trees frequently interrupted by grasslands (Marr 1967). The dominant tree species are ponderosa pine and Douglas fir, with Rocky Mountain juniper common on dry sites. Mountain mahogany, blue gramma and other grasses, shrubs, and herbs form the understory. Narrow-leaf cottonwood, Colorado blue spruce, river birch, and willow are common along stream courses. North-facing slopes support dense forests of Douglas fir and ponderosa pine. Douglas fir is more common on steeper slopes, while higher percentages of ponderosa pine may indicate the presence of coarse soils. On south-facing slopes, the forests are less dense and ponderosa pine is favored over Douglas fir.

The Upper Montane Zone

The vegetation in this zone is similar to that of the lower montane, but the overall density of the forest is greater. Douglas fir and ponderosa pine are still the dominant tree species. Limber pine is present where strong winds persist, soils are coarse, or both (Marr 1967). Aspen and lodgepole pine form dense secondary successional forests after fires, logging, or other disturbances (Windell, Willard, and Foster 1986). Aspen stands typically have a well-developed understory of forbs and grasses, in contrast to ponderosa pine stands, which have a sparse understory (Peet 1988). Aspen is generally found on wetter sites and lodgepole pine on drier sites (Griffths and Rubright 1983). Narrow-leaf cottonwood and Colorado blue spruce are the dominant tree species on the valley floors, with willow, alder, and birch forming complex stands along stream banks (Marr 1967).

The Subalpine Zone

The subalpine zone forms the highest, most continuous and pristine forests in Colorado (Mutel and Emerick 1992). Engelmann spruce and subalpine fir trees occur as dense unbroken forests (Marr 1967). These spruce-fir forests are replaced by stands of bristlecone pine on open, dry, south-facing slopes (Peet 1978) and by limber pine on windy, rocky, and exposed sites (Arno and Hammerly 1984). Dwarf juniper occur throughout this zone (Short 1985). Disturbance areas are indicated by stands of quaking aspen and lodgepole pine.

Spruce and fir have somewhat different ecological characteristics, with spruce more tolerant of extreme conditions than fir (Peet 1988). Spruce is the dominant species on very wet or boggy sites, as well as on the driest sites (Peet 1981). Where conditions are the driest, fir is absent from the subalpine zone (Peet 1988). Spruce is more successful at establishing itself on mineral soils following fires (Alexander 1974; Peet 1981), whereas fir is more successful at establishing itself in the shade and on organic substrates (Knapp and Smith 1982). In areas where stands are mixed, the forest canopy is usually dominated by spruce, but most of the seedlings and saplings in the understory are fir (Peet 1988).

The Krummholz ecotone is the transition between the subalpine and alpine zones. The trees in this region are stunted and dwarfed by the harsh environmental conditions. The lower boundary of the Krummholz ecotone is called timberline, the upper elevational limit of upright, erect trees. The upper boundary of this ecotone is termed tree line, the upper limit of any tree establishment. Trees here rarely grow taller than a meter or two and are often flagged, that is, they lack branches on the windward side because of desiccation by the wind.

The Alpine Zone

The alpine zone consists of low-growth shrubs, cushion plants, small forbs, sedges, and grasses (Mutel and Emerick 1992). Trees are completely absent in this zone. The majority of the plants are perennial, and there is less species diversity than in the subalpine. Most of the shrubs are in the willow family (Thilenius 1975). The important herb families include saxifrage, pink, rose, and buckwheat (Short 1985).

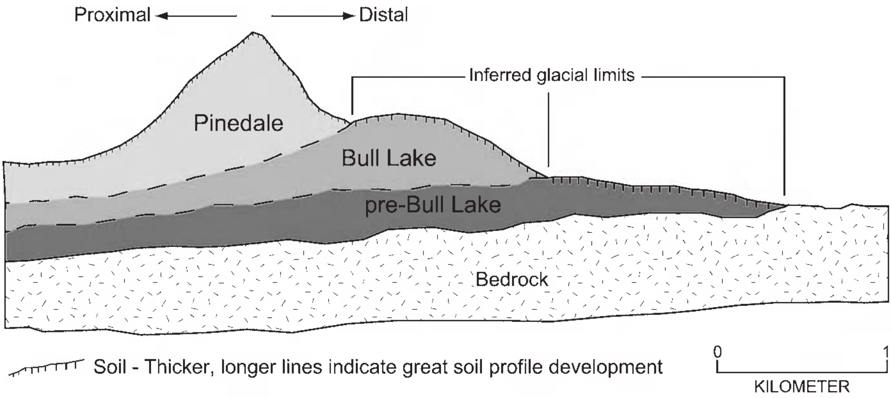
GLACIAL CHRONOLOGIES

The Late Pleistocene

In the Rocky Mountains, at least three distinct periods of Pleistocene glaciation (pre–Bull Lake, Bull Lake, and Pinedale) are recognized (Madole 1976; Table 1.2). Each glaciation can be distinguished based upon moraine morphology, soil development, and surface weathering characteristics (Madole, VanSistine, and Michael 1998; Figure 1.3). The most recent episode of Pleistocene glaciation, the Pinedale, is named after terminal moraines first studied near the town of Pinedale, Wyoming (Blackwelder 1915). The term Pinedale is now widely accepted for both the time and the deposits of the last extensive glaciation in the Rocky Mountains (Benson et al. 2005).

In Colorado, both early and late glacial advances occurred during the Pinedale; however, the chronology of the earlier advance(s) is not well defined by numerical age (radiocarbon dates) determinations. All ages discussed in this chapter are given in radiocarbon years (RCYBP) unless otherwise noted. Madole and Shroba (1979) showed that early Pinedale advances were more extensive than late Pinedale advances. Porter and colleagues (1983) noted that, during the late Pinedale, glaciers were at or near their maximum extent ca. 20,000 RCYBP. Madole (1986) has suggested that the maximum extent of the late Pinedale advance occurred between ca. 23,500 and 19,000 RCYBP. Benson and colleagues (2004) also concluded that glaciers in the Front Range reached their maximum extent by ca. 22,000 RCYBP.

Madole and colleagues (1998) mapped the extent of Pinedale glaciation in the Upper Platte River Basin and found that in the Front Range the maximum length of most glaciers was between 12 and 20 km, with maximum thicknesses between 180 and 350 m. Most large glaciers in the Front Range terminated in deep valleys at elevations between 2,500 and 2,700 m. None of the Front Range glaciers reached as far east as the mountain front, terminating between 17 and 40 km west of eastern foothills and piedmont.

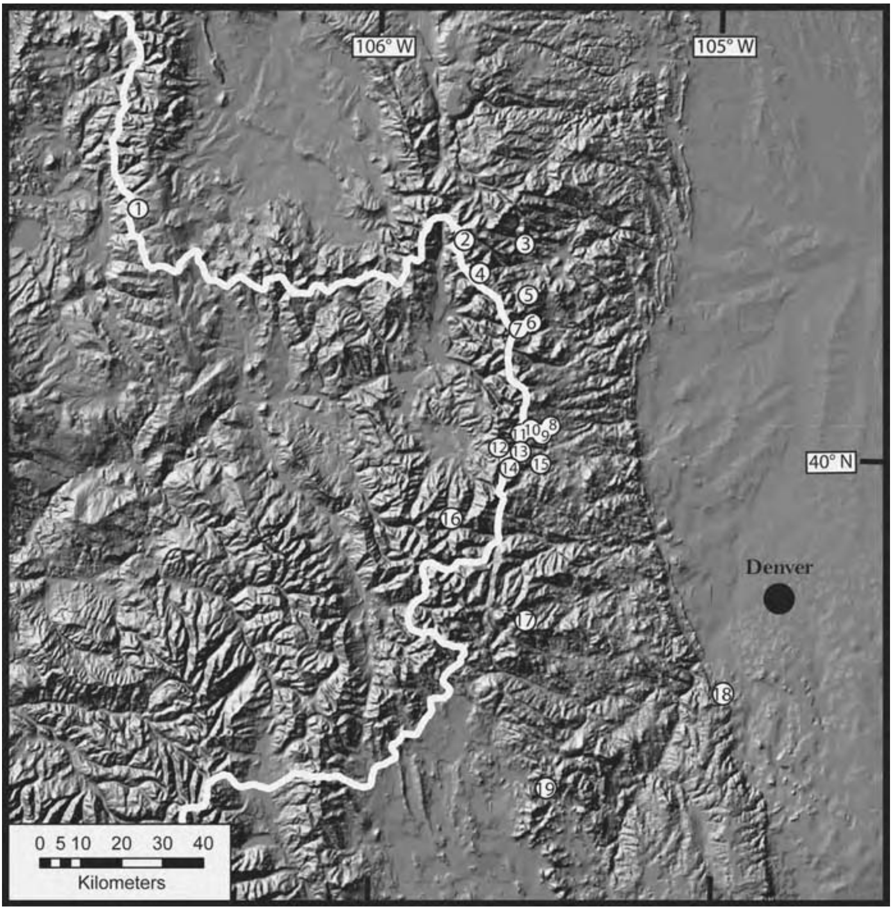


1.3. Schematic diagram showing how till of Pinedale, Bull Lake, and pre–Bull Lake glaciations differ in position, topographic expression, and degree of weathering and soil development (Madole, VanSistine, and Michael 1998).

Table 1.2. Quaternary Time Chart and Provisional Ages of Glaciations (Madole, VanSistine, and Michael 1998).

Formal Time Division		*Informal Time Division*		*Glaciations*	*Age*
Quaternary Period	Holocene Epoch				10,000
	Pleistocene Epoch				
		late Pleistocene		Pinedale	~30,000
		middle Pleistocene	late	Bull Lake	~130,000
					~300,000
			middle	pre–Bull Lake	
					~620,000
			early		
					~788,000
		early Pleistocene			
					~1,800,000

Pierce (2004) provides a comprehensive discussion of Pleistocene glaciation studies from the Rocky Mountains. Additional reviews of the glacial history of the Colorado Front Range are provided by Madole (1972, 1976, 1986), Meierding and Birkeland (1980), and Benson and colleagues (2004). Studies from the Front Range (Table 1.3; Figure 1.4) show spatial variability in the timing of Pinedale deglaciation. Madole (1980) used geologic and radiocarbon analyses of samples from sites at La Poudre Pass and near Buffalo Pass in the Park Range to suggest that the termination of Pinedale glaciation occurred by ca. 11,000 RCYBP. In a



1.4. Shaded relief map of Colorado Front Range region showing locations of glacial and paleoecological sites discussed in this chapter. Circled numbers refer to sites listed in Table 1.3. Solid white line shows the Continental Divide.

study from Lake Devlin, Madole (1986) suggested deglaciation occurred sometime between ca. 15,000 and 12,000 RCYBP.

Nelson and colleagues (1979), working at the Mary Jane site (Frazier River Valley) on the west slope of the Front Range, concluded that Pinedale deglaciation began just prior to ca. 13,000 RCYBP. In the Mount Evans region, sediments from a high-altitude lake (Echo Lake) suggest deglaciation started as early as ca. 15,000 RCYBP and was completed by ca. 13,000 RCYBP (Doerner, Sullivan, and Briles 1998). Benson and colleagues (2004) concluded that Pinedale glaciers in the Front Range had disappeared from all but the highest elevations by 14,000 RCYBP.

There is also evidence for late Pleistocene glacial advances following Pinedale deglaciation and climatic warming. Benedict (1973, 1981) found evidence of ice

Table 1.3. Late Quaternary Glacial and Paleoecological Sites and References from the Colorado Front Range. Site numbers are keyed to Figure 1.4.

Site Number	*Site Name*	*Elevation (m)*	*Reference(s)*
1	Buffalo Pass	3,146	Madole 1980
2	La Poudre Pass	3,103	Madole 1980
			Elias 1983, 1985
			Short 1985
3	Lawn Lake Fen	3,357	Doerner, Brunswig, and Sanborn 2002
4	Mount Ida Pond	3,520	Elias 1983, 1985
5	Beaver Meadows	2,530	Doerner, Brunswig, and Lane 2001
			Doerner 2004
6	Lock Vale	3,322	Nash 2000
7	Sky Pond	3,320	Reasoner and Jodry 2000
8	Redrock Lake	3,100	Pennak 1963
			Maher 1972
9	Lefthand Reservoir	3,224	Elias 1983, 1985
10	Long Lake	3,210	Short 1985
11	Lake Isabelle Bog	3,310	Elias 1983, 1985
			Short 1985
12	Caribou Lake Valley	3,410	Benedict 1973,1981
			Davis and Osborn 1987
13	Arapaho Valley		
	(Butterfly and Triple lakes)	3,474	Benedict 1973,1981
			Davis and Osborn 1987
14	Fourth of July Valley	3,415	Benedict 1973,1981
			Davis and Osborn 1987
15	Devlin Lake	2,953	Legg and Baker 1980
			Madole 1986
16	Mary Jane		
	(Frazier River Valley)	2,882	Nelson et al. 1979
			Short and Elias 1987
17	Echo Lake (Mt. Evans)	3,230	Doerner 1994
			Doerner, Sullivan, and Briles 1998
18	Lamb Spring	1,713	Elias 1986, 1996a
			Elias and Toolin 1990
19	Lost Park Meadow	3,079	Vierling 1998

advancement in the Indian Peaks region estimated to have occurred between ca. 12,000 and 11,000 RCYBP. Locally, this advance is known as the 'Satanta Peak' advance. Menounos and Reasoner (1997) also provide evidence from Sky Pond (3,320 m) in Rocky Mountain National Park for a limited glacial advance that correlates to the 'Satanta Peak' advance. They attribute this advance to a short-term cooling associated with the Younger Dryas chronozone. This conclusion is supported by a recent geomorphic study by Nash (2000), working in the nearby Loch Vale area of Rocky Mountain National Park. He used geomorphic mapping and clast analysis to infer that glacial advances associated with the Younger Dryas chronozone in the Loch Vale area (approximately 3,322 m) were between a few meters to 0.5 km in extent.

The Holocene

Holocene glacial deposits are found in many cirques throughout Colorado (Burke and Birkeland 1983). Davis (1988) suggested that alpine areas in Colorado provide some of the best-dated and most detailed Holocene glacial chronologies in the American Cordillera. The majority of the Colorado chronologies are from northern Colorado sites whose chronologies are briefly discussed below.

Benedict (1973, 1981) reconstructed a chronology of Holocene advances for cirques in the Front Range. Working at three sites (Caribou Lake Valley, Fourth of July Valley, and Arapaho Valley) in the Indian Peaks region, Benedict found evidence of at least four intervals of glacier expansion. The first is termed the 'Ptarmigan' advance, dating to ca. 7,500 to 6,400 RCYBP. The second advance is the 'Triple Lakes,' which occurred between ca. 5,000 and 3,000 RCYBP. The 'Audubon' advance was the third expansion, which occurred between ca. 1,850 and 950 RCYBP. The most recent advance was the 'Arapaho Peak' advance, dated between ca. 350 and 150 RCYBP.

Birkeland and Shroba (1974) used relative-dating methods to suggest that the Triple Lakes advance could be as old as the Satanta Peak advance. Davis (1988) pointed out that the radiocarbon ages described by Benedict are only minimum-limiting ages. Davis and Osborn (1987) used radiocarbon dates obtained from Butterfly Lake, inside the Triple Lakes moraine, to suggest that the Triple Lakes advance in the Arapaho Valley could be equivalent in age to the Satanta Peak advance in the Caribou Lake Valley.

The most common Holocene glacial deposits occurring in the Front Range date to the "Little Ice Age" (ca. A.D. 1350–1850) period. These glacial advances reached their maximum extent by the mid-nineteenth century and have been retreating ever since (Elias 2001). Despite some of the best-dated (radiocarbon) glacial chronologies in the Rocky Mountains, the nature of Holocene glacier fluctuations in Colorado remains unresolved.

PALEOECOLOGICAL STUDIES

Environmental changes during the late Quaternary were substantial. Paleoenvironmental research is essentially an interdisciplinary endeavor, drawing from fields as diverse as archaeology, botany, climatology, geography, and geomorphology. The ultimate aim of such research is to collect and synthesize information in an effort to reconstruct the past and explain the present. In this section, results of paleoecological studies conducted in the Front Range are summarized (Table 1.3; Figure 1.4).

The Late Pleistocene

Limited information is available concerning vegetation and climate from Colorado during and subsequent to Pinedale glaciation (Pennak 1963; Maher 1972; Legg and Baker 1980; Elias 1983, 1985; Short 1985; Short and Elias 1987).

This is the case in part because of the absence of research sites (lakes and fens) with a continuous record of deposition extending from the late Pleistocene to the present. Organic-rich deposits are relatively scarce in Colorado (Elias 2001). At lower elevations, Holocene aridity has resulted in desiccation of natural lakes and wetlands. In alpine regions, organic deposits were stripped and drainages were reconfigured as glaciers advanced and retreated. Additionally, basins dammed by terminal moraines were directly influenced by upstream glaciers. Glacial ice tends to collect significant amounts of pollen over time, and as glaciers melted, pollen was released into meltwater streams and transported downstream. Lakes, which received glacial runoff, were consequently loaded with anomalously high concentrations of pollen. This effect has tended to dilute the climatic signal, and, as a result, it is frequently difficult to separate "noise" from the true signal. In addition, glacial sediments may be reintroduced into basins long after glaciers have retreated, thus complicating their stratigraphic records.

At Devlin Lake in the Indian Peaks Wilderness Area, Legg and Baker (1980) analyzed a sediment core recovered from a late-glacial–aged lake. The core contained pollen and a sequence of varved sediments with radiocarbon dates from ca. 22,400 to 12,200 RCYBP. The date of ca. 12,200 RCYBP was interpreted to be the minimum age for the retreat of glacial ice in the area. The pollen from this core was dominated by sagebrush (40–60 percent) and pine (10–25 percent). Other pollen types that were consistently present but in lesser amounts included grass (5–17 percent), goosefoot and amaranth (3–17 percent), daisy (5–10 percent), juniper (2–10 percent), and spruce (1–4 percent). The pollen data indicate that an open tundra environment prevailed during the full-glacial period. During this time, Devlin Lake was about 100 m above tree line. This implies that tree line was about 500 m lower than its present elevation during the last glacial maximum.

Elias (1986, 1996a, 1996b) and Elias and Toolin (1990) used fossil beetle assemblages to reconstruct late Pleistocene temperature parameters for the Lamb Spring site (1,713 m), located 3 km east of the mountain front on the piedmont zone just south of Denver. A beetle assemblage dated to ca. 14,500 RCYBP was found to reflect full-glacial conditions. The reconstructed mean July temperatures were estimated as 10–11°C colder than present, while reconstructed mean January temperatures were 26–30°C colder than present.

At the Mary Jane site (2,882 m), Short and Elias (1987) used fossil insects and pollen from a series of alternating lake sediments and glacial tills to reconstruct late-glacial conditions. The oldest lake sediments were dated to ca. 30,000 RCYBP (Nelson et al. 1979) and were overlain by glacial till. The next-youngest lake bed deposits yielded a radiocarbon age of ca. 13,750 RCYBP. Prior to the last Pinedale glacial advance (ca. 30,000 RCYBP), the vegetation was characterized as an open spruce-fir forest (Elias 2001). This was followed by a colder phase, in which alpine tundra replaced the subalpine forest. The existence of tundra vegetation at the site translates into a depression of upper tree line by more than 500 m. Late-glacial warming at Mary Jane appears to have begun sometime between ca. 13,800 to

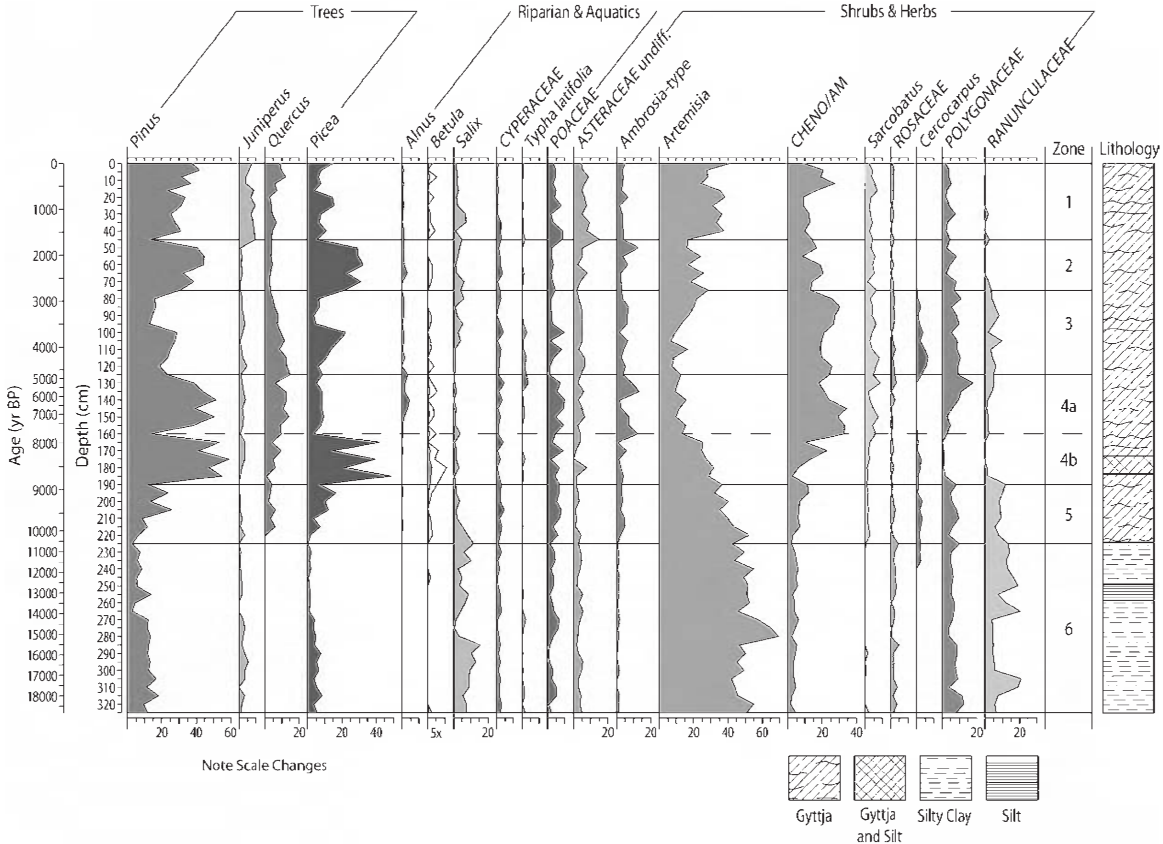
12,300 RCYBP (Short and Elias 1987). Its fossil assemblages indicate that summer temperatures had risen well above full-glacial levels by this time and were only 3–4°C cooler than modern values (Elias 1996b).

In the Mt. Evans region, Doerner (1994) used pollen and sediment analysis from Echo Lake (3,230 m) to reconstruct an 18,500-year record of paleoenvironmental change. The pollen diagram from this site indicates that during the last glacial maximum, sagebrush along with other shrubs and herbs dominated the vegetation (Figure 1.5). The limited representation of tree pollen (pine 15 percent, spruce 5 percent) indicates that tundra was probably much more extensive than at present. Colder temperatures depressed tree line by at least 300 m. By ca. 14,000 RCYBP, pollen data show a subtle change in response to climatic warming. A decrease in sagebrush pollen signals the onset of deglaciation and change toward post-glacial environmental conditions. At that time, tree line was still well below the elevation of the lake, and tundra plants continued to dominate the landscape.

The climatic warming that characterized the terminal Pleistocene was interrupted by a brief but intense cool period. Reasoner and Jodry (2000) provide paleobotanical evidence from Sky Pond (3,320 m) that shows a clear and immediate response to cooling associated with the Younger Dryas chronozone (ca. 10,800 and 10,100 RCYBP). They found that tree line was displaced downward between 60 and 120 m in elevation. This change in the position of tree line resulted from a cooling of about 0.4–0.9°C in summer temperature. The Younger Dryas oscillation was expressed in the percentages and accumulation rates of both spruce and fir pollen. In addition, this oscillation was reflected in the signal from two other taxa, oak and pine, which represent regional conditions.

The Holocene

Vierling (1998) analyzed the pollen in a sediment core extracted from Lost Park (3,079 m), a dry meadow in the Tarryall Mountains. This core was divided into three pollen zones and produced a 12,000-year record of environmental change. The lowest zone (ca. 12,000 to 9,000 RCYBP) is characterized by relatively high values of spruce, pine, and sedge. The high percentages of conifer pollen indicate the presence of a spruce-pine forest in the region. Stratigraphy in this zone suggests a wetland existed in the valley bottom, and the high values of sedge pollen support this idea. Vierling concluded that during this period temperatures were cooler than present, with a winter-dominated precipitation regime. The middle zone (ca. 9,000 to 1,800 RCYBP) shows increasing percentages of pine and Chenopodiaceae-type (goosefoot family) pollen. Beginning around 9,100 RCYBP there was a shift to warmer conditions accompanied by an increase in summer (monsoonal) precipitation. There was also a general decline in sedge pollen along with a stratigraphic change from peaty to alluvial sediments. This implies an increase in fluvial activity that resulted in scour and fill of the valley floor. Pine pollen percentages peaked between ca. 7,800 and 6,000 RCYBP,



1.5. Echo Lake pollen diagram.

indicating that maximum Holocene temperatures occurred at this time. The period between ca. 6,000 and 4,000 RCYBP was characterized by drier summers, less monsoonal precipitation, and frequent occurrences of charcoal fragments in the sediment. This suggests the onset of drier summer conditions accompanied by a weakening of summer monsoonal circulation. The upper zone (ca. 1,800 RCYBP to the present) represents the establishment of modern climate conditions. Spruce becomes a more important taxon, as indicated by rising ratios of spruce to pine pollen. Sagebrush pollen percentages also increased in this zone. Sedge percentages remained low; however, the core's stratigraphy shows that peat replaced the alluvial sediments of the prior zone. Sedge percentages may be artificially low because of an increase in the percentage of spruce and sagebrush pollen. The late Holocene Period is characterized by cooler conditions.

The Echo Lake site (3,230 m) shows significant post-glacial warming had begun by about ca. 10,200 RCYBP (Doerner 1994). Warming is indicated by increases in arboreal (tree) pollen percentages with concomitant declines in non-tree pollen types, reflecting expansion of coniferous forests at the expense of tundra (Figure 1.5). Between ca. 10,200 and 9000 RCYBP, pine and spruce pollen increased as sagebrush pollen declined. Sagebrush along with plants from the goosefoot and amaranth group are shade-intolerant species, and their decline indicates a gradual reduction in the spatial extent of tundra and the establishment of an open pine-spruce forest. The increase in pine and spruce pollen suggests that tree line was near the lake but likely remained below its elevation. There were also increases in oak and mountain mahogany pollen, which indicates warmer conditions at lower elevations. These taxa do not grow at higher elevations in the Front Range; therefore, they likely represent upslope transport of pollen into the basin. The change in the pollen spectra was accompanied by a depositional change (from inorganic silty clay to organic-rich lake sediment). The increase in organic-rich sediment is interpreted as a signal of increased temperatures. As conditions became more favorable (i.e., warmer and wetter), there was greater organic productivity in the lake and on the surrounding landscape.

The Echo Lake pollen spectra between ca. 9,000 and 4,800 RCYBP suggest that the climate of this period was warmer than the present climate because pine dominates the pollen rain and non-arboreal pollen types have the lowest values in the core. The reduction of sagebrush indicates that tree line had risen to an elevation above the lake. The expansion of spruce and the decline in sagebrush suggest that warmer conditions existed at higher elevations. Maximum warming, as indicated by peaks in tree pollen (spruce and pine), occurred between ca. 9,000 and 7,800 RCYBP. After that time, conditions became slightly cooler and drier, but temperatures remained warmer than the present. Another significant change occurred between ca. 4,800 to 2,900 RCYBP and marked the onset of regional climatic cooling. Pine pollen dropped sharply in this zone, and spruce percentages showed an initial increase, followed by a significant decline. Chenopodiaceae-type percentages remained high (about 20 percent), and cattail

Table 1.4. Echo Lake Vegetation and Climate History.

Pollen Zone	*Age Range* (RCYBP)	*Vegetation*	*Inferred Climate*
1	ca. 1,600 to present	Modern forest (pine and spruce decrease, sagebrush and juniper increase)	Cool and dry
2	ca. 3,000 to 1,600	Forest expansion/higher tree line (pine and spruce increase)	Warm
3	ca. 4,800 to 3,000	Forest thinning (oak and sagebrush increase)	Cool
4a	ca. 7,800 to 4,800	Open pine forest (pine increases, spruce declines)	Warm and dry
4b	ca. 9,000 to 7,800	Tree line expansion (spruce increases)	Maximum warming
5	ca. 10,200 to 9,000	Establishment of subalpine forest	Cool and moist (with increasing temperatures)
6	ca. 18,500 to 10,200	Tundra (sagebrush dominated)	Cold and dry

pollen peaked. Increases in pollen from these taxa suggest dry conditions, fluctuations in lake levels, or both. There were also increases in sagebrush pollen along with pollen from plants in the buckwheat, buttercup, and rose families. Increases in pollen from these plants suggest cooler conditions and the lowering of tree line, the thinning of the forest, or both. Pollen from oak and mountain mahogany also increased at this time. These taxa are not local, and their expansion probably represents a change to more favorable environmental conditions in the foothills to the east of the site.

Climatic warming occurred between ca. 2,900 to 1,500 RCYBP, as the percentages and absolute influx rates of pine and spruce rise rapidly. Sagebrush pollen declines slightly, as does the pollen from the buckwheat, buttercup, and rose families. The decline in these taxa implies either forest expansion or higher tree line. During the last 1,500 RCYBP, the pollen spectra indicate a general cooling trend. Pine and spruce frequencies decline while sagebrush pollen increases. This indicates cooler or drier conditions and the possible retreat (lowering) of tree line to its modern position. Juniper pollen achieves maximum representation during this time. This period represents the establishment of modern conditions for the Mt. Evans region. Table 1.4 summarizes the vegetation and climatic changes reconstructed from the Echo Lake site.

Pennak (1963) analyzed pollen from four sites in the Front Range, ranging in elevation from 2,617 to 3,247 m. Two of the sites were mountain lakes and two were dry meadows. The most complete pollen data were recovered from Redrock Lake (3,100 m), which yielded a 7,000-year record of environmental change. Because results from all four sites are consistent, only the Redrock Lake record is discussed here. The pollen assemblage recovered from this site is relatively simple, with pine, spruce, sagebrush, and grass usually accounting for more than 90 percent of all pollen grains. An open mesic (moist) forest surrounded the lake at ca. 7,000 RCYBP. Five hundred years later there was a dry boreal interval in which climatic conditions were considerably drier than at present. This interval was quickly replaced by a sharply defined warm, dry period that extended from ca. 6,000 to 3,000 RCYBP. Pennak interprets this period as the post-glacial interval, also known as the *Altithermal* (Antevs 1948). This period is notable for the dominance of grasses (35–40 percent), fewer pines (30 percent), and only about 10 percent spruce. Modern vegetation cover became established about 3,000 RCYBP and has remained essentially unchanged.

Maher (1972) reanalyzed the Redrock Lake site and was able to extend the Holocene record back to ca. 9,500 RCYBP. Maher's analysis suggests that between ca. 10,000 and 7,600 RCYBP, the climate was cooler, wetter, or both than the present, with tree line about 150 m lower. From ca. 7,600 to 6,700 RCYBP, both climate and the location of tree line were similar to present conditions. During the period from ca. 6,700 to 3,000 RCYBP, tree line was lower and the climate was cooler, wetter, or both than the present. Establishment of the modern vegetation cover and climate occurred at about 3,000 RCYBP.

Elias (1983, 1985) used fossil insect evidence from La Poudre Pass Bog (3,103 m), Lefthand Reservoir (3,224 m), Lake Isabelle Bog (3,310), and Mount Ida Pond (3,520 m) to establish a chronology of Holocene environmental change. Elias found that summer temperatures were warmer than the present during the early Holocene. Maximum ratios of forest-tundra taxa suggest that Holocene temperatures peaked between ca. 9,000 and 7,000 RCYBP. The climate remained warmer than the present until 4,500 RCYBP. At that time, the macrofossil records suggest cooling that lasted from 4,500 to 3,100 RCYBP. The climate warmed from ca. 3,000 to 2,000 RCYBP, followed by a general cooling trend that extends to the present.

Short (1985) analyzed the fossil pollen from three sites in the northern Front Range to test Maher's (1972) climatic reconstruction. Her sites included La Poudre Pass Bog (3,103 m), Lake Isabelle Bog (3,310 m), and Long Lake (3,210 m). The climatic signal from these three sites is consistent. Short interprets the period from ca. 12,000 to 10,500 RCYBP as a tundra environment dominated by sagebrush and grasses, with some birch and willow. Spruce and fir arrived by ca. 10,500 RCYBP at Long Lake and were present at La Poudre Pass by ca. 9,800 RCYBP, when peat accumulation started. Despite the presence of trees in the region, tundra vegetation was still dominant until ca. 9,000 RCYBP. From ca. 9,000 to 6,500 RCYBP, a spruce-fir forest replaced the tundra vegetation, and the climate was similar to or warmer than the present. At Lake Isabella Bog the climate was suitable for the initiation of peat growth and accumulation between ca. 8,000 and 7,000 RCYBP. Short suggests that an increase of lodgepole pine during this period indicates warmer temperatures. Maximum upward extension of the tree line and maximum temperatures for the region occurred between ca. 6,500 and 3,500 RCYBP, when pine achieved its maximum representation. After ca. 3,500 RCYBP, there was a decrease in arboreal pollen representation and an increase in shrub and herb pollen (especially sagebrush and grass). This suggests that tree line was either lowered slightly or that forests thinned and that the decrease in pine may have resulted from a climatic cooling. Modern vegetation distributions were established sometime after ca. 3,500 RCYBP.

Paleoecological studies from Rocky Mountain National Park also provide evidence of Holocene environmental change (Doerner, Brunswig, and Lane 2001; Doerner, Brunswig, and Sanborn 2002). These pollen studies are in broad agreement. The middle Holocene Period (ca. 7,400 to 5,400 RCYBP) was characterized by warm and possibly dry conditions as compared to the present day. At higher elevations (Lawn Lake Fen, 3,357 m), warmer summers and longer growing seasons resulted in rapid peat growth. At lower elevations (Beaver Meadows, 2,530 m), increased temperatures and lower effective precipitation led to the disappearance of wetlands from valley bottoms. After ca. 5,400 RCYBP, regional climate change brought cooler conditions to the area. Modern vegetation associations and climate conditions were likely established by ca. 1,800 RCYBP. Tree-ring studies from the park show that tree invasion and establishment at upper tree line (about 3,414–3,511 m) are likely related to the development of favorable climatic conditions (warmer and wetter) since the end of the "Little Ice Age" (ca. A.D. 1850) (Hessl and Baker 1997a, 1997b; Hessl, Baker, and Weisberg 1996).

Maher's early and middle Holocene reconstruction contrasts with other paleoecological records from the region. An alternative interpretation is offered by Nichols (1982), who notes that a peak in the absolute influx of spruce pollen occurred at 8,500 RCYBP in the Redrock Lake core. High spruce values at this time are also reported at other sites in Colorado (Andrews et al. 1975; Peterson and Mehringer 1976; Short 1985; Doerner 1994). Temperatures during this period are interpreted as warmer than the present. Nichols remarked that spruce values in the Redrock Lake core fluctuated during the middle Holocene until a marked decline occurred between 4,000 and 3,000 RCYBP. This suggests that cooler conditions were established at that time. If one accepts Nichols's reinterpretation of Maher's data, then the Redrock Lake record is in agreement with other Front Range reconstructions.

DISCUSSION

The Late Pleistocene

The late-glacial paleoecological records from the Front Range reflect high sagebrush pollen with moderate values of pine, grass, and other herb taxa. These records are in general agreement with many of the pollen records from the western United States, which are characterized by high frequencies of non-arboreal pollen taxa, including sagebrush, juniper, and grass (Beiswenger 1991; Davis and Pitblado 1995; Doerner and Carrara 2001; Fall, Davis, and Zielinski 1995; Mehringer 1985; Whitlock 1993). The vegetation reconstructions for this period indicate that an open tundra environment existed throughout the Front Range

(Legg and Baker 1980; Short and Elias 1987; Doerner 1994). Elias (1995) suggests this vegetation likely developed under cold, dry conditions similar to the climate found in parts of northern Siberia today.

Paleotemperature reconstructions developed from fossil beetle assemblages suggest that during the full glacial (ca. 14,500 RCYBP), mean July temperatures were as much as 10–11°C colder and mean January temperatures were depressed by 26–29°C compared with modern parameters (Elias 2001). Temperatures were cold enough to support the growth of glacial ice, but there was likely insufficient winter precipitation to allow for glacier expansion. Colder summer temperatures depressed the upper tree line by as much as 300 m below present-day limits on Mt. Evans (Doerner 1994). At the Devlin Lake and Mary Jane sites, upper tree line was depressed by 500 m below modern limits.

While complete agreement is lacking, it appears deglaciation in the Front Range began between ca. 15,000 and 12,000 RCYBP (Nelson et al. 1979; Legg and Baker 1980; Short 1985; Madole 1986). The transition from full-glacial conditions to warmer post-glacial conditions began sometime after ca. 14,000 RCYBP. Fossil insect assemblages from the Mary Jane site indicate that summer temperatures had risen well above full-glacial levels by ca. 13,000 and were only 3–4°C cooler than modern values (Short and Elias 1987; Elias 1996b). Pollen data from several sites indicate that vegetation was responding to climatic warming, but tundra plants still dominated the landscape (Short 1985; Doerner 1994).

Post-glacial warming that began after ca. 14,000 RCYBP was interrupted by a dramatic but short-term period of renewed glaciation and cooling soon after 11,000 BP (Benedict 1973, 1981; Birkeland and Shroba 1974; Davis and Osborn 1987; Menounos and Reasoner 1997; Nash 2000; Reasoner and Jodry 2000). This climatic oscillation is associated with the Younger Dryas chronozone, a pronounced 1,000-year cool period beginning about 11,000 radiocarbon years ago. Termination of this event appears to have resulted in rapid warming and the appearance of essentially modern-equivalent climatic conditions associated with the Holocene.

The Holocene

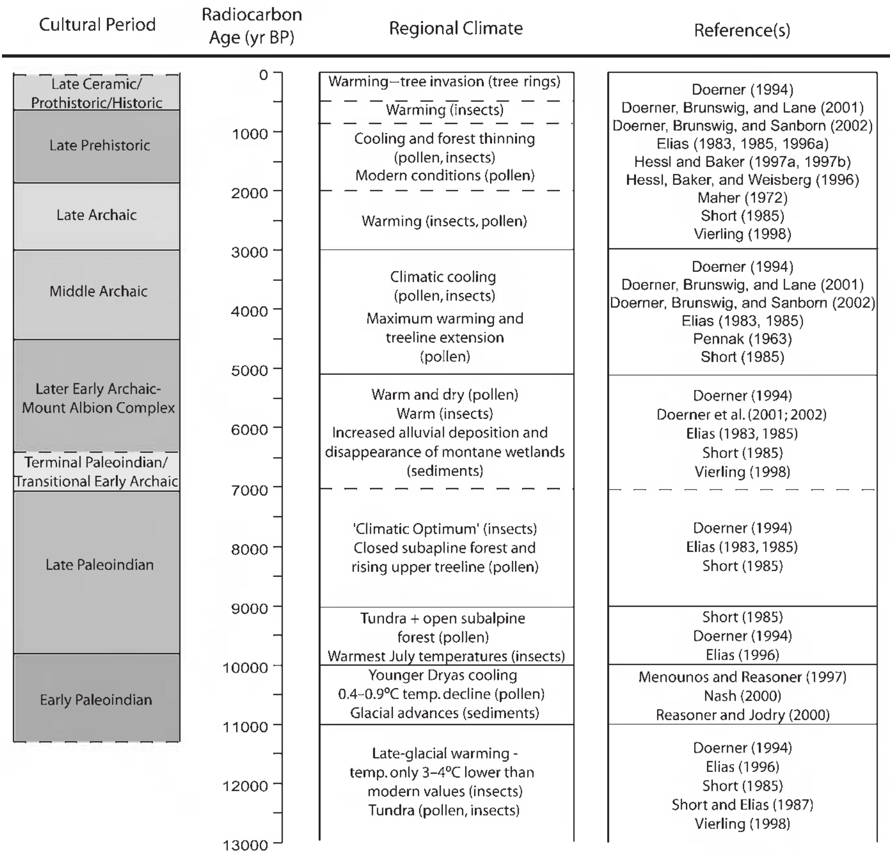
During the Holocene, Colorado experienced a series of climatic fluctuations. Strong evidence suggests the time interval from 10,000 to 8,000 BP was the warmest period of the Holocene. Research from numerous sites in western North America supports this premise (e.g., Kearney and Luckman 1983; Ritchie, Cwynar, and Spear 1983; Hebda and Mathewes 1984; Davis, Sheppard, and Robertson 1986; Carrara, Trimble, and Rubin 1991; Doerner and Carrara 1999). This period has been named the "early Holocene Xerothermic" (Hebda and Mathewes 1984). A possible explanation for the early Holocene thermal maximum is provided by Kutzbach (1983). From about 10,000 to 9,000 BP, orbital variations produced a solar radiation maximum during the Northern Hemisphere summer. Kutzbach (1983: 274) states that "about 9000 RCYBP, obliquity was 24.23° (compared to 23.45°

at present), perihelion was in Northern Hemisphere summer (July 30 compared to January 3 at present), and eccentricity was 0.0193 (compared to 0.0167 at present). These factors combined to produce increased solar radiation for July and decreased radiation for January."

July insolation was more than 8 percent higher than today, resulting in temperatures in the interior of North America 2–3°C higher than at present (Kutzbach and Guetter 1984, 1986). This greater summer radiation increased temperatures and decreased effective moisture, producing vegetation assemblages more xerothermic than those of present-day environments (Barnosky 1989). Trees responded to this warming by migrating upslope to higher elevations. By ca. 9,000, subalpine forests were well established throughout the Front Range, and tree line reached up to and perhaps beyond modern-day limits.

The period from ca. 8,000 to 4,500 RCYBP was warm, with possibly dry summers; however, this period was not as warm as the early Holocene. This contradicts the traditional view that the middle Holocene (ca. 7,000 to 4,000 RCYBP) was the warmest period of the Holocene. Drier summer conditions during the middle Holocene are indicated by changes in pollen assemblages and meadow stratigraphy. Lower-elevation sites in the Front Range (i.e., Lost Park Meadow [Vierling 1998], Beaver Meadows [Doerner 2004; Doerner, Brunswig, and Lane 2001], and Doolittle Ranch [Doerner 1994]) contain stratigraphic sequences in which peaty sediments overlie alluvial sediments. This suggests that scour and fill was occurring on the valley floors, as summer moisture levels were not high enough to support wet meadows. Less effective precipitation (e.g., greater evaporation rates) at lower elevations could account for the disappearance of the wet meadows.

Warmer-than-modern conditions prevailed in the Front Range until the late Holocene (ca. 4,800–4,500 RCYBP), when regional climate change brought cooler conditions to the area. The regional pollen signal shows a decrease in tree pollen and an increase in shrub and herb pollen (especially sagebrush and grasses). This suggests that either tree line lowered or forests thinned in response to colder summer temperatures. Echo Lake pollen data indicate that this cooler interval lasted from ca. 4,800 to 2,900 RCYBP; however, fossil beetle data suggest the cooling occurred between ca. 4,500 and 3,100 RCYBP. This timing difference may be a result of the interpolation of radiocarbon ages between well-dated intervals in the cores or of differences in response times to environmental change between plants and insects. This cold period was interrupted by a warm interval that lasted between 1,000 and 1,400 years (pollen data, ca. 2,900 to 1,500 RCYBP; fossil beetles, ca. 3,000 to 2,000 RCYBP). Cooler conditions again returned to the Front Range sometime between ca. 2,000 and 1,500 RCYBP, as modern conditions were likely established. The most recent millennium is characterized by warmer than modern ("Medieval" warming), to cooler than modern ("Little Ice Age" cooling), and back to warmer temperatures. Most of the proxy data regarding these most recent climatic reversals is derived from fossil insect assemblages,



1.6. Correlation of cultural-chronological periods with paleoenvironmental reconstructions from the Colorado Front Range.

glacial reconstructions, and tree-ring studies, as the pollen records from the Front Range have generally been insensitive to these short-term fluctuations.

CONCLUSION

From this review of paleoenvironmental studies of the Colorado Front Range, it is clear that substantial changes in climate and vegetation occurred during the late Quaternary. Figure 1.6 provides a summary of paleoenvironmental conditions from the terminal Pleistocene to Historic times correlated with cultural-chronological periods for the Front Range region.

During the late Pleistocene, climate in the Front Range was cold and dry, and vegetation was dominated by tundra. Global cooling associated with full-glacial times gave way to warming between ca. 14,000–11,000 years ago, when glaciers in Colorado began to melt and retreat. This warming was interrupted by a brief but intense cool period known as the Younger Dryas. This nearly 1,000-year event displaced tree line downward between 60 and 120 m and allowed cirque glaciers to re-advance. Post-glacial warming resumed by ca. 10,000 RCYBP, and vegetation responded by advancing upslope. About 9,000 years ago, warming intensified in response to increases in summer insolation. By ca. 8,000 years ago, warmer conditions were well entrenched in the region, with cycles of intense to more moderate warming and aridity, nearly always warmer and drier than the present. Eventually, by the late Holocene (ca. 4,500 years ago), broadly modern climatic parameters were established, although several less intense cycles of cooling ("neo-glaciations") and warming are known to have existed at various times throughout this period. The most recent cold phase was that of the so-called Little Ice Age that began in the fourteenth century A.D. and only ended in the mid-nineteenth century A.D., with increasingly warm conditions of the modern era.

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