Late Quaternary Glacial and Vegetative History of the Glacier National Park Region, Montana

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Cover. Agassiz Glacier and Kintla Peak (3,080 m) from the Boulder Pass area. Photograph by W.C. Alden (no. 705), August 5, 1913. U.S. Geological Survey Photo Library, Denver, CO 80225.

LATE QUATERNARY GLACIAL AND VEGETATIVE HISTORY OF THE GLACIER NATIONAL PARK REGION, MONTANA

Frontispiece. Agassiz Glacier and Kintla Peak (3,080 m) from the Boulder Pass area, August 17, 1981.

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By PAUL E. CARRARA

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Late Quaternary Glacial and Vegetative History of the Glacier National Park Region, Montana

By Paul E. Carrara

Abstract

Deglaciation was at least 90 percent complete in the Glacier National Park region by the time the Glacier Peak G (11,200 yr. BP) and Mount St. Helens Jy (11,400 yr. BP) ashes were deposited. The presence of these two ashes at Marias Pass, on the Continental Divide, indicates that the Marias Pass area was deglaciated before 11,400 yr. BP. The fact that the Two Medicine Glacier, a large piedmont glacier on the plains of Montana, was supplied by ice that flowed through Marias Pass indicates that this glacier retreated from the plains of Montana long before 11,400 yr. BP. Hence, by that time, former tributary glaciers along the eastern side of the Lewis Range in Glacier National Park no longer flowed east beyond the mountain front but were confined to their respective mountain valleys.

On the west side of Glacier National Park, the large trunk glacier that filled the valley of the North Fork Flathead River had retreated at least 10 km north of the site of the present-day town of Polebridge by the time the Glacier Peak G and Mount St. Helens Jy ashes were deposited. Similarly, the valley of the Middle Fork Flathead River was ice-free by the time the Glacier Peak G ash was deposited. Furthermore, stratigraphic evidence in both valleys suggests that the large trunk glaciers that occupied these valleys may have retreated from these areas several hundred to several thousand years before the deposition of the Glacier Peak G and Mount St. Helens Jy ashes.

By 11,200 yr. BP, glaciers in the Lewis, Livingston, Whitefish, and Flathead Ranges that were formerly tributary to the North and Middle Fork trunk glaciers were confined to their local mountain valleys. This conclusion is indicated by a radiocarbon age of 11,150±90 yr. BP (W–5043) from a bog on a moraine at the lower end of Bowman Lake and by the presence of the Glacier Peak G ash in a bog on a moraine deposited by the former McDonald glacier about 1 km southeast of the site of the present-day town of West Glacier.

By 11,000 to 10,000 yr. BP, remaining glaciers in Glacier National Park were probably confined to the same cirques and well-shaded niches where present-day glaciers and snowfields lie. Although evidence was not found in the park, data from other areas in the Rocky Mountains suggest that by as early as 11,000 yr. BP remaining glaciers may have been similar in size to present-day glaciers. Radiocarbon ages near present-day glaciers in Banff and Jasper National Parks, Alberta, and Yoho National Park, British Columbia, indicate that by 10,000 yr. BP, late Wisconsin glaciers had receded to positions close to those of present-day glaciers. A similar amount of deglaciation by these times is inferred for the Glacier National Park region.

Initial revegetation of the Glacier National Park region had also occurred by the time the Glacier Peak G and Mount St. Helens Jy ashes were deposited. At Marias Pass, a conifer needle and an alder strobile were found beneath the Glacier Peak G ash, and several small willow fragments were found beneath the Mount St. Helens Jy ash. Pollen and insect macrofossils below the Mount St. Helens Jy ash at this site indicate that when this ash was deposited treeline was at least 500 m lower than present and possibly as much as 700 m lower.

Reforestation by at least 10,000 yr. BP is documented in several areas in Glacier National Park. In the McGee Meadow area, radiocarbon ages on lodgepole pine fragments indicate that this species was established by at least 10,000 yr. BP. Stratigraphic evidence from one bog in this area suggests that lodgepole pine may have been established as early as 11,000 yr. BP. In the Howe Lake area, a radiocarbon age on a wood fragment indicates that either spruce or larch was growing in this area by at least 10,600 yr. BP. In the Kootenai Lakes area, a radiocarbon age on lodgepole pine fragments suggests that this species was growing there by about 10,000 yr. BP. On Snyder Ridge, a radiocarbon age on gyttja containing coniferous wood fragments indicates the presence of conifers there by about 10,000 yr. BP.

Two groups of moraines have been identified fronting the present-day glaciers and snowfields in the higher areas of Glacier National Park. Mazama ash in the soil overlying moraines of the older group establishes a minimum age of 6,845 yr. BP for these moraines. Vegetative evidence and correlation with other moraines elsewhere in the Rocky Mountains suggest that the older moraines are latest Pleistocene in age. Whether these moraines were formed by the last advance or stillstand of Wisconsin glaciers or formed by a separate advance after the Wisconsin glaciation is not known.

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The younger group of moraines was formed by advances that culminated in the mid-19th century as indicated by tree-ring studies at several localities in the park. During that time, Glacier National Park contained more than 150 glaciers. The climatic cooling of the mid-19th century that produced these moraines, although mild compared to full-glacial conditions, was the most severe climatic deterioration in the Glacier National Park region since the end of the Wisconsin glaciation.

Since the mid-19th century, glaciers in Glacier National Park have shrunk dramatically. From the mid-19th century until about 1920 retreat rates were slow. From 1920 to the mid-1940's the Glacier National Park region experienced a period of above average summer temperatures and below average annual precipitation that caused the glaciers to retreat drastically; many disappeared altogether. After the mid-1940's, the overall retreat rates slowed. Between 1966 and 1979, several of the larger glaciers in the Mount Jackson area advanced slightly.

INTRODUCTION

Purpose and Methods

This report presents the late Quaternary glacial and vegetative history of the Glacier National Park region, northwestern Montana, based on field mapping, volcanic ash identification, radiocarbon age determinations, and tree-ring analysis. Field mapping was done during the summers of 1981 to 1985 as part of a U.S. Geological Survey project to update the geological knowledge of Glacier National Park.

Three volcanic ashes erupted from volcanos in the Cascade Range of the Pacific Northwest have been identified in postglacial deposits in the Glacier National Park region. These ashes are the (1) Mazama, (2) Glacier Peak G, and (3) Mount St. Helens Jy. The Mazama ash, erupted in the middle Holocene, is a useful time-stratigraphic marker because it is commonly found in bogs and exposures throughout the Glacier National Park region. In addition, because this ash is also found in soils overlying glacial deposits in the higher areas of the park, it provides a minimum age for these deposits that could not be radiocarbon dated because of a lack of associated organic material (Osborn, 1985; Carrara, 1987).

The Glacier Peak G and Mount St. Helens Jy ashes were erupted in the late Pleistocene, when the Glacier National Park region was undergoing extensive deglaciation. Because these ashes would not be preserved if they had fallen on glacial ice, the presence of these ashes at a site provides information concerning the extent of deglaciation at the time they were deposited (Carrara, 1986). Also, at sites where these ashes are associated with pollen, plant fossils, and insect macrofossils, information concerning environmental conditions during the late Pleistocene has been obtained (Carrara and others, 1986). Because of the problems associated with the radiocarbon dating of late Pleistocene events in the Glacier National Park region, the Glacier Peak G and Mount St. Helens Jy ashes are invaluable time-stratigraphic markers.

Thirteen radiocarbon ages were obtained from organic material collected from bogs and exposures in the Glacier National Park region. Radiocarbon dating of the lowermost organic material at these sites has yielded minimum ages for deglaciation and the beginning of organic sedimentation. Radiocarbon dating of wood fragments from some of these sites has yielded minimum ages for the establishment of shrubs and trees following deglaciation.

Cores were taken from trees within the forest trimlines fronting the Agassiz and Jackson Glaciers and also from trees immediately beyond the fresh, bouldery moraines of several other glaciers in the park. These cores were analyzed to date the glacial advance of the mid-19th century.

This report documents the presence of the three late Quaternary ashes in the Glacier National Park region and describes the stratigraphy at the sites of these ashes and (or) organic material on which radiocarbon age determinations were made. Next, this report discusses the implications of these ashes and radiocarbon ages to the late Quaternary history of the Glacier National Park region. The use of the Mazama ash to establish a minimum age for glacial deposits in the higher areas of the park is also discussed. In addition, this report documents the use of tree-ring analysis to date the minor glacial advance of the mid-19th century and subsequent recession.

Geology and Geography

Glacier National Park, located in northwestern Montana adjacent to the Canadian border and comprising parts of Flathead and Glacier Counties, covers about 4,000 km^2 (fig. 1). The park is an area of diverse topography underlain by rocks ranging in age from Precambrian to Tertiary (fig. 2), which are in many places mantled by surficial deposits of late Quaternary age (Carrara, 1989).

On the west, trending northwest to southeast, are the valleys of the North and Middle Forks Flathead River (fig. 1) whose valley floors range in altitude from about 1,280 to 975 m. These valleys have been down-dropped on the east and are underlain by about 5,000 m of the upper Paleogene Kishenehn Formation, consisting of lacustrine and fluvial sedimentary rocks (Constenius,

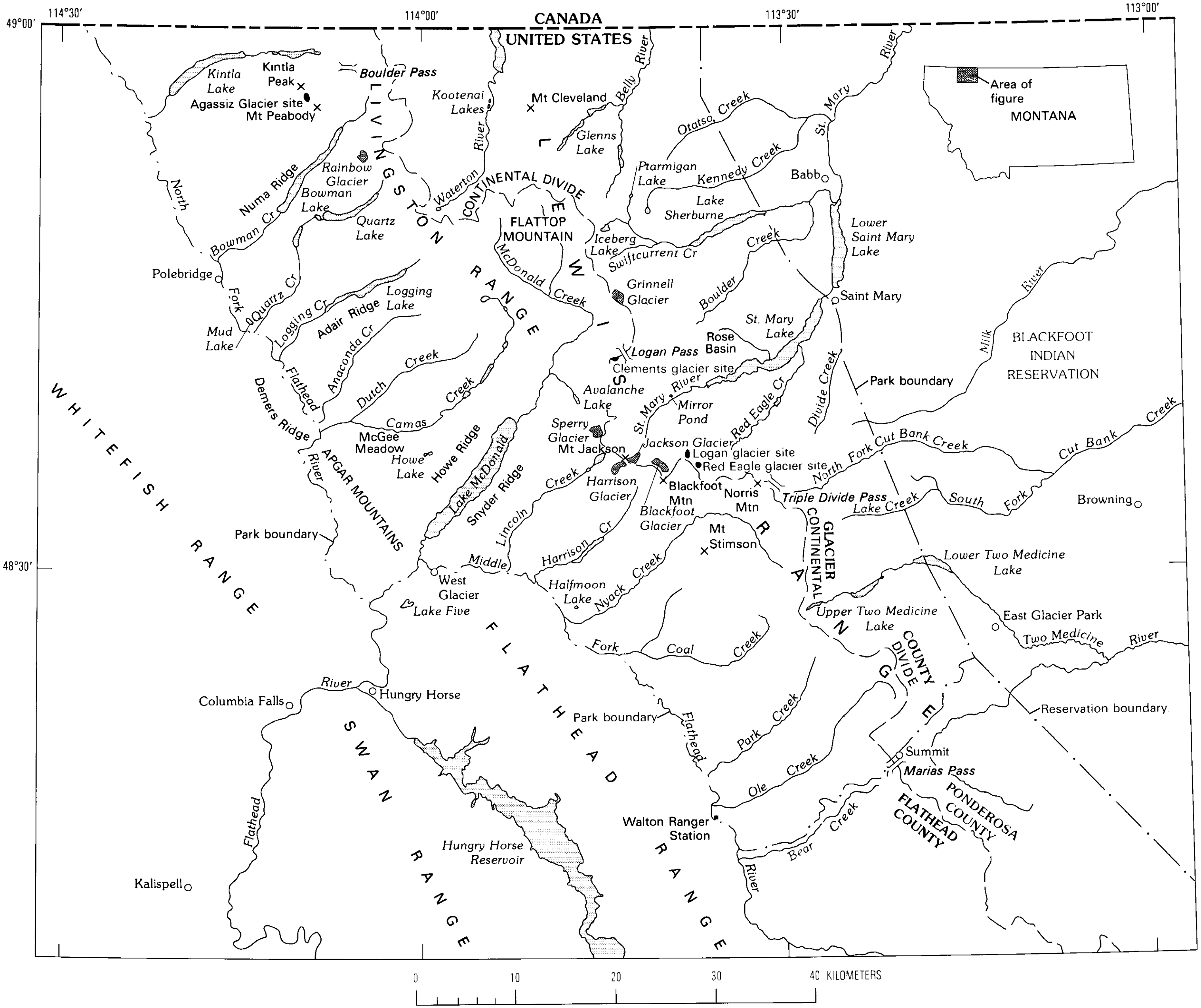

Figure 1. General location and place names in the Glacier National Park region, Montana.

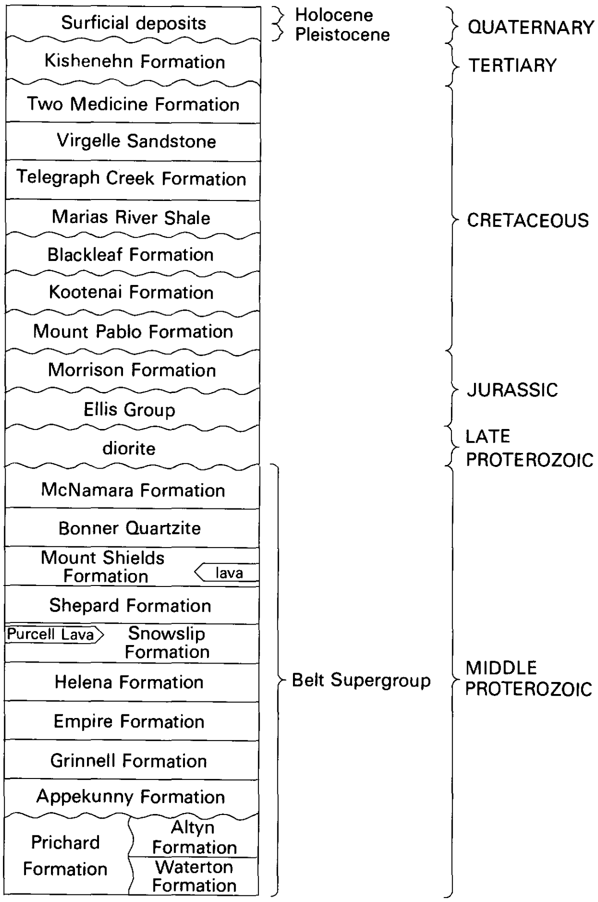

Figure 2. General stratigraphic section in the Glacier National Park region (modified from Ross, 1959; Mudge and Earhart, 1983; Raup and others, 1983; Whipple and others, 1984; J.W. Whipple, written commun., 1984; and Earhart and others, 1989).

1981). In some places, these rocks have failed, resulting in large landslides, such as the one in the Halfmoon Lake area. In other areas the Kishenehn Formation is covered by late Wisconsin till as much as 30 m thick. Depressions within this till contain many bogs filled with peats, organic muds, and volcanic ashes. Valley glaciers emerging from the western flank of the Livingston Range carved large, U-shaped valleys in the underlying Kishenehn Formation. Many of these valleys now contain large, deep lakes (including Kintla, Bowman, Quartz, Logging, and McDonald Lakes).

The central part of Glacier National Park is dominated by two rugged mountain ranges (fig. 1), trending northwest to southeast, that contain many small glaciers and snowfields. The Livingston Range on the west extends about 35 km from the Canadian border south to the Lake McDonald area. The Lewis Range on the east extends about 100 km from the border south through the park to Marias Pass. Large areas of these two ranges lie above timberline (about 2,000 m), and many peaks exceed 2,800 m in altitude. Mount Cleveland (3,191 m), about 8 km south of the border in the Lewis Range, is the highest peak in Glacier National Park. Local relief between many valleys and the surrounding summits is as much as 1,500 m. The Continental Divide follows the crest of the Lewis Range northward to about 16 km south of the border, then swings westward and follows the crest of the Livingston Range to Canada.

Both mountain ranges are underlain by Proterozoic sedimentary rocks of the Belt Supergroup (fig. 2) ranging from about 1,600 Ma to 800 Ma (Raup and others, 1983). These rocks consist of a sequence of argillites, siltites, and carbonates with a maximum stratigraphic thickness of about 5,200 m (Whipple and others, 1984). Although these rocks have been affected by low-grade regional metamorphism, many of their original sedimentary features (ripple marks, mud cracks, salt casts, and fossil algal stromatolites) are well preserved. In places, these rocks have been intruded by diabasic and gabbroic sills and dikes. Near the upper part of the section, mafic lava flows 10–60 m thick are present (McGimsey, 1985). In the Paleocene, this entire sequence of rocks was displaced about 65 km east over Cretaceous sedimentary rocks along the Lewis thrust (Raup and others, 1983). Today these rocks form a broad syncline whose axis closely parallels the Continental Divide (Ross, 1959).

The Lewis and Livingston Ranges contain various surficial deposits (Carrara, 1989). A large block slide is present immediately west of Avalanche Lake and several block slides are also present in the area north of Glenn Lake. Steep valley sides, whose forest is scarred by snow avalanche tracks, are mantled by colluvium. Talus cones and sheets are common at the base of steep slopes and cliffs. Late Wisconsin till mantles the floors of many of the mountain valleys. In the higher areas of both ranges, two groups of moraines front the present-day glaciers and snowfields (Osborn, 1985; Carrara, 1987; Carrara and McGimsey, 1988). On the high, unglaciated upland surfaces, stone stripes, sorted polygons, and solifluction deposits have been formed on the regolith by severe frost action (Carrara, 1989).

East of the Lewis thrust lies the western margin of the Montana plains. Here altitudes range from about 1,370 to 2,000 m. This area consists of low, rounded hills mantled in places with a veneer of glacial till and underlain mainly by a dark-gray marine mudstone of the Upper Cretaceous Marias River Shale (Mudge and Earhart, 1983). In many places immediately east of the Lewis thrust, Pleistocene valley glaciers gouged deep, steep-sided valley walls in these soft rocks as they emerged from the Lewis Range. These rocks have since failed and formed large landslides (Carrara, 1989).

Two large rivers, the North and Middle Forks Flathead River (fig. 1), drain the west side of Glacier National Park. The North Fork Flathead River, one of the major rivers of northwestern Montana, heads in British Columbia and flows southeast about 120 km, draining an area of about 4,200 km^2 (Erdmann, 1947). Streams draining the eastern flank of the Whitefish Range and the western flank of the Livingston Range in Glacier National Park (Kintla, Akokala, Bowman, Quartz, Logging, Anaconda, Dutch, and Camas Creeks) are tributary to this river. The Middle Fork Flathead River heads in the Lewis and Clark Range, south of Glacier National Park in the Great Bear Wilderness, and flows northwest about 100 km to the town of West Glacier, draining an area of about 2,900 km^2 (Erdmann, 1947). Streams draining the eastern flank of the Flathead Range and the western flank of the Lewis Range in Glacier National Park south of Lake McDonald (Lincoln, Harrison, Nyack, Coal, Park, and Ole Creeks) are tributary to this river. Near the town of West Glacier the Middle Fork Flathead River is joined by McDonald Creek, which drains a large area in the north-central part of the park enclosed by the Livingston Range on the west and the Lewis Range on the east. About 7 km southwest of the town of West Glacier, the North and Middle Forks Flathead River join and flow south 10 km to the town of Hungry Horse, where they are joined by the northwest-flowing South Fork Flathead River. This river, which heads in the Swan Range and the Lewis and Clark Range about 140 km southeast in the Bob Marshall Wilderness, drains an area of about 4,450 km^2 (Erdmann, 1947). The combined Flathead River then flows west about 3 km through Bad Rock Gap, where it turns south and flows about 65 km to Flathead Lake. The Flathead River is a tributary of the Clark Fork of the Columbia River.

The east side of Glacier Park is drained by several rivers that head along the Continental Divide (fig. 1). The Waterton, Belly, and St. Mary Rivers drain the northern and northeastern areas of Glacier National Park. These rivers flow north and northeast into Canada and are part of the Hudson Bay drainage system. The Two Medicine, Cutbank, and Milk Rivers drain valleys along the east flank of the Lewis Range south of St. Mary Lake. The Two Medicine and Cutbank Rivers flow east about 100 km to the Marias River. The Milk River flows northeast into Canada, then southeast and east back into the United States about 400 km to the Missouri River.

Climate

The Continental Divide separates the Glacier National Park region into two areas dominated by different climatic systems. The area west of the Continental Divide, which is the wettest part of Montana, is dominated mainly by Pacific air masses. However, during the winter, cold continental polar air masses from Canada sometimes pour over the mountains into this area. Also, in the spring and summer, air masses from the Gulf of Mexico sometimes invade this area (Cunningham, 1982). In this area, winter is the wettest season and snowfalls are quite heavy. However, both the towns of Polebridge and West Glacier record significant precipitation during May and June (U.S. Dept. of Commerce, 1982), reflecting the effect of the invasion of warm moist air masses from the Gulf of Mexico. The period of minimum precipitation on both sides of the park is July and August (Finklin, 1986).

The area east of the Continental Divide is dominated by cold, dry continental-polar air masses during the winter. Here winter precipitation is slight. From October to March, the towns of Babb and Browning receive only 28 percent and 33 percent, respectively, of their annual precipitation (U.S. Dept. of Commerce, 1982). During the winter, strong Chinook winds can occur when Pacific air masses invade this area. These air masses, which lose their moisture as they ascend the western slopes, warm adiabatically as they descend the eastern slopes, producing winds that may exceed 160 km/hr (Cunningham, 1982; Finklin, 1986). In this area, the wet season is from April through June and is caused by the incursion of air masses from the Gulf of Mexico after the northward retreat of the polar front in the spring (Cunningham, 1982).

These air masses from the Gulf of Mexico can cause extensive flooding when heavy rains from them fall on late-lying snowpack. For instance, in June 1964 as much as 37 cm of rain fell over a 30-hour period in the central region of Glacier National Park (U.S. Dept. of Commerce, 1964). The towns of Browning, East Glacier Park, and Summit (fig. 1) received about 20 cm of rain during this period. Record floods resulting from heavy rain on snowpack occurred in northwestern Montana on both sides of the Continental Divide. The floods were so severe that several dams were destroyed, including the Two Medicine Dam at Lower Two Medicine Lake (U.S. Dept. of Commerce, 1964).

Climatic data from several stations in the Glacier National Park region are presented in table 1. Some climatic data also exist for the higher areas of the park. Annual precipitation near the Grinnell Glacier (fig. 1) (2,135 m altitude) averaged 265 cm between 1950 and 1966. Annual precipitation in the Grinnell valley below the glacier averaged 375 cm between 1956 and 1966 (Dightman, 1961, 1967). A storage precipitation gage on Flattop Mountain (fig. 1), at an altitude of 1,920 m, indicates an average annual precipitation of 200 cm (Finklin, 1986), most of which falls as snow. Snow may fall in the higher areas of the park during any time of the year, and will often linger there until mid-summer.

Table 1. Climatic data (1951–1980) from stations in the Glacier National Park region[1]

[n.a., data not available]

Station	Altitude (m)	Average annual precipitation (cm)	Average Jan. temperature (°C)	Average July temperature (°C)
Kalispell	903	39.0	-5.5	19.1
West Glacier	962	76.0	-6.3	17.3
Polebridge	1125	59.3	-8.3	15.8
Babb	1311	47.6	n.a.	n.a.
Browning	1328	39.4	-8.4	16.8

[1]United States Department of Commerce (1982).

Forests

Forests in Glacier National Park reflect the two different climatic systems that affect this region. Because of the influence of Pacific air masses west of the Continental Divide, many trees common in the Pacific Northwest reach their eastern limits in the western part of Glacier National Park (Hansen, 1948). These trees include western hemlock (*Tsuga heterophylla*), western red cedar (*Thuja plicata*), western white pine (*Pinus monticola*), and Pacific yew (*Taxus brevifolia*). Grand fir (*Abies grandis*), also common in the Pacific Northwest, extends into Glacier National Park and areas farther to the southeast (Hansen, 1948). These trees, along with western larch (*Larix occidentalis*), Douglas fir (*Pseudotsuga menziesii*), western paper birch (*Betula papyifera*), and quaking aspen (*Populus tremuloides*) form dense forests in moist sites at lower altitudes. Douglas fir and Rocky Mountain juniper (*Juniperus scopulorum*) can also be found at dry sites along the valley of the North Fork Flathead River. Lodgepole pine (*Pinus contorta*) and ponderosa pine (*Pinus ponderosa*) are common at lower altitudes along the west side of the park, and usually grow in drier, open areas. Trees in the subalpine zone on the west side of the park include subalpine fir (*Abies lasiocarpa*), Engelmann spruce (*Picea engelmannii*), subalpine larch (*Larix lyalli*), and whitebark pine (*Pinus albicaulis*).

Forests in Glacier National Park east of the Continental Divide contain species also found in the southern Rocky Mountains of Colorado and New Mexico. These forests are more open and drier than those on the west side of the park. Trees here include lodgepole pine, quaking aspen, Douglas fir, limber pine (*Pinus flexilis*), subalpine fir, and Engelmann spruce. Lodgepole pine inhabit low and mid-altitudes and form almost pure stands in some areas. Aspen is found at lower altitudes in moist sites and also on the many landslide deposits east of the Lewis thrust. Douglas fir inhabit low, dry sites in this area. Limber pine is found at dry, windy sites at lower altitudes, such as along the shores of St. Mary and Lower St. Mary Lakes, as well as near upper timberline. Engelmann spruce and subalpine fir are found from higher altitudes to treeline.

Timberline in Glacier National Park is at an altitude of about 2,000 m. Above this altitude, thickets of dwarf, windswept (krummholz) subalpine fir and occasional Engelmann spruce extend upward to treeline, where they give way to tundra vegetation, at an altitude of about 2,050 m. Studies indicate that treeline in Glacier National Park is migrating upward (Habeck, 1969).

Thickets of thinleaf alder (*Alnus* sp.) and willow (*Salix* sp.) are found on both sides of the park. These shrubs commonly inhabit moist sites, such as stream banks and the margins of bogs.

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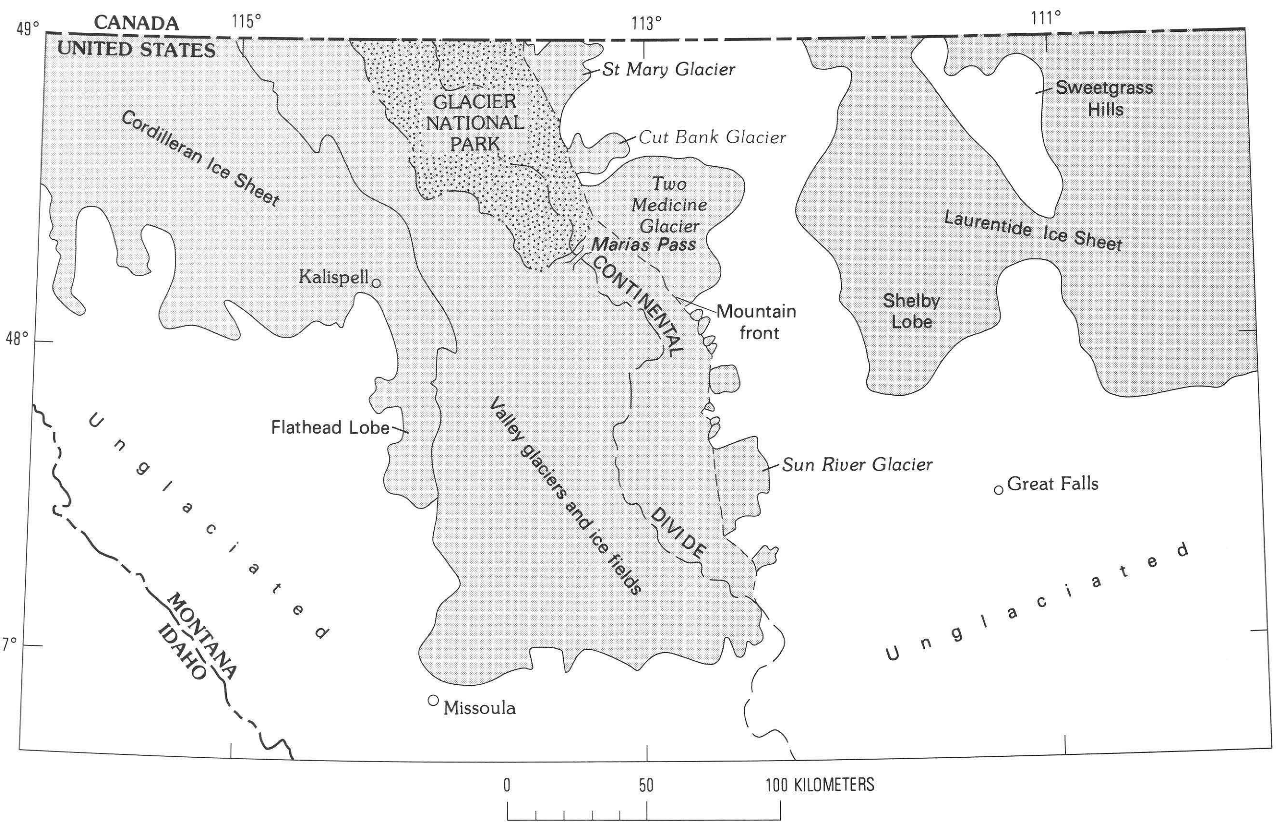


Figure 3. The Glacier National Park region showing late Wisconsin ice limits (modified from Alden, 1932, 1953; Calhoun, 1906; Colton and others, 1961; Mickelson and others, 1983; Richmond, 1986; and Waitt and Thorson, 1983).

Christensen and J.T. Quirk (Center for Wood Anatomy Research, Madison, Wisconsin) kindly identified wood fragments found in the various bogs and exposures. Emmett Evanoff (University of Colorado Museum) identified the fossil mollusks recovered from several of the bogs reported in this study. S.C. Porter (Quaternary Research Center, University of Washington) kindly provided samples of the Glacier Peak B and G ashes from the Trinity Mine site, Washington. D.M. Cheney and J.E. Jenness (U.S. Geological Survey) prepared the ash samples for chemical analysis; D.M. Cheney also provided the grain-size analyses. A.J. Bartel (U.S. Geological Survey) performed the X-ray fluorescence analyses. A.M. Sarna-Wojcicki analyzed the ash samples from Guardipee Lake. S.A. Elias and S.K. Short (Institute of Arctic and Alpine Research, University of Colorado) identified the fossil insects and pollen respectively from the Marias Pass exposure. This manuscript, or parts of it, benefited substantially from reviews by L.M. Carter, A.J. Donatich, W.R. Hansen, R.F. Madole, G.M. Richmond, V.S. Williams (U.S. Geological Survey), P.T. Davis (Bentley College), P.J. Mehringer, Jr. (Washington State University), and J.E. Beget (University of Alaska).

GLACIER NATIONAL PARK DURING THE LATE WISCONSIN GLACIATION

During the height of the late Wisconsin glaciation, much of northwestern Montana, including Glacier National Park, was covered by glacial ice (fig. 3). West of Glacier National Park lay the Cordilleran ice sheet, whose southern margin stretched from the Puget Sound area of Washington eastward to western Montana. This large ice sheet formed from the coalescence of large valley glaciers within the mountains of British Columbia. It then flowed south, overriding low mountain ranges in Washington, Idaho, and Montana, and sending ice lobes into the major valleys. The easternmost lobe of this ice sheet, the Flathead lobe, flowed south into the present-day Flathead Lake area (Alden, 1953; Richmond and others, 1965; Waitt and Thorson, 1983; Richmond, 1986a).

On the plains of Montana east of Glacier National Park lay the southwestern lobe of the Laurentide ice sheet. This ice sheet covered most of Canada, and its southern margin stretched from Montana east to the

Atlantic Ocean. The southwestern lobe of this ice sheet, the Shelby lobe, advanced south from Canada onto the plains of Montana (fig. 3) (Alden, 1932; Mickelson and others, 1983).

Between these two large ice sheets, in the mountains along the Continental Divide, valley glaciers and icefields covered the Glacier National Park region beneath hundreds of meters of ice (Alden, 1953; Richmond, 1986a). Only the higher peaks protruded above this ice cover as nunataks. Radiocarbon age determinations from the Colorado Rockies suggest that late Wisconsin valley glaciers reached their maximum extent in that area about 23,000 to 22,000 yr. BP (Rosenbaum and Larson, 1983; Madole, 1986). Obsidian hydration dating from the Yellowstone region suggests that late Wisconsin ice reached its maximum extent there as early as 30,000 yr. BP (Pierce and others, 1976; Pierce, 1979).

West of the Continental Divide in the Glacier National Park region, a large trunk glacier flowing southeast down the valley of the North Fork Flathead River was fed by glaciers flowing from valleys along the western flank of the Livingston Range and from valleys along the eastern flank of the Whitefish Range. Many of these tributary glaciers were 10–20 km long and several hundred meters thick. This large trunk glacier headed about 30 km north of the U.S.-Canadian border and filled the valley of the North Fork Flathead River with at least 1,000 m of ice. This ice thickness is indicated by striations and erratics along the crest of the Apgar Mountains and Demers Ridge (Alden, 1953), and by many sinuous meltwater channels, trending northwest-southeast, on the interfluves on the west side of the Livingston Range, such as Adair Ridge (Carrara, 1989). This ice overrode the Apgar Mountains, flowed southwest, and merged with the Flathead lobe of the Cordilleran ice sheet.

Another large trunk glacier filled the valley of the Middle Fork Flathead River. This glacier was fed by glaciers flowing from valleys along the western flank of the Lewis Range and from valleys along the eastern flank of the Flathead Range. This trunk glacier, which headed in the Lewis and Clark Range about 100 km to the southeast in the present-day Great Bear Wilderness, formed a local ice field west of Marias Pass. Some of this ice flowed northwest and joined the Lake McDonald glacier (which drained the Flattop Mountain area) and the large trunk glacier flowing down the valley of the North Fork Flathead River. This combined ice body then flowed south-southwest and merged with the Flathead lobe of the Cordilleran ice sheet. Still other ice from the ice field west of Marias Pass flowed east across the Continental Divide as a major tributary of the Two Medicine glacier.

East of the Continental Divide in the Glacier National Park region, some glaciers advanced beyond the mountain front onto the adjoining plains, where they formed large piedmont glaciers. The largest of these piedmont glaciers was the Two Medicine glacier (fig. 3). Glaciers along a 60-km length of the mountain front merged to form this glacier, including glaciers from Two Medicine valley and from the ice field west of Marias Pass. The Two Medicine glacier had a maximum width of about 50 km, extended about 55 km beyond the mountain front, and left a scattered veneer of hummocky till over an area of 2,000 km^2 (Calhoun, 1906; Alden, 1932).

To the north, other glaciers heading in Glacier National Park extended beyond the mountain front onto the plains. Glaciers in the valleys of Lake and Cutbank Creeks flowed east and merged on the plains into a small piedmont lobe that extended about 15 km beyond the mountain front (Calhoun, 1906; Alden, 1953). Farther north, a large glacier in the St. Mary River valley flowed beyond the mountain front and into Canada (fig. 3), where its terminal deposits lie beneath till of a subsequent advance of the Laurentide ice sheet (Calhoun, 1906; Alden, 1932; Horberg, 1954). The St. Mary glacier was fed by tributary glaciers flowing from the valleys of Divide, Red Eagle, Boulder, Swiftcurrent, Kennedy, and Otatso Creeks in Glacier National Park. At Lower St. Mary Lake this glacier was about 370 m thick (Alden, 1932). In the interior of the park, in the valleys of the Belly and Waterton Rivers, large glaciers about 400 m thick flowed north into Canada.

VOLCANIC ASHES IN THE GLACIER NATIONAL PARK REGION

The Mazama, Glacier Peak G, and Mount St. Helens Jy volcanic ashes (from youngest to oldest) have been found in the Glacier National Park region. These late Quaternary ashes were identified on the basis of (1) stratigraphic position, (2) radiocarbon dating, (3) mafic mineral assemblage, and (4) various properties of their glass fraction including shard morphology, refractive index, and major element values. Because these ashes were used to establish a chronology for late Quaternary events in the Glacier National Park region, a brief summary discussion of each ash follows.

Mazama Ash

The Mazama ash is the most widespread late Quaternary ash in the western United States and southwestern Canada. It has been found in Oregon, Washington, Idaho, northern California, northern Nevada, northern Utah, western Montana, western Wyoming, southern British Columbia, and southern Alberta (fig. 4). After some initial confusion this ash was identified as having originated from the great eruption of Mount Mazama (present site of Crater Lake, south-central Oregon) (Powers and Wilcox, 1964), a former volcano about 900 km southwest of Glacier National Park. This volcano was a complex assemblage of overlapping shield and stratovolcanic deposits whose climatic eruption about 6,845±50 yr. BP produced the Mazama ash (Bacon, 1983). This ash has been identified in about 25 bogs or exposures in the Glacier National Park region (Carrara, 1989).

Glacier Peak G Ash

Glacier Peak is a stratovolcano in north-central Washington, about 550 km west of Glacier National Park (fig. 4). Powers and Wilcox (1964) identified a widespread late Pleistocene ash as originating from this volcano, and subsequent work has shown that Glacier Peak has erupted many times since the late Pleistocene (Porter, 1978; Beget, 1982a, 1982b). Porter (1978) identified at least nine ashes erupted from Glacier Peak since the late Pleistocene. However, only two ashes, designated B (youngest) and G, have been identified as far east as Montana; of these only the G ash was found in the Glacier National Park region. These two ashes were deposited from distinct plumes by closely spaced eruptions about 11,200 yr. BP (Mehringer and others, 1984). Layer B spread southeast from Glacier Peak (fig. 4) and has been reported as far east as Lost Trail Pass bog in the Lemhi Range on the Idaho-Montana border (Mehringer, Arno, and Petersen, 1977; Mehringer, Blinman, and Petersen, 1977; Mehringer and others, 1984), and Cub Creek Pond in Yellowstone National Park (Waddington and Wright, 1974; Westgate and Evans, 1978).

Layer G spread east from Glacier Peak and has been reported as far as southern Alberta (Westgate and Evans, 1978). Layer G has also been found in the Kootenai River area, Montana (Mierendorf, 1984); Sheep Mountain bog, about 18 km northeast of Missoula, Mont. (Mehringer and others, 1984); the Sun River Canyon area, Montana (Lemke and others, 1975; Carrara and others, 1986); and the Glacier National Park region (Carrara, 1986; Carrara and others, 1986). In the Glacier National Park region the Glacier Peak G ash has now been identified at 10 sites (Carrara, 1989).

Mount St. Helens Jy Ash

Mount St. Helens, a stratovolcano in southern Washington about 700 km west-southwest of Glacier National Park (fig. 4), erupted many times during the late Quaternary (Mullineaux and Crandell, 1981; Mullineaux, 1986). Most ashes produced by these eruptions are of only local extent, but several have been found far to the east.

Ash layers of the Mount St. Helens set J were the product of explosive eruptions between about 11,500 and 10,800 yr. BP (D.R. Mullineaux, written commun.,1986). East of Mount St. Helens, three ash layers have been recognized in set J. From youngest to oldest these ash layers are the Jb, Jy, and Js. In addition, west of Mount St. Helens, another layer, Jg, thought to be the youngest set J ash, is present (Mullineaux, 1986). The Jy ash consists of at least two voluminous, overlapping beds without known evidence of a significant time lapse between them (Mullineaux, 1986). This ash, which has been found in the Glacier National Park region stratigraphically below the Glacier Peak G ash, was estimated by Carrara and others (1986) to date from about 11,400 yr. BP.

Both layers Jb and Jy are extensive enough to be useful marker beds far from Mount St. Helens. The Jb ash spread southeast as a narrow plume, whereas the Jy ash covered a broader area from northeast to southwest of Mount St. Helens (fig. 4) (D.R. Mullineaux, oral commun., 1986). Set J ashes have been reported previously from two localities in Montana. An ash of set J (probably Jy, based on the K:Fe ratio) has been reported in the Kootenai River valley, about 150 km west of Glacier National Park (Mierendorf, 1984). In addition, the Jy ash has been reported near Marias Pass immediately south of Glacier National Park (Carrara and others, 1986). In the Glacier National Park region the Mount St. Helens Jy ash has been identified at three sites (Carrara, 1989), including the Marias Pass site. At these three sites it is stratigraphically below the Glacier Peak G ash.

Although the Glacier Peak G and Mount St. Helens Jy ashes were not identified in many of the bogs inspected in this study and were not found in soils in the high areas of the park, their absence cannot be taken as conclusive evidence that the underlying deposits postdate these ashes. Where found, the Glacier Peak G and Mount St. Helens Jy ashes are usually less than 2 cm thick. As the majority of these ash localities are in bogs, the ashes were recovered by augering several meters below the surface. Thus, it is possible that in some bogs these ashes were present but not identified. In addition, the likelihood of preservation of these ashes on a deposit in the higher areas of the park would be minimal in a sparsely vegetated, recently deglaciated landscape. In support of the idea that ash preservation would be minimal in a recently deglaciated terrain, Jackson and others (1982) found that the early postglacial period was a time of active geomorphic processes not conducive to

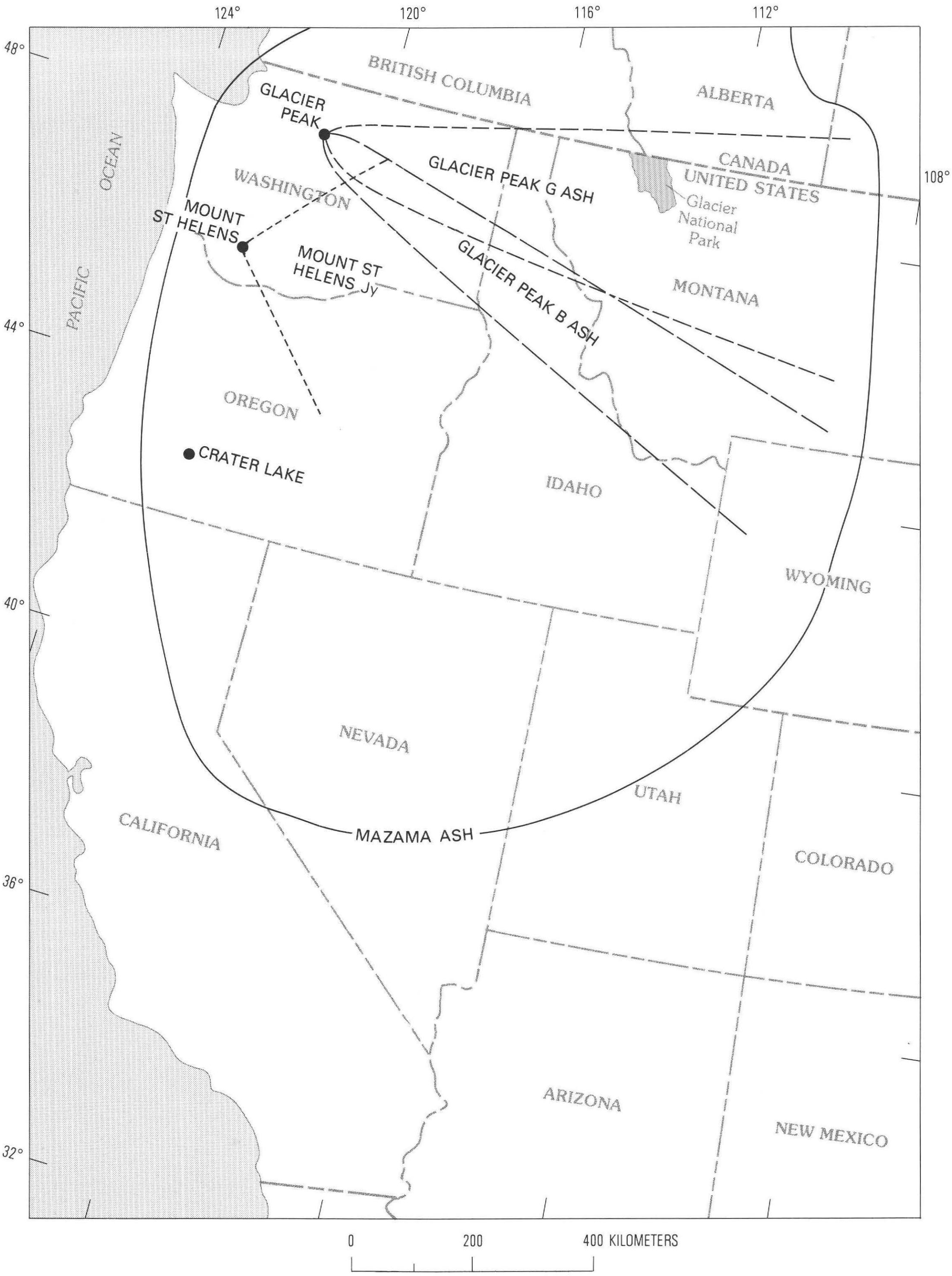


Figure 4. Distribution pattern of the Mazama, Glacier Peak B and G, and Mount St. Helens Jy ashes (modified from Porter, 1978; Westgate and Gorton, 1981; Sarna-Wojcicki and others, 1983; and Mullineaux, 1986).

ash preservation. Hence, the absence of the Glacier Peak G or Mount St. Helens Jy ashes in a bog, in an exposure, or in a soil does not prove that a given deposit postdates these ashes.

Identification of Ashes in the Glacier National Park Region

In the field the middle Holocene Mazama ash may tentatively be distinguished from the late Pleistocene Glacier Peak G and Mount St. Helens Jy ashes by stratigraphic position and the amount of mafic crystals.

Stratigraphically, in bogs or exposures, the Mazama ash is usually within 0.5 to 2 m of the surface in the highly organic, dark-brown (10 *YR* 3/3) to black (10 *YR* 2/1),[1] fine-grained sediment characteristic of the middle Holocene in the Glacier National Park region. The ash consists of a silty layer ranging in thickness from 3 cm to as much as 100 cm. Grain-size analyses of several samples of this ash yielded an average sand:silt:clay ratio of 13:63:24. Where thick and containing little organic matter, the ash is light gray (10 *YR* 7/1); where thin and contaminated with detrital sediment and organic matter, the ash is pale brown (10 *YR* 6/3) to dark yellowish brown (10 *YR* 3/6). However, because of its gritty feel, caused by its sand-sized fraction, and because of its lighter color, which contrasts with the darker surrounding organic sediment, the Mazama ash is easily recognized even when highly contaminated.

Stratigraphically, in bogs or exposures, the Glacier Peak G and Mount St. Helens Jy ashes are usually within the lowermost organic sediments, or in the underlying inorganic sediments deposited soon after deglaciation. The Glacier Peak G ash ranges in thickness from 3 mm to 3 cm, whereas the Mount St. Helens Jy ash is from 1 mm to 1 cm thick. Grain-size analyses of several samples of the Glacier Peak G ash yielded an average sand:silt: clay ratio of 18:57:25. Grain-size analysis of the Mount St. Helens Jy ash from the Marias Pass exposure (Carrara and others, 1986) yielded a sand:silt:clay ratio of 9:69:22. Both ashes are white (10*YR* 8/1) to light gray (10*YR* 7/1) in color.

The Mazama ash has also been identified in soil in many places in Glacier National Park, and is common in areas above treeline (Osborn, 1985; Carrara, 1987). In these places, the ash is a light-yellowish-brown (10 *YR* 6/4) to reddish-brown (2.5 *YR* 5/6) (due to mafic mineral oxidation) silt, 5–10 cm thick in small depressions at a depth of 10–20 cm. Grain-size analyses of two samples yielded an average sand:silt:clay ratio of 8:67:25.

Mafic crystals, easily seen with a hand lens, are relatively abundant in the Glacier Peak G ash compared with the Mazama and Mount St. Helens Jy ashes (Powers and Wilcox, 1964; Lemke and others, 1975). In the Glacier National Park region, mafic crystals compose about 15–20 percent by volume of the Glacier Peak G ash. In contrast, mafic crystals make up only about 5 percent of the Mazama and Mount St. Helens Jy ashes.

Microscopically, the Mazama ash is also easily distinguished from the Glacier Peak G and Mount St. Helens Jy ashes. Although all three ashes contain crystals of plagioclase, hornblende, hypersthene, and opaque oxides, the Mazama ash also contains crystals of augite and apatite (Lemke and others, 1975, table 1; and Sarna-Wojcicki and others, 1983, table 5–2). Numerous elongate frothy glass shards containing tubular voids separated by thin glass walls are characteristic of Mazama ash (fig. 5*A*). In addition, the refractive index of the glass is >1.502, higher than any other Holocene or late Pleistocene ash reported in Montana (Wilcox, 1965, fig. 2). Glass shards within the Glacier Peak G ash contain small crystals, are vesicular, and relatively equidimensional in comparison to shards in the Mazama ash (fig. 5*B*). Glass shards within the Mount St. Helens Jy ash are similar in appearance to those of the Glacier Peak G ash but are relatively free of crystals (fig. 5*C*).

Chemically, glass from the Mazama ash can also be distinguished from that of other common Holocene and late Pleistocene ashes, including the Glacier Peak G and Mount St. Helens Jy ashes. Mazama glass has less silica and more iron and titanium than these other ashes. For instance, Mazama glass commonly is about 73 percent SiO_2, 2 percent FeO, and 0.4 percent TiO_2, whereas common values for glass within the Glacier Peak B, G, and Mount St. Helens Jy ashes are 76–77 percent, 76–77 percent, 75–76 percent SiO_2; 1.2 percent, 1 percent, 1.3 percent FeO; and 0.2 percent, 0.2 percent, 0.2 percent TiO_2 respectively (Lemke and others, 1975; Westgate and Evans, 1978; Westgate and Gorton, 1981; Sarna-Wojcicki and others, 1983; Mehringer and others, 1984). The higher iron content of glass within the Mazama ash compared to Glacier Peak B and G, and Mount St. Helens Jy ashes is also shown by K:Fe ratios. In this study, K:Fe ratios determined by wavelength dispersive X-ray fluorescence for 11 samples of Mazama ash from bogs in the Glacier National Park region ranged from 1.37:1 to 1.51:1 and averaged 1.46:1 (table 2). K:Fe ratios also determined by X-ray fluorescence for five samples of Mazama ash from soils in the high areas of Glacier National Park ranged from 1.46:1 to 1.51:1 and averaged 1.49:1 (table 3). As will become apparent in the following discussion, these K:Fe values for the Mazama ash are significantly lower than K:Fe values of the Glacier Peak B, and G, and Mount St. Helens Jy ashes.

[1]All colors given in this report are for moist or wet sediment.

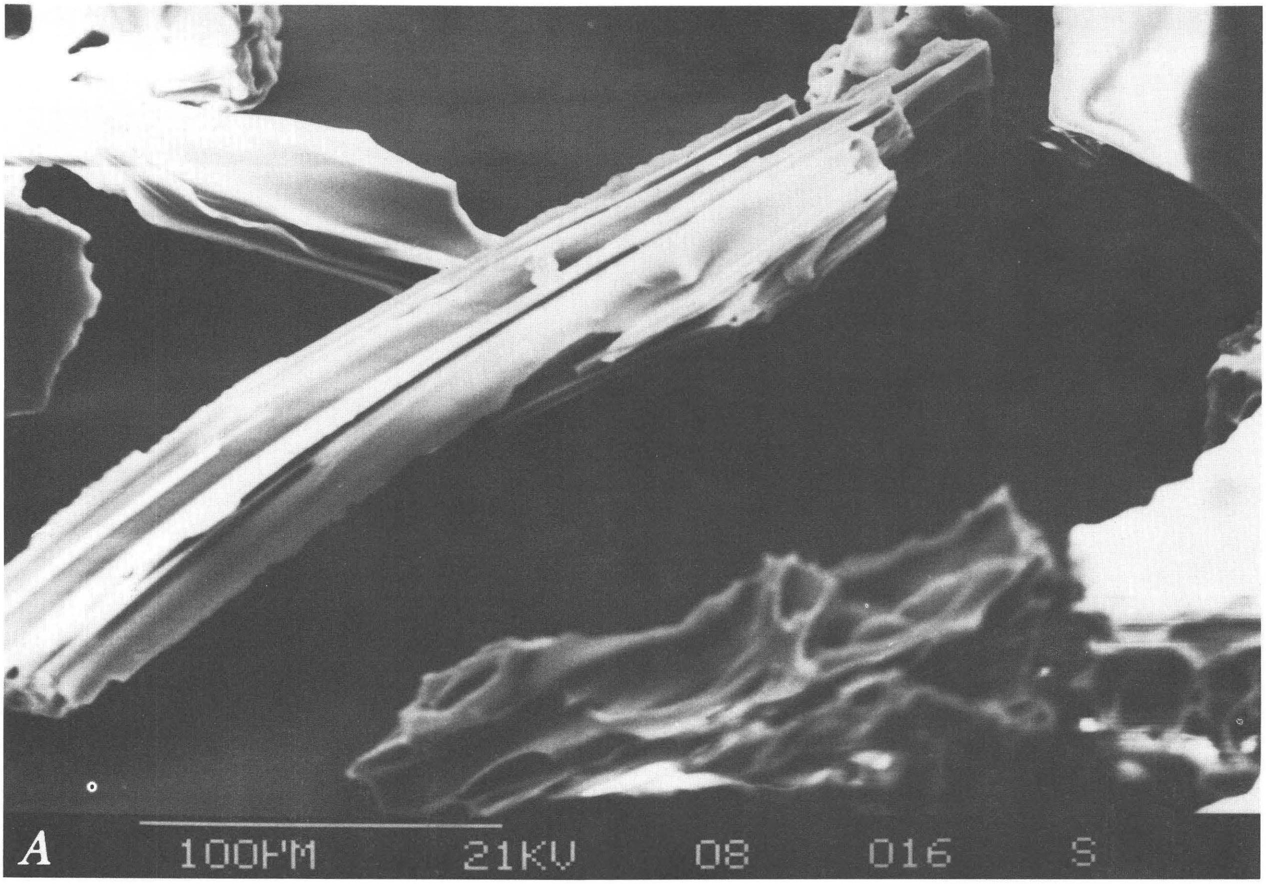


Figure 5 (above and following pages). Scanning electron micrographs of glass shards. *A*, Mazama ash.

Although the Glacier Peak B ash was not found in the Glacier National Park region, at the start of this study its presence could not be ruled out. Hence, its mineral and chemical characteristics are discussed and contrasted with the Glacier Peak G and Mount St. Helens Jy ashes that are present in the region.

The Glacier Peak B and G and Mount St. Helens Jy ashes all contain crystals of hornblende, hypersthene, plagioclase, and opaque oxides. The refractive index of the glass shards of all these ashes is about 1.498. Hence, these ashes cannot be distinguished from one another under a microscope. Because these ashes are very close in age, they could be considered as a single time-stratigraphic marker in order to establish a chronology for upper Quaternary deposits in the northwestern United States and southwestern Canada. However, their specific identifications are important in gaining a better understanding of their distribution and age relationships.

Fortunately, these ashes can be distinguished by chemical analysis of their glass fractions. The Glacier Peak B ash contains more calcium and iron and less potassium than the G ash (Westgate and Evans, 1978). Glass within the Glacier Peak B ash commonly contains about 1.3–1.4 percent CaO, 1.2 percent FeO, and 3 percent K_2O, whereas glass within the Glacier Peak G ash contains about 1.2–1.3 percent CaO, 1 percent FeO, and 3.5 percent K_2O (Lemke and others, 1975; Westgate and Evans, 1978; Sarna-Wojcicki and others, 1983; Mehringer and others, 1984). These seemingly small chemical differences are readily apparent in the K:Fe ratios. Smith and others (1977) found that K:Fe ratios of 7 samples of B ash ranged from 2.7:1 to 2.9:1 and averaged 2.8:1, and that K:Fe ratios of 10 samples of G ash ranged from 3.1:1 to 3.7:1 and averaged 3.4:1.

In this study, K:Fe ratios, determined by X-ray fluorescence for nine samples of Glacier Peak G ash from the Glacier National Park region, ranged from 3.05:1 to 3.41:1 and averaged 3.23:1 (table 4). The ratio for the Guardipee Lake sample determined by electron microprobe was 3.69:1.

The Mount St. Helens J ashes contain more iron (except the Js layer), aluminum and sodium, and less

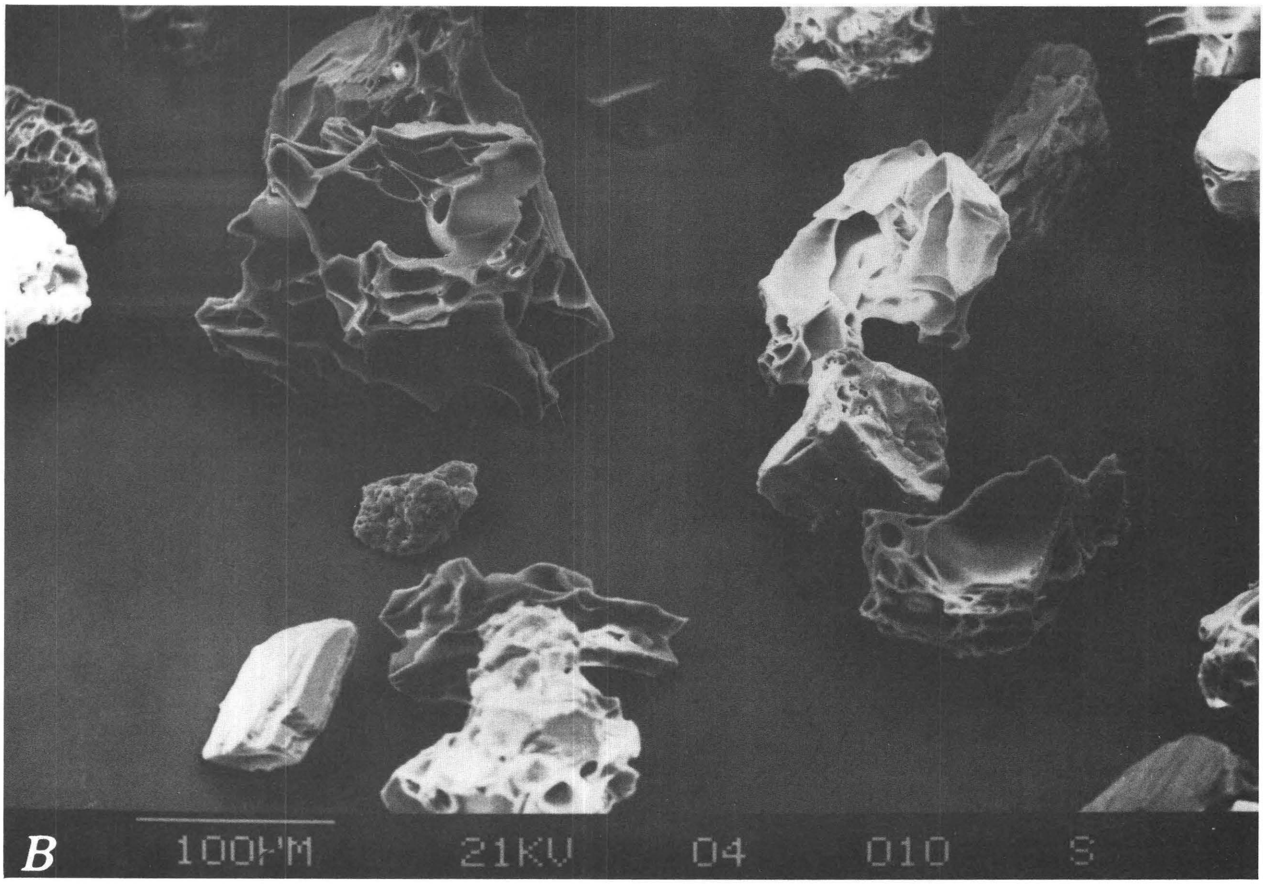


Figure 5—Continued. Glass shards in *B*, Glacier Peak G ash.

potassium, than the Glacier Peak B and G ashes (Westgate and Evans, 1978). The two widespread ashes of set J, Jb and Jy, can be distinguished from one another in that Jb contains more iron, magnesium, and calcium than does the Jy ash. Glass in the Mount St. Helens Jb ash is about 2 percent FeO, 0.75 percent MgO, and 2 percent CaO, whereas glass in the Jy ash is about 1.35 percent FeO, 0.35 percent MgO, and 1.35 percent CaO (Westgate and Evans, 1978, table 3; D.R. Mullineaux, oral commun., 1984). Because of their different iron contents, these two ash layers can also be distinguished by their K:Fe ratios. Samples UA 482 and UA 483 (Westgate and Evans, 1978, table 3) are from layers Jb and Jy respectively (D.R. Mullineaux, oral commun., 1984). Layer Jb has a K:Fe ratio of 1.23:1; layer Jy has a ratio of 1.96:1. A sample of set J ash, probably Jy (D.R. Mullineaux, oral commun., 1984) analyzed by Smith and others (1977) yielded a K:Fe ratio of 1.9:1.

In this study, K:Fe ratios determined by X-ray fluorescence for the two samples of Mount St. Helens Jy ash from the Glacier National Park region were both 1.91:1 (table 5). The ratio for the Guardipee Lake sample determined by electron microprobe was 2.14:1.

RADIOCARBON AGES AND VOLCANIC ASHES FROM BOGS AND EXPOSURES IN THE GLACIER NATIONAL PARK REGION

In this report, the bogs and exposures from which radiocarbon ages were obtained and volcanic ashes were identified have been placed into five groups based upon their geographical location or related glacial history. Most of these sites are on the west side of Glacier National Park because bogs are abundant in this area, especially in the valley of the North Fork Flathead River and in the Lake McDonald area, whereas they are sparse in the Livingston and Lewis Ranges and the eastern areas of the park. Not all bogs and exposures inspected in this study are described in this report; only those that yielded information pertaining to the late Quaternary history of the Glacier National Park region are discussed. Areas containing important bogs and (or) exposures are (1) Bowman Lake, three sites; (2) Lake McDonald, six sites; (3) Marias Pass, four sites; (4) Lower St. Mary Lake, one site; and (5) Kootenai Lakes, two sites. Stratigraphy at these sites is described from youngest to oldest.

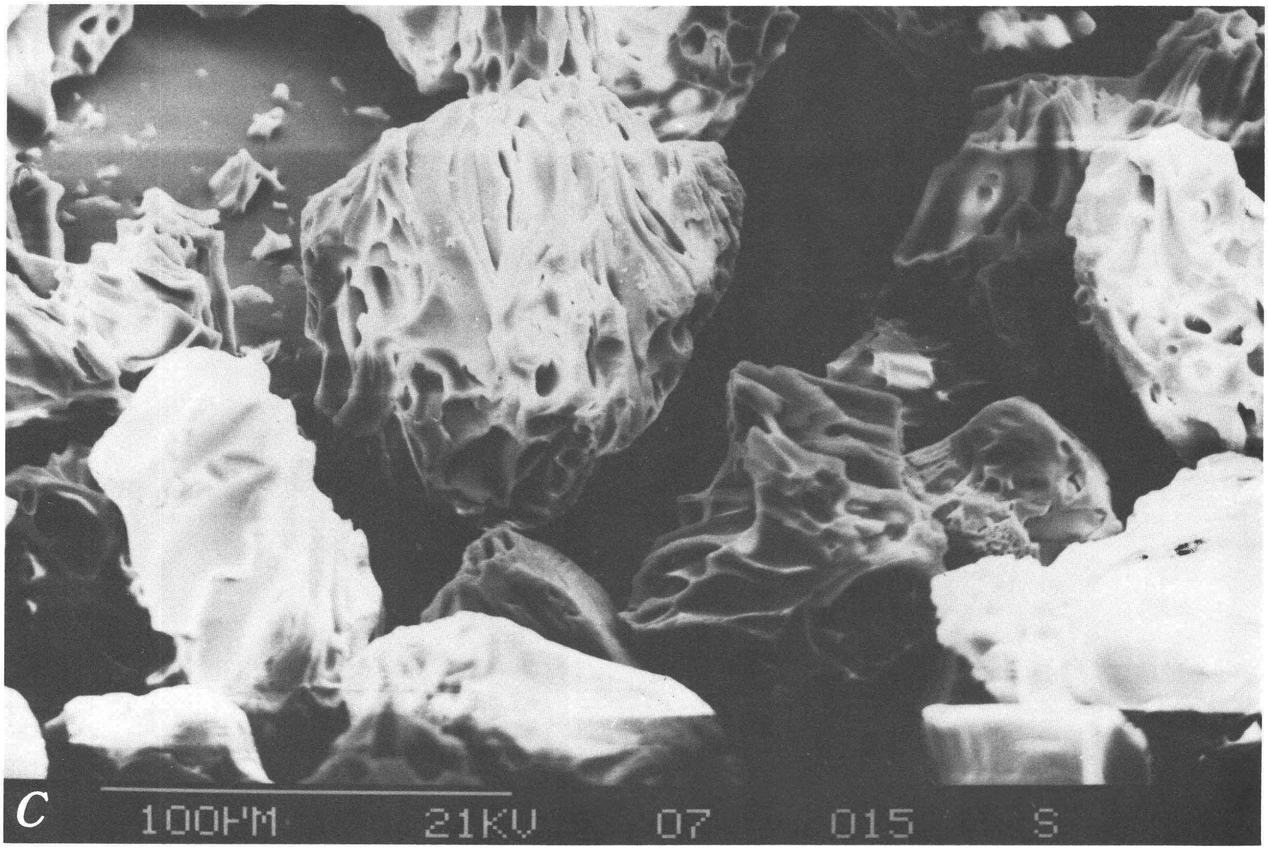


Figure 5—Continued. Glass shards in *C*, Mount St. Helens Jy ash.

Bowman Lake Area

Bowman Lake is in the northwestern corner of Glacier National Park (fig. 1). It is a deep lake, 10.4 km long, occupying a glacial valley and dammed by a hummocky end moraine of late Wisconsin age deposited by the former Bowman glacier (Carrara, 1989). The moraine contains several bogs. During late Wisconsin time, a large glacier that headed along the Continental Divide on the west side of the Livingston Range flowed down the Bowman valley for about 20 km and joined the large trunk glacier occupying the valley of the North Fork Flathead River. This trunk glacier was about 120 km long and 1,000 m thick. Three sites were investigated in the Bowman Lake area: (1) Cabin Creek exposure, (2) Bowman Lake bog, and (3) Numa Ridge bog (fig. 6).

Cabin Creek Exposure

The Cabin Creek exposure is in the valley of the North Fork Flathead River at an altitude of 1,122 m, about 9 km north of the Polebridge Ranger Station (fig. 6). The exposure is about 0.3 km east of the river and about 20 m above it in a former bog along the east side of Glacier Route 7. The former bog is drained and incised by a small north-flowing stream, which has exposed stratigraphic sections along its banks. The area is underlain by late Wisconsin till deposited by the trunk glacier that filled the valley of the North Fork Flathead River. This till in places exceeds 30 m in thickness, and it overlies the Kishenehn Formation. Forests in this area consist primarily of ponderosa and lodgepole pine, Engelmann spruce, and subalpine fir.

The Cabin Creek exposure, which is almost 4 m high, lies in a stream cut at the north end of the former bog (fig. 7). In addition, the stratigraphy was extended another 2.3 m by auger. Although the stratigraphic section at this exposure does not extend from the late Pleistocene to the most recent sediments, younger sediments including the Mazama ash were found in the stream bank farther upstream (south). Hence, a complete stratigraphic section spanning the last 12,000 years is represented by the various sections exposed along the stream bank.

At the Cabin Creek exposure the Glacier Peak G and Mount St. Helens Jy ashes were identified in a massive, inorganic, dark-grayish-brown (2.5 *Y* 4/2), clayey-silt lake sediment (tables 4 and 5; fig. 8). This lake sediment has a sand:silt:clay ratio of 1:56:43. The Glacier Peak G ash is a white (10 *YR* 8/1) silty layer 3–6 mm thick. The Mount St. Helens Jy ash, about 25 cm below the

Table 2. Major-element values of glass in the Mazama ash from bogs in the Glacier National Park region[1]

Site--	Dry[2]	Bowman Lake	Mud Lake[3]	McGee Meadow	Glacier Route 7	Ben Ryan	Snyder Ridge	Marias Pass[4]	Middle Fork	Chewing Blackbones	Mirror Pond[5]	Average
SiO_2	72.8	72.8	72.5	72.9	71.9	72.4	73.1	73.6	72.6	73.0	72.9	72.8
Al_2O_3	14.6	14.6	14.7	14.8	15.7	14.8	14.5	13.9	14.7	14.5	14.5	14.6
[6]FeO	1.97	2.01	2.10	2.02	1.92	1.99	1.99	2.15	1.98	2.01	2.01	2.01
MgO	.54	.53	.61	.58	.56	.51	.56	.53	.52	.52	.52	.54
CaO	1.83	1.75	1.94	1.83	1.91	2.07	1.77	1.69	1.94	1.74	1.80	1.84
Na_2O	5.00	5.05	5.04	4.78	4.90	5.08	4.91	4.69	5.15	5.07	5.09	4.99
K_2O	2.79	2.81	2.70	2.68	2.69	2.68	2.72	2.82	2.71	2.79	2.78	2.75
TiO_2	.41	.42	.45	.43	.44	.43	.45	.40	.44	.43	.41	.43
Total---	99.94	99.97	100.07	100.02	100.02	99.96	100.00	100.05	100.04	100.06	100.01	99.96
K:Fe	1.51:1	1.49:1	1.37:1	1.42:1	1.49:1	1.43:1	1.46:1	1.40:1	1.46:1	1.49:1	1.48:1	1.46:1

[1]Weight percent, without water, recalculated to 100 percent. Analysis by A.J. Bartel, U.S. Geological Survey.

[2]Located 150 m south of the Cabin Gulch exposure (see fig. 6).

[3]Located 5 km south-southeast of Polebridge (see fig. 1).

[4]Located 0.5 km southwest of Marias Pass.

[5]Located in upper St. Mary River drainage area (see fig. 18).

[6]Total iron as FeO.

Table 3. Major-element values of glass in the Mazama ash from soils in the Glacier National Park region[1]

Site--	Sperry Glacier	Comeau Pass[2]	Red Eagle Glacier	Grinnell Glacier	Clements Glacier	Average
SiO_2	72.6	72.0	73.7	72.5	72.7	72.7
Al_2O_3	14.7	15.1	13.9	14.8	14.7	14.6
[3]FeO	1.94	1.86	2.06	1.98	1.93	1.95
MgO	.50	.44	.46	.51	.50	.48
CaO	1.95	2.31	1.49	1.81	1.89	1.89
Na_2O	5.16	5.26	5.09	5.19	5.13	5.17
K_2O	2.71	2.54	2.91	2.74	2.73	2.73
TiO_2	.41	.41	.39	.43	.41	.41
Total--	99.97	99.92	100.00	99.96	99.99	99.93
K:Fe	1.49:1	1.46:1	1.51:1	147:1	1.51:1	1.49:1

[1]Weight percent, without water, recalculated to 100 percent. Analysis by A.J. Bartel, U.S. Geological Survey.

[2]Located 0.8 km west of Sperry Glacier (see fig. 30).

[3]Total iron as FeO.

Glacier Peak G ash (fig. 7), is a white (10 *YR* 8/1) silty layer, 1 mm thick. The Mount St. Helens Jy ash is underlain by 3.4 m of fine-grained sediment that overlies till or outwash.

Bowman Lake Bog

The Bowman Lake bog is at an altitude of 1,238 m, about 0.6 km southwest of the outlet of Bowman Lake (fig. 6). Forest surrounding this bog consists mostly of lodgepole pine with some Engelmann spruce and subalpine fir. The bog is on the hummocky end moraine of late Wisconsin age (Carrara, 1989) deposited by the former Bowman glacier after it separated from the large trunk glacier in the valley of the North Fork Flathead River and retreated upvalley.

At the Bowman Lake bog a hole was augered 280 cm into the underlying sediments (fig. 7). The Mazama ash (table 2) forms as a dark-yellowish-brown (10 *YR* 3/6) layer 10 cm thick at a depth from 140 to 150 cm. Grain-size analysis of the ash from this bog gave a sand:silt:clay ratio of 12:66:22. Below the Mazama ash, at a depth from 260 to 275 cm, near the base of the hole, a sample of gray (5 *Y* 6/1) clay was obtained. Although this sample had a low organic content, its organic matter was concentrated in the laboratory (Kihl, 1975) and yielded a radiocarbon age of 11,150±90 yr. BP (W–5043) (table 6).

Numa Ridge Bog

Numa Ridge bog, at an altitude of 1,535 m, is on Numa Ridge about 1 km northwest of Bowman Lake (fig. 6). The bog borders a small lake (fig. 9) and is surrounded by forest containing lodgepole pine, Engelmann spruce, and subalpine fir. The bog is underlain by late Wisconsin till deposited by the large trunk glacier that filled the valley of the North Fork Flathead River.

At Numa Ridge bog a hole was augered 430 cm into the underlying sediments (fig. 7). Here the Mazama ash is a light-gray (10 *YR* 7/1) layer 15 cm thick at a depth from 275 to 290 cm. Grain-size analysis of this ash gave a sand:silt:clay ratio of 11:62:27. The ash is in a dark-

Table 4. Major-element values of glass in the Glacier Peak G ash from bogs and exposures in the Glacier National Park region[1]

Site--	Cabin Gulch	Glacier Route 7	Howe Lake	Ben Ryan	Snyder Ridge	Marias Pass	Middle Fork	Goat Lick	Chewing Blackbones	Guardipee Lake[2]	Average[3]
SiO_2	74.9	76.2	75.6	73.9	75.9	76.5	75.7	75.4	76.6	78.5	75.6
Al_2O_3	14.1	13.3	13.5	14.6	13.4	13.3	13.7	14.0	12.6	12.0	13.6
[4]FeO	1.08	1.05	1.06	1.07	1.08	1.06	1.06	1.06	1.05	.95	1.06
MgO	.47	.39	.45	.39	.45	.40	.40	.42	.42	.23	.42
CaO	2.10	2.10	2.17	2.41	1.87	1.53	1.90	1.91	1.90	1.21	1.99
Na_2O	4.05	3.70	3.75	4.17	3.96	3.63	3.88	3.87	3.97	3.58	3.89
K_2O	3.09	3.11	3.30	3.28	3.16	3.39	3.19	3.22	3.26	3.29	3.22
TiO_2	.17	.20	.19	.17	.19	.19	.18	.18	.17	.21	.18
Total--	99.96	100.05	100.02	99.99	100.01	100.00	100.01	100.06	99.97	99.97	99.96
K:Fe	3.05:1	3.16:1	3.32:1	3.27:1	3.12:1	3.41:1	3.21:1	3.24:1	3.31:1	3.69:1	3.23:1

[1]Weight percent, without water, recalculated to 100 percent. Analysis by A.J. Bartel, U.S. Geological Survey.
[2]Analysis by A.M. Sarna-Wojcicki, U.S. Geological Survey, using electron microprobe (written commun., 1985).
[3]Includes only those samples (9) collected by author and analyzed by XRF.
[4]Total iron as FeO.

Table 5. Major-element values of glass in the Mount St. Helens Jy ash from the Glacier National Park region[1]

Site--	Cabin Gulch	Marias Pass	Guardipee Lake[2]
SiO_2	75.6	75.9	78.4
Al_2O_3	13.7	13.8	12.3
[3]FeO	1.29	1.32	1.12
MgO	.40	.41	.26
CaO	1.88	1.68	1.40
Na_2O	4.56	4.40	4.15
K_2O	2.31	2.36	2.25
TiO_2	.18	.20	.18
Total--	99.92	100.07	100.06
K:Fe	1.91:1	1.91:1	2.14:1

[1]Weight percent, without water, recalculated to 100 percent. Analysis by A.J. Bartel, U.S. Geological Survey.

[2]Analysis by A.M. Sarna-Wojcicki, U.S. Geological Survey, using electron microprobe (written commun., 1985).

[3]Total iron as FeO.

brown (10 *YR* 3/3) organic clay. Below the Mazama ash at a depth of 410 to 430 cm, near the base of the hole, an olive-gray (5 *Y* 4/2) gyttja was sampled. This material yielded a radiocarbon age of 10,090±130 yr. BP (W-5185) (table 6).

Discussion

The presence of the Glacier Peak G and Mount St. Helens Jy ashes at the Cabin Creek exposure indicates that this part of the valley of the North Fork Flathead River was deglaciated before 11,400 yr. BP. Hence, the large trunk glacier that filled this valley during late Wisconsin time had retreated substantially and lay somewhere to the north of the Cabin Creek exposure by 11,400 yr. BP. This date (the estimated age of the Mount St. Helens Jy ash, Carrara and others, 1986), which represents a minimum date of deglaciation, may be minimum by several hundred or thousand years because the Mount St. Helens Jy ash at this exposure is underlain by 340 cm of fine-grained sediment (fig. 7). Because the Glacier Peak G ash (11,200 yr. BP) and the Mount St. Helens Jy ash (11,400 yr. BP) are both present at this exposure, a rate of sedimentation can be calculated between these two ashes and then extrapolated down to the base of the massive lake sediment unit in which these two ashes occur. In this way, an age of about 13,700 yr. BP is obtained for the base of the lake sediment at a depth of 570 cm (fig. 7). This age is of course only an estimate, but it suggests that the age of the Mount St. Helens Jy ash may be a very minimum date of deglaciation for this section of the valley of the North Fork Flathead River.

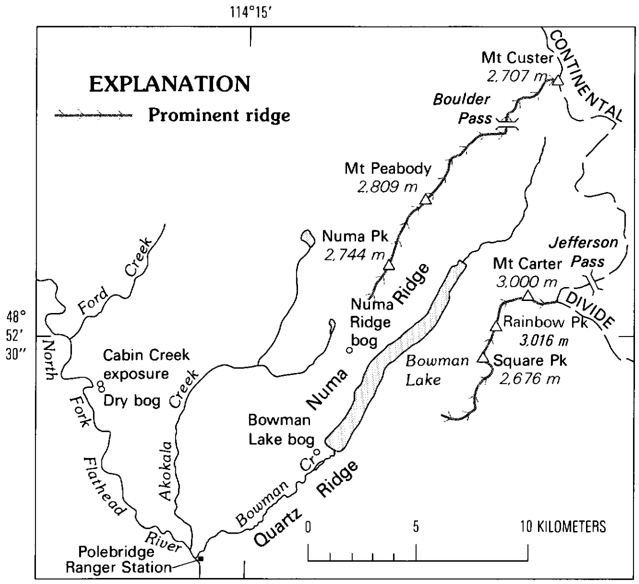


Figure 6. The Bowman Lake area.

The radiocarbon age of 11,150±90 yr. BP (W-5043) from the Bowman Lake bog is a minimum age for the hummocky end moraine on which this bog lies and is a minimum date for the withdrawal of ice from this section of Bowman Creek valley. This date indicates that by 11,150 yr. BP the Bowman glacier had shrunk drastically from its maximum late Wisconsin extent and was confined to its own valley in the Livingston Range, if it still existed at all.

Although the radiocarbon age of 10,090±130 yr. BP (W-5185) from the Numa Ridge bog is not as old as the radiocarbon age from Bowman Lake bog or the volcanic ashes from the Cabin Creek exposure, the date from Numa Ridge is important in that it provides a minimum date for revegetation following deglaciation at a site 300 m above Bowman Lake. Note that, as the bottom of the Numa Ridge bog was not reached by auger, the radiocarbon age from this bog is not thought to represent a close limiting date.

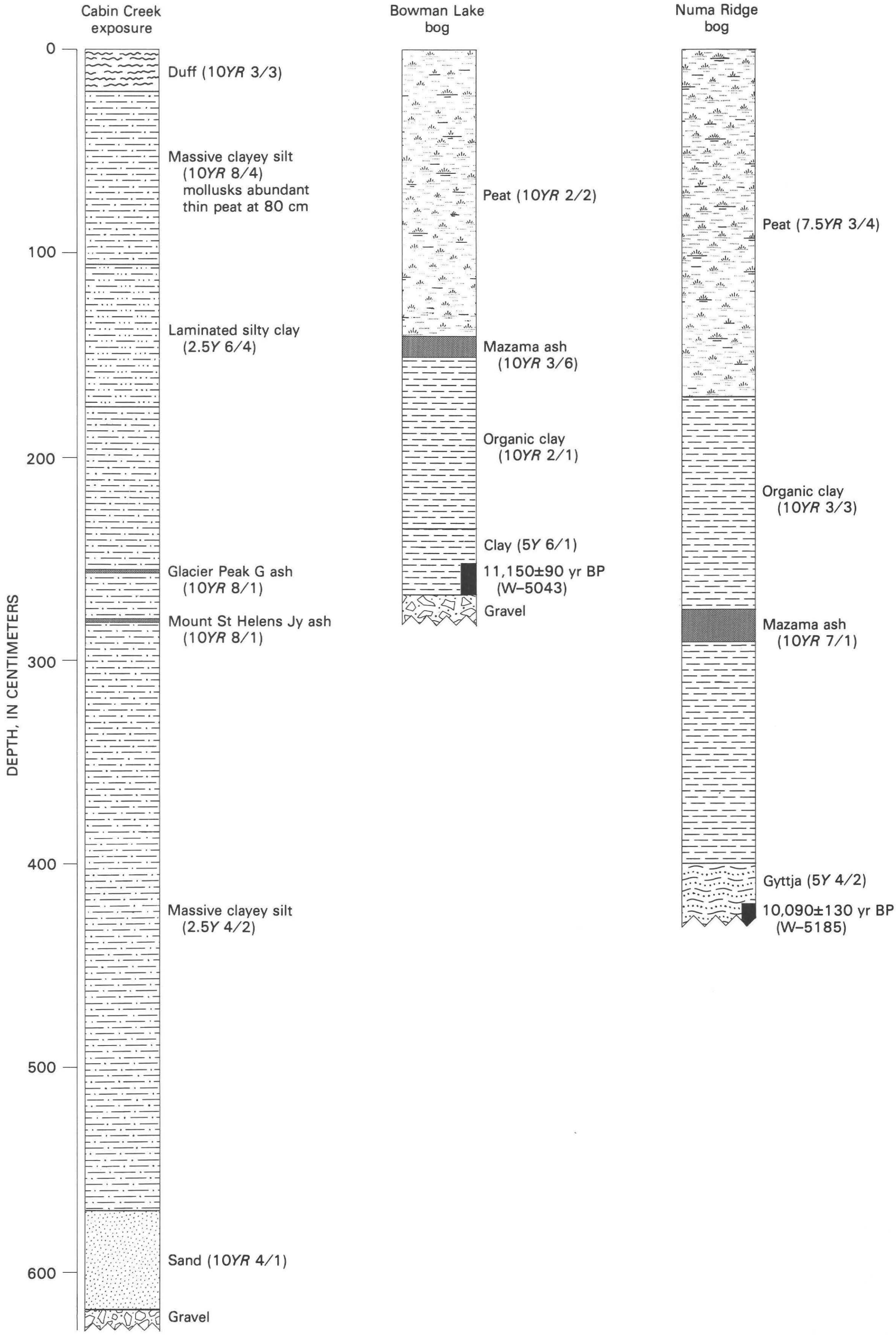


Figure 7. Stratigraphy of the three bogs investigated in the Bowman Lake area.

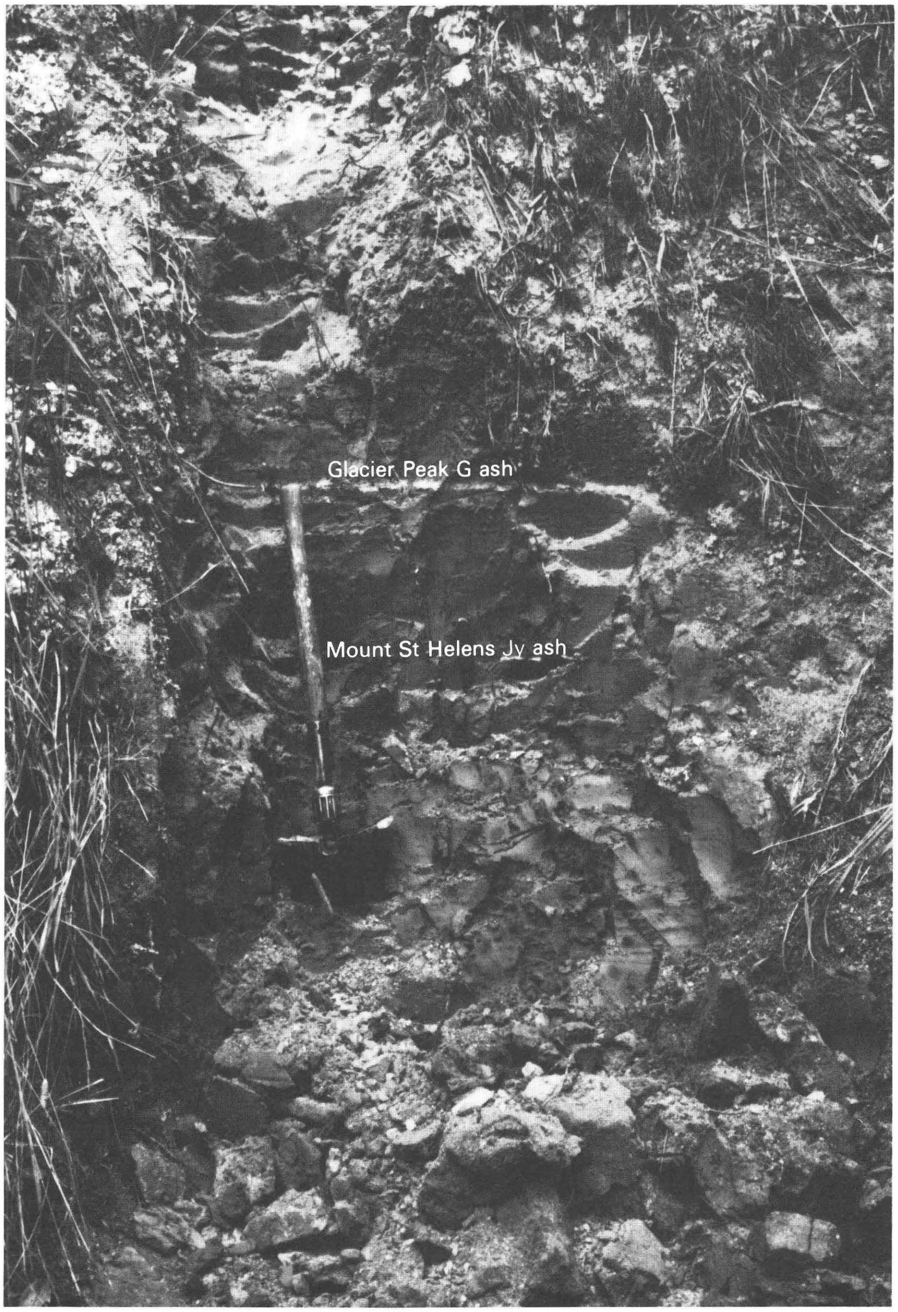


Figure 8. The Cabin Creek exposure, showing the Glacier Peak G and Mount St. Helens Jy ashes; shovel handle is 50 cm long.

Lake McDonald Area

Lake McDonald is in the west-central part of Glacier National Park. It is a deep lake, 15 km long, occupying the lower end of McDonald Valley (figs. 1 and 10) and dammed by till and outwash. During late Wisconsin time, a large glacier in the McDonald Valley headed in cirques along the western side of the Continental Divide in the Logan Pass and Flattop Mountain area. The McDonald glacier was tributary to the large trunk glaciers that filled the valleys of the North and Middle Forks Flathead River. To the south, Snyder Ridge was overrun by ice from the trunk glacier of the valley of the Middle Fork Flathead River. To the north, the McGee Meadow area and Howe Ridge were overrun by ice from the trunk glacier in the valley of the North Fork Flathead River. The Apgar Mountains, west of McGee Meadow, were also overridden by this glacier, indicating that in the vicinity of McGee Meadow this glacier was at least 800 m thick.

Table 6. Radiocarbon age determinations from the Glacier National Park region

^{14}C age	Lab. No.	Material	Site
780±70	W-5169	wood[1]	Halfmoon Lake area.
7,940±200	W-5572	*Salix*	Snyder Ridge bog.
8,730±80	W-5041	*Pinus*[2]	McGee bog.
9,510±350	W-5571	*Pinus*[2]	Mount Cleveland bog #1.
9,650±250	W-5575	gyttja, contained conifer fragments[2]	Snyder Ridge bog.
9,850±260	W-5577	gyttja, contained *Salix* fragments[3]	Mount Cleveland bog #4.
9,910±100	W-5191	*Pinus*[2]	East McGee bog.
9,920±120	W-5187	peat, contained *Pinus contorta* cone[2]	Glacier Route 7 bog.
10,090±130	W-5185	gyttja[3]	Numa Ridge bog.
10,630±250	W-5348	*Picea* or *Larix*[2]	Howe Lake bog.
11,150±90	W-5043	organic material (<0.125 mm)[3]	Bowman Lake bog.
>27,000	W-5166	organic material (<0.125 mm)[4]	Marias Pass exposure.
>40,000	W-5213	organic material (<0.125 mm)[4]	Marias Pass exposure.

[1]Wood at base of landslide exposed in section along Middle Fork Flathead River (see Carrara, 1989).

[2]Minimum date of reforestation following deglaciation.

[3]Minimum date of organic sedimentation following deglaciation.

[4]Samples associated with the Glacier Peak G and Mount St. Helens Jy ashes; samples contaminated by old carbon (see discussion in text).

Six sites were investigated in the Lake McDonald area: (1) McGee bog, (2) East McGee bog, (3) Glacier Route 7 bog, (4) Howe Lake bog, (5) Ben Ryan bog, and (6) Snyder Ridge bog (fig. 10).

McGee Bog

The McGee bog, at an altitude of 1,177 m, is about 12 km north-northwest of the town of West Glacier, in McGee Meadow near the headwaters of McGee Creek (fig. 10). Forest in the McGee Meadow area includes lodgepole pine, Engelmann spruce, subalpine fir, western larch, paper birch, and black cottonwood. The area is underlain by late Wisconsin till that in places exceeds 30 m in thickness (Carrara, 1989), and that overlies the Kishenehn Formation.

At the McGee bog a hole was augered 290 cm into the underlying sediments (fig. 11). The Mazama ash (table 2) was found near the surface, indicating a slow sedimentation rate since its deposition. The ash is a light-gray (10 *YR* 7/2) layer 60 cm thick. Grain-size analysis of the ash indicated a sand:silt:clay ratio of 16:61:23.

Several small wood fragments were recovered below the Mazama ash, at a depth between 130 and 140 cm at the base of a brown (10 *YR* 4/3), organic, silty clay (fig. 11). These fragments were identified as "yellow pine" (D.J. Christensen, written commun., 1981) and are probably lodgepole pine. These fragments yielded a radiocarbon age of 8,730±80 yr. BP (W–5041) (table 6).

Figure 9. Numa Ridge bog; view is to the east; Rainbow Peak (3,016 m) and Square Peak (2,676 m), left and right respectively, are in the background.

East McGee Bog

The East McGee bog, at an altitude of 1,180 m, is about 1 km southeast of the McGee bog (fig. 10). At the East McGee bog a hole was augered 520 cm into the underlying sediments (fig. 11). Here the Mazama ash is a light-brownish-gray (10 *YR* 6/2) layer, 35 cm thick at a depth between 135 and 170 cm. Grain-size analysis of this ash yielded a sand:silt:clay ratio of 14:63:23.

Several coniferous wood fragments were recovered below the Mazama ash at a depth between 500 and 520 cm, in a very dark brown (10 *YR* 2/2) gyttja. These fragments, also identified as "yellow pine" (D.J. Christensen, written commun., 1983), are probably lodgepole pine and yielded a radiocarbon age of 9,910±100 yr. BP (W–5191) (table 6).

Glacier Route 7 Bog

The Glacier Route 7 bog, at an altitude of 1,192 m, is about 100 m west of the Glacier Route 7 road, about 1 km south of the East McGee bog (fig. 10). At the Glacier Route 7 bog a hole was augered to a depth of 450 cm (fig. 11). Here the Mazama ash (table 2) is a dark-yellowish-brown (10 *YR* 4/4) layer 20 cm thick at a depth from 82 to 102 cm, overlying a black (5 *Y* 2.5/3) peat. Grain-size analysis of the ash yielded a sand:silt:clay ratio of 11:66:23.

Abundant fossil mollusks were found in a very pale brown (10 *YR* 7/3), organic (10 percent), silty-clay unit at a depth from 315 to 415 cm, containing thin lenses of peat. Two species of aquatic snails, *Gyraulus parvus* and *Planorbella trivolvis*, and two species of freshwater clams, *Pisidium compressum* and *Pisidium variable*, were identified (Emmett Evanoff, written commun., 1985). These species are very widespread today in North America, ranging from the central United States to the northern limit of boreal forest in Canada; hence, these mollusks provide no information concerning past climates. However, they do indicate that this bog was a perennial pond rich in aquatic vegetation at the time this unit was deposited (Emmett Evanoff, written commun., 1985).

A lodgepole pine cone was recovered at a depth of 400 cm in the mollusk-bearing unit. Unfortunately, the cone was too small for a conventional radiocarbon age

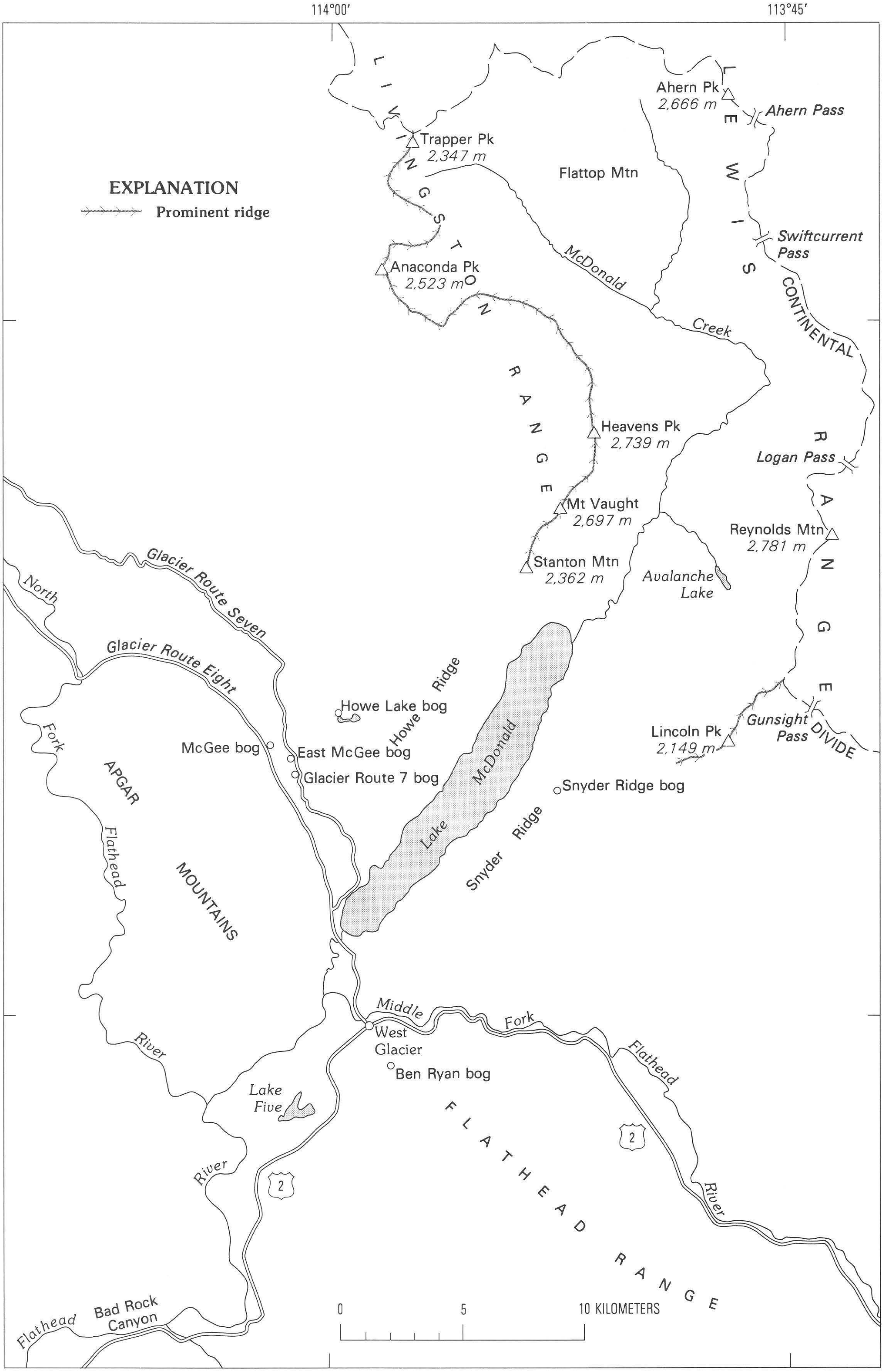


Figure 10. The Lake McDonald area.

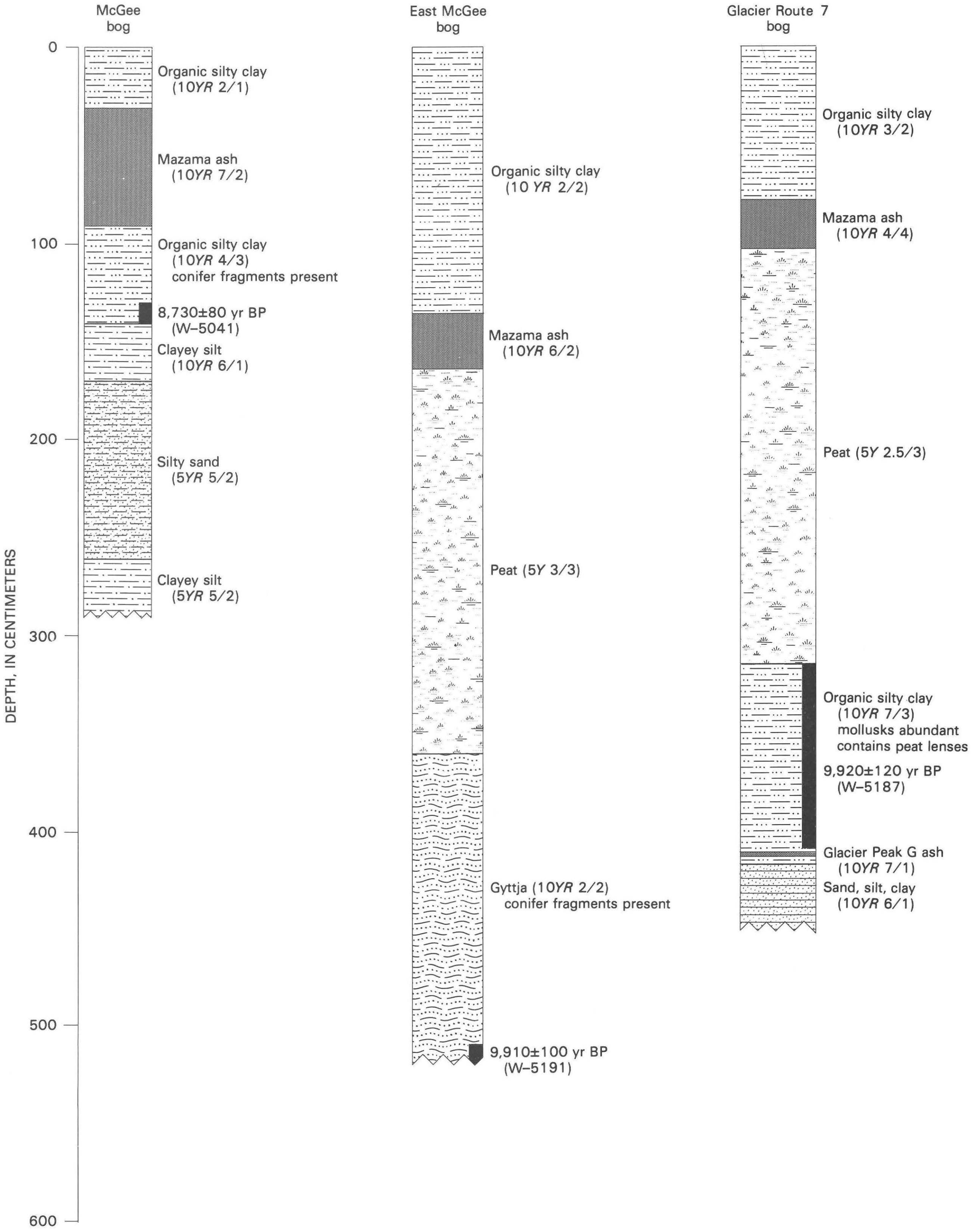


Figure 11. Stratigraphy of the three bogs investigated in the McGee Meadow area.

determination, so it was combined with the peat from the lenses in this unit and yielded a radiocarbon age of 9,920±120 yr. BP (W–5187) (table 6).

The Glacier Peak G ash (table 4) was found at a depth of 410 cm near the bottom of the mollusk-bearing unit. The ash forms a light-gray (10 *YR* 7/1) layer about 1 cm thick. Grain-size analysis of this ash indicated a sand:silt:clay ratio of 12:54:34. The mollusk-bearing unit overlies a light-gray (10 *YR* 6/1) pebbly sand, silt, and clay unit, that could only be penetrated about 35 cm by the auger.

Howe Lake Bog

The Howe Lake bog, at an altitude of 1,256 m, is on Howe Ridge north of Lake McDonald about 13 km north of the town of West Glacier (fig. 10). The area is underlain by late Wisconsin till that overlies the Kishenehn Formation. A nearby exposure in a landslide scarp indicates that the till is more than 30 m thick in this area. Forest surrounding this bog consists of lodgepole pine, Engelmann spruce, subalpine fir, and western larch.

The Howe Lake bog lies immediately northwest of Howe Lake. Here, a hole was augered to a depth of 355 cm (fig. 12). The upper 300 cm of sediment consisted of a black (10 *YR* 2/1) gyttja that contained abundant wood fragments. One fragment from a depth of 270 cm, identified as either spruce or larch (D.J. Christensen, written commun., 1983), yielded a radiocarbon age of 10,630±250 yr. BP (W–5348) (table 6).

Fossil mollusks were also recovered at the Howe Lake bog. Aquatic snails (*Gyraulus parvus*) and freshwater clams (*Pisidium variable*) were found at a depth of 300 to 310 cm (Emmett Evanoff, written commun., 1985). Because these mollusks are overlain by the wood fragment that yielded a radiocarbon age of 10,630±250 yr. BP and underlain by the Glacier Peak G ash, they indicate that this bog was a perennial pond rich in aquatic vegetation about 11,000 yr. BP.

The Mazama ash was not found in this bog. However, the Glacier Peak G ash was identified at a depth of 355 cm (table 4; fig. 12). The ash occurs in a very dark grayish brown (10 *YR* 3/2) gyttja as a light-gray (10 *YR* 7/1) sandy-silt layer about 1 cm thick. Grain-size analysis of this ash yielded a sand:silt:clay ratio of 35:46:19.

Ben Ryan Bog

Ben Ryan bog, at an altitude of 1,107 m, is about 2 km south-southeast of the town of West Glacier (fig. 10). Forest around the bog includes western white pine, lodgepole pine, Engelmann spruce, subalpine fir, western larch, and paper birch. The surrounding area is underlain by a hummocky end moraine deposited by the late Wisconsin McDonald glacier that contains many ponds and bogs.

At the Ben Ryan bog a hole was augered 260 cm into the underlying sediments (fig. 12). Here the Mazama ash (table 2; fig. 12) is a yellowish-brown (10 *YR* 5/4) silty layer, about 5 cm thick, at a depth from 100 to 105 cm at the base of a black (10 *YR* 2/1) peat. The Glacier Peak G ash (table 4; fig. 12)) was found at a depth of 255 cm. This ash is a white (10 *YR* 8/1) layer 1 cm thick in a grayish-brown (10 *YR* 5/2) organic clayey silt.

Snyder Ridge Bog

Snyder Ridge bog, at an altitude of 1,351 m, is on Snyder Ridge south of Lake McDonald and about 12 km northeast of the town of West Glacier (fig. 10). The bog is underlain by late Wisconsin till that overlies the Kishenehn Formation. Till in this area locally exceeds 30 m in thickness. Forest surrounding this bog consists of lodgepole pine, Engelmann spruce, subalpine fir, Douglas fir, and western larch.

At the Snyder Ridge bog a hole was augered to a depth of 410 cm (fig. 12). Here the Mazama ash (table 2) is a yellowish-brown (10 *YR* 5/4) layer, 15 cm thick at a depth between 50 and 65 cm. Grain-size analysis of this ash yielded a sand:silt:clay ratio of 11:66:23.

Two radiocarbon ages were obtained on samples below the Mazama ash (fig. 12). An age of 7,940±200 yr. BP (W–5572) was obtained on wood fragments from a depth of 110 cm in a black (10 *YR* 2/1) peat (table 6). These wood fragments were identified as willow (J.T. Quirk, written commun., 1984). An age of 9,650±250 yr. BP (W–5575) was obtained on yellowish-brown (10 *YR* 5/4) gyttja that contained coniferous wood fragments from a depth from 185 to 195 cm. The lower part of this gyttja contained the Glacier Peak G ash (table 4; fig. 12). Here the ash consisted of a white (10 *YR* 8/1) silty layer about 1 cm thick at a depth of 235 cm. This gyttja unit is underlain by 150 cm of a light-gray (10 *YR* 7/1) clay.

Discussion

The presence of the Glacier Peak G ash at four sites in the Lake McDonald area (Glacier Route 7, Howe Lake, Ben Ryan, and Snyder Ridge bogs) indicates extensive deglaciation of this area before 11,200 yr. BP. The fact that the Glacier Peak G ash at Snyder Ridge bog is underlain by 25 cm of gyttja and 150 cm of clay suggests that this site may have been ice free for some time before the deposition of the ash, and indicates that the glacier that filled the valley of the Middle Fork Flathead River no longer overtopped Snyder Ridge. In fact, because both the radiocarbon age of 9,650±250 yr.

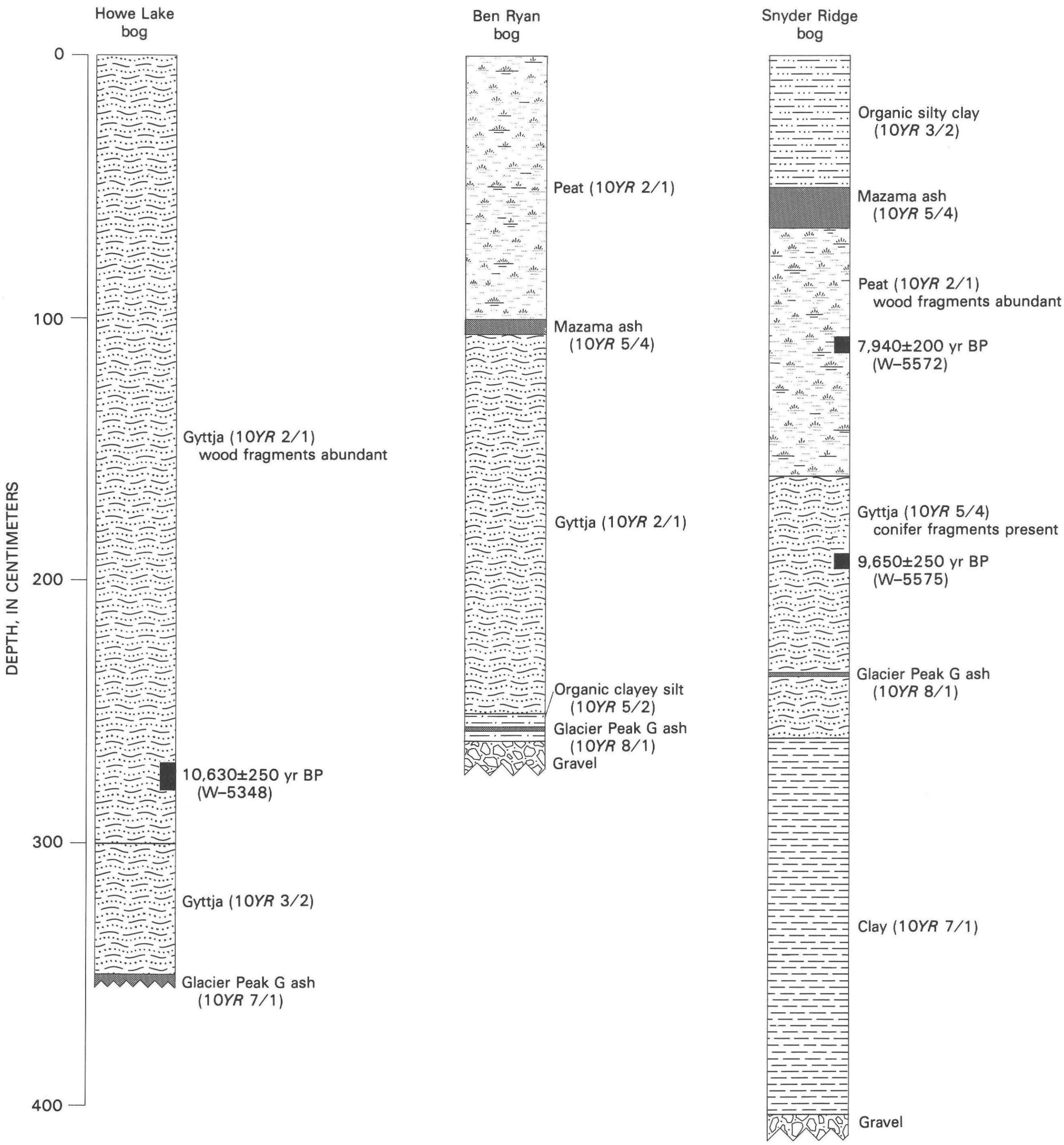


Figure 12. Stratigraphy of the Howe Lake, Ben Ryan, and Snyder Ridge bogs.

BP and the Glacier Peak G ash occur in the yellowish-brown gyttja (fig. 12), a rate of sedimentation was calculated between them and then extrapolated down to the base of the gyttja. In this way, an estimated age of about 12,000 yr. BP was obtained for the base of the gyttja at a depth of 260 cm. This age, determined by a sedimentation rate, is of course only an estimate, but does suggest that the area around the Snyder Ridge bog may have been ice free for some time before the deposition of the Glacier Peak G ash.

The presence of the Glacier Peak G ash at Ben Ryan bog provides a minimum age for the hummocky end moraine that this bog overlies and a minimum date for the withdrawal of ice from this section of the McDonald valley. Richmond (1986a) has suggested that this hummocky till, which he referred to as the Five Lakes moraine, marks the outer limit of a significant readvance of local mountain glaciers about 12,000 to 11,500 yr. BP. However, instead of representing a significant advance, this moraine could also have been formed by a stillstand during a general retreat. Whatever the Five Lakes moraine represents, the McDonald glacier had shrunk drastically from its late Wisconsin maximum by 11,200 yr. BP, and if it still existed was confined solely to its valley. This moraine is correlative with the hummocky end moraine at the lower end of Bowman Lake from which a minimum radiocarbon age of 11,150±90 yr. BP (W–5043) was obtained.

Minimum dates of revegetation in the Lake McDonald area are indicated by stratigraphic relationships and several radiocarbon ages of coniferous wood fragments from the bogs inspected in this area. That a lodgepole pine cone overlay the Glacier Peak G ash by about 10 cm at the Glacier Route 7 bog suggests that reforestation by that species may have taken place about 11,000 yr. BP. As previously mentioned, this cone was too small for a conventional radiocarbon age determination; it was combined with peat from overlying lenses within the mollusk-bearing unit and yielded an age of 9,920±120 yr. BP (W–5187). Hence, this age represents a very minimum date of reforestation. The radiocarbon age of 9,910±100 yr. BP (W–5191) of pine fragments from East McGee bog also is a minimum date of reforestation because the bottom of the dark-brown gyttja containing these coniferous wood fragments was not reached. Hence, the radiocarbon ages from both the Glacier Route 7 and East McGee bogs suggest the reintroduction of lodgepole pine into the Lake McDonald area before 10,000 yr. BP. In addition, the age of 10,630±250 (W–5348) on a spruce or larch fragment at Howe Lake bog provides a minimum age for the reintroduction of either of these species into this area.

Although the coniferous fragments from Snyder Ridge bog could not be identified to genus, their radiocarbon age of 9,650±250 yr. BP (W–5575) is a minimum date for the return of conifers to Snyder Ridge, the highest bog investigated in the Lake McDonald area. Based on sedimentation rates, an estimated age of about 12,000 yr. BP represents the beginning of organic sedimentation at this bog.

Marias Pass Area

Marias Pass, at an altitude of 1,595 m, is on the Continental Divide immediately south of Glacier National Park, about 15 km west of the mountain front (figs. 1 and 13). The area around the pass has a broad rolling topography and is mantled with thin scattered deposits of till and glaciofluvial sediments (Carrara, 1989), underlain mainly by gray and dark-gray Cretaceous marine mudstone (Mudge and Earhart, 1983). During late Wisconsin time, the Marias Pass area was a major accumulation area and thoroughfare for ice flowing east across the divide to the Two Medicine piedmont glacier. Calhoun (1906) indicated that the ice was about 550 m thick in the Marias Pass area at that time.

A local ice field west of Marias Pass supplied much of the ice to the Two Medicine glacier. This ice field may have been substantially thickened by the Flathead lobe of the Cordilleran ice sheet, which effectively blocked local mountain ice west of the divide, causing this ice field to thicken and spill east over Marias Pass (Richmond, 1986a). The Two Medicine glacier, fed by glaciers along a 60-km length of the mountain front, flowed east onto the plains to form the largest piedmont glacier in Montana: it extended about 55 km beyond the mountain front, had a maximum width of 50 km, and covered about 2,000 km^2 (fig. 13) (Calhoun, 1906; Alden, 1932).

Three sites were investigated in the Marias Pass area. In addition, a fourth site was investigated by C.W. Barnosky (Barnosky and others, 1987). Although widely separated, these sites provide information regarding the former Two Medicine glacier. They are (1) Marias Pass exposure, (2) Middle Fork bog, (3) Goat Lick exposure, and (4) Guardipee Lake (fig. 13).

Marias Pass Exposure

The Marias Pass exposure, at an altitude of 1,550 m, is about 2 km southwest of Marias Pass in a road cut along the south side of U.S. Highway 2. Forest surrounding this exposure includes lodgepole pine, Engelmann spruce, and subalpine fir. Here, a Glacier Peak ash was found in a section of postglacial laminated clayey-silt lake sediment (Richmond, 1972). Actually, the lake sediment contains two ashes, as the Glacier Peak G ash is underlain by the Mount St. Helens Jy ash (tables 4 and 5; fig. 14) (Carrara and others, 1986).

An exposure about 5 m high was excavated at the Marias Pass site (fig. 14). Most of the deposits consist of finely laminated (varves?), brown (10 *YR* 4/3) clayey-silt lake sediments. Grain-size analysis of these sediments indicated a sand:silt:clay ratio of 5:59:36. The laminae (940 were counted) consist of thin (1–3 mm) alternating fine and coarse layers. Several layers of brown (10 *YR* 4/3) fluvial gravelly sand 5–10 cm thick are present. Due to postglacial stream erosion and highway construction, it is not clear how the former lake was dammed. However, a nearby deposit of till suggests that the lake was dammed either by till or by a landslide of till shortly after deglaciation.

The upper ash, identified as the Glacier Peak G (Carrara and others, 1986), is 3 cm thick and consists of two layers. Its basal layer, 1 cm thick, is light gray (2.5 *Y* 7/1) and contains noticeably more mafic crystals and coarser glass shards than does the overlying fine-grained white (2.5 *Y* 8/2) layer. A combined grain-size analysis of these two layers indicated a sand:silt:clay ratio of

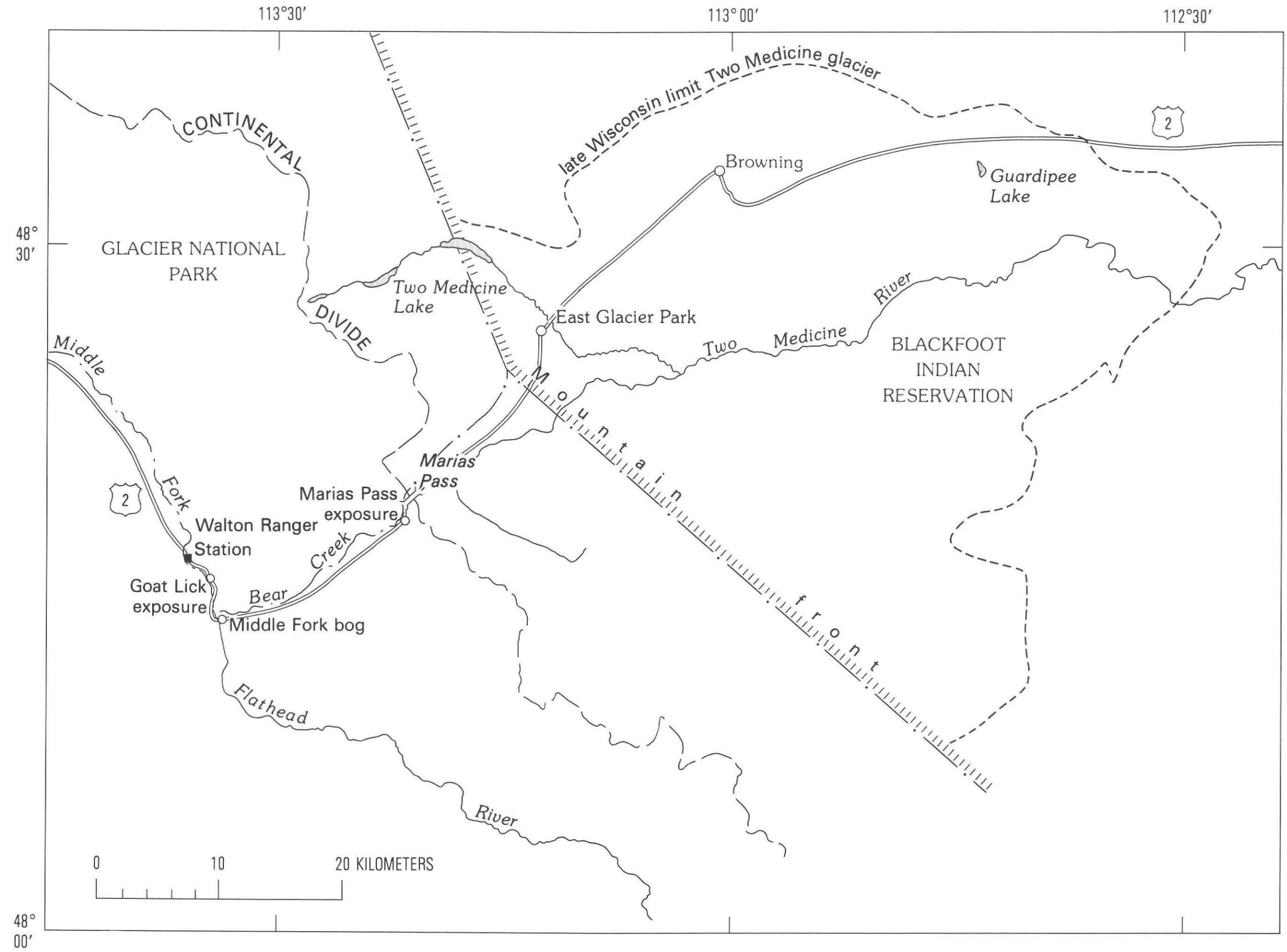


Figure 13. The Marias Pass area.

15:69:16. Above the ash is a layer of ash-rich, wavy-rippled clayey-silt lake sediment 2 cm thick (figs. 14 and 15). This layer of rippled sediment may be a turbidite (Bouma, 1962), and its presence suggests that the deposition of the Glacier Peak G ash interrupted normal sedimentation in the small lake. Above the wavy-rippled sediment is a zone of light-gray (10 *YR* 6/2), ash-rich, laminated lake sediment 13 cm thick.

The lower ash, a white (10 *YR* 8/1) layer 1 cm thick, identified as the Mount St. Helens Jy (Carrara and others, 1986), was found about 1 m below the Glacier Peak G ash near the bottom of the exposure (fig. 14). Grain-size analysis of this ash yielded a sand:silt:clay ratio of 9:69:22. Overlying this ash is a pale-brown (10 *YR* 6/3), ash-rich, laminated clayey-silt lake sediment, 10 cm thick, that grades upward into laminated, brown (10 *YR* 4/3) lake sediment.

Five samples of lake sediment from the Marias Pass exposure were examined for pollen. These samples were collected at about 30 cm intervals from immediately below the Glacier Peak G ash to immediately below the Mount St. Helens Jy ash. The pollen of these samples (samples 1 to 5, fig. 16) is compared to those of two modern surface samples (samples 0 and 00, fig. 16) from nearby forest duff. Pollen concentration values were generally low in the fossil samples. For samples 2 and 3, two slides were needed to count >100 grains.

The two modern samples (0 and 00) are dominated by pine pollen (82 and 88 percent) (fig. 16). Fir, alder, sagebrush (*Artemisia*), grass (Gramineae), spruce, and Douglas fir/larch percentages are low (<5 percent). Other pollen types were rare or absent in these samples.

The fossil pollen samples are also dominated by pine pollen (68 to 74 percent) (fig. 16). However, several differences distinguish the fossil pollen spectra from the modern pollen spectra: (1) the fossil samples contain more haploxylon pine pollen (whitebark pine type) than the modern samples, although diploxylon pine pollen (lodgepole pine type) still predominates; (2) spruce is more abundant in the fossil samples (reaching a high of 9 percent in sample 5); (3) sagebrush, grass, and other nonarboreal pollens are also more abundant in the fossil

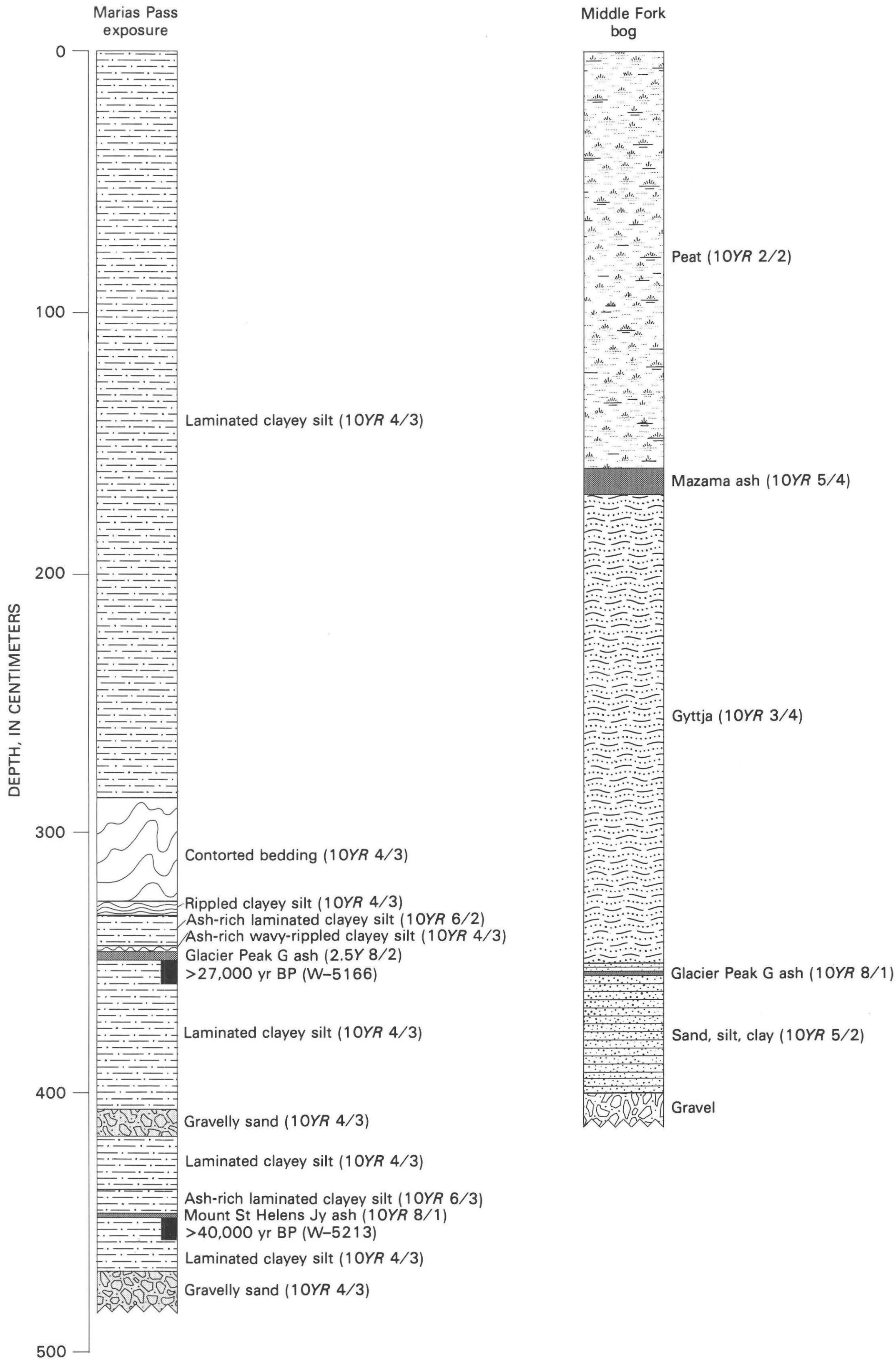


Figure 14. Stratigraphy of the Marias Pass exposure and Middle Fork bog.

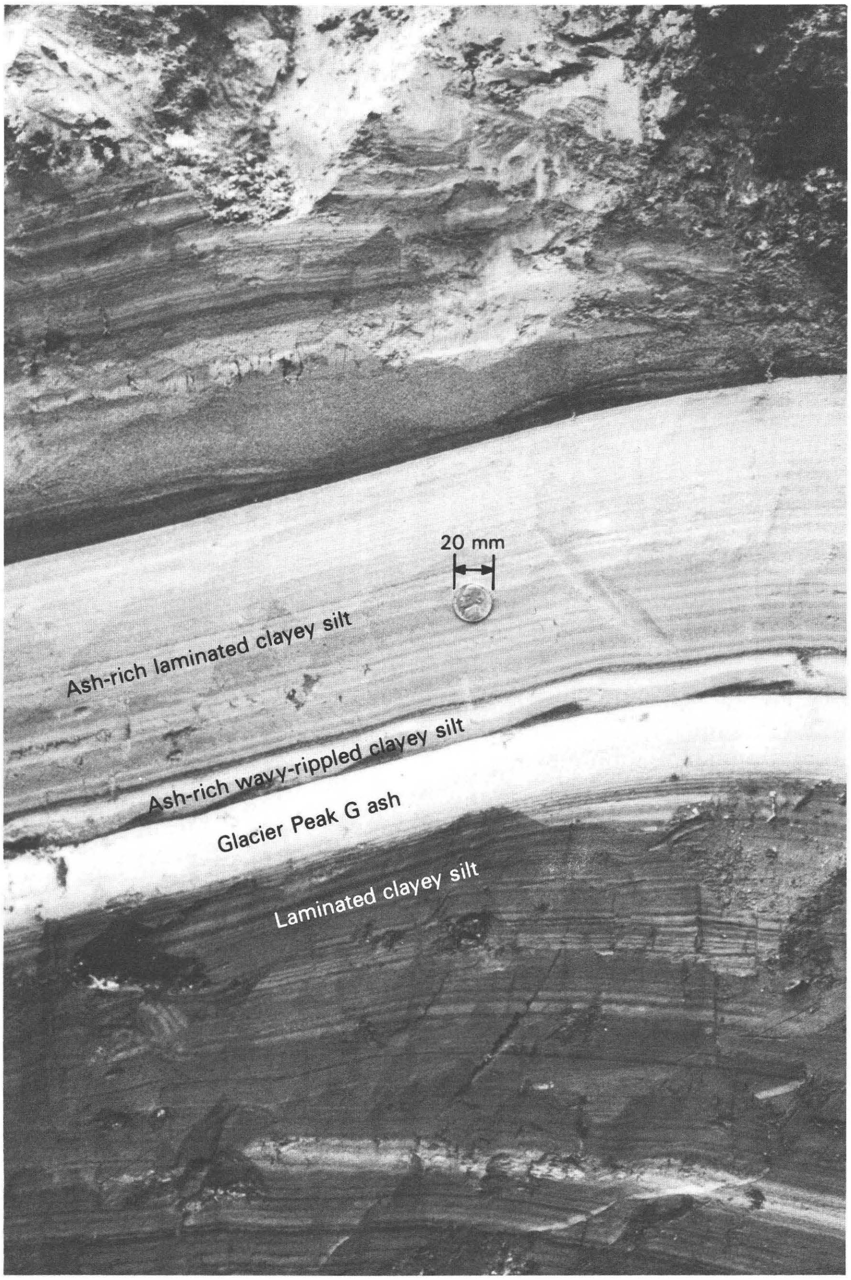
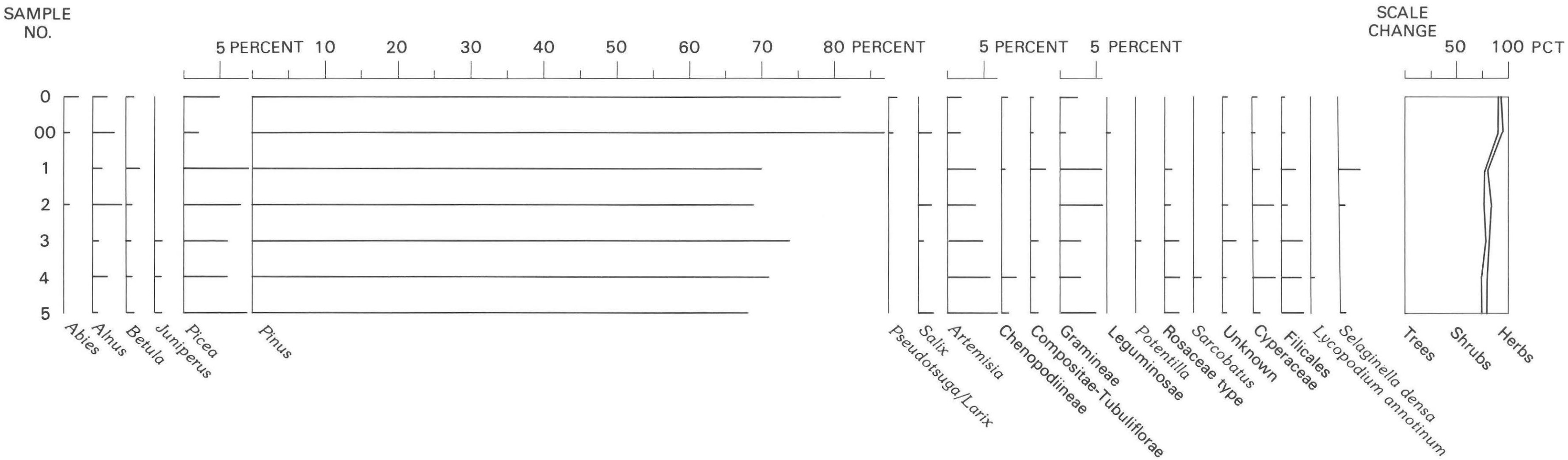


Figure 15. Glacier Peak G ash at the Marias Pass exposure.

samples; (4) fir pollen is rare and Douglas fir/larch pollen is absent in the fossil samples; and (5) several spore types, ferns (Filicales), stiff clubmoss (*Lycopodium annotinum*), and rock selaginella (*Selaginella densa*), that are typical of subalpine areas are present only in the fossil samples.

Plant macrofossils were found in the lake sediment below both ashes at the Marias Pass exposure. These macrofossils were recovered by washing the sediment through a 125 μm sieve. Although insufficient in quantity for conventional radiocarbon dating, these macrofossils indicate environmental conditions in the Marias Pass area at the time the Glacier Peak G and Mount St. Helens Jy ashes were deposited.

A conifer needle, an alder strobilus, and an unidentifiable wood fragment were found in the sediment immediately below the Glacier Peak G ash. The sediment immediately below the Mount St. Helens Jy ash contained most of the macrofossils. Several small woody fragments were identified as either *Salix* (willow) or



EXPLANATION

Sample No.	Location	No. of grains counted
0	Surface sample	300
00	Surface sample	300
1	Immediately below Glacier Peak G ash	198
2	23 cm below sample No. 1	134
3	58 cm below sample No. 1	144
4	88 cm below sample No. 1	200
5	Immediately below Mount St. Helens Jy ash	200

Figure 16. Percentage pollen diagram from the Marias Pass exposure.

Populus (aspen, cottonwood, and poplar) (D.J. Christensen, written commun., 1983). As *Populus* does not presently inhabit the area, these woody fragments are probably willow.

Fossil insects were also identified in the sediment immediately below the Mount St. Helens Jy ash (Elias, 1988). Insects provide environmental information, especially those that are predaceous or scavengers: because they can respond quickly to changing environments, predators and scavengers can exploit newly deglaciated regions before revegetation, which may take several hundred or possibly several thousand years. Insects identified include; (1) *Tachyporus rulomus*, (2) *Bembidion sulcipenne hyperboroides*, (3) *Amara discors*, (4) *Thanatophilus trituberculatus*, and (5) *Formica neorufibarbis*.

T. rulomus, a staphylinid beetle, commonly is found across boreo-arctic and montane North America. This species lives in wet habitats including damp leaf litter, mosses, and the borders of lakes and marshes, and is the most common arctic *Tachyporus* (Elias, 1988). *B. sulcipenne hyperboroides*, a riparian ground beetle, commonly inhabits the gravel banks of streams. This species is known only from northern British Columbia, the Yukon Territory, and Alaska (Lindroth, 1963, 1969). *A. discors*, a ground beetle, lives in sandy, open habitats (Lindroth, 1968) within the boreo-montane regions of North America. This species has been reported from three alpine sites in Glacier National Park (Edwards, 1975). *T. trituberculatus*, a beetle associated with carrion, reaches the arctic tundra in Canada and is not known from the lower 48 States (Elias, 1988). *F. neorufibarbis*, an ant, commonly is found in the upper subalpine and just above treeline in the Rocky Mountains and as far north as the arctic treeline. It is the most cold-adapted species of North American *Formica* (Francoeur, 1973).

Middle Fork Bog

The Middle Fork bog, at an altitude of 1,317 m, is near the southwest corner of Glacier National Park about 6 km south of the Walton Ranger Station (fig. 13). The area around this bog is underlain by rock of the Belt Supergroup covered in places with a veneer of late Wisconsin glacial and fluvial deposits. Forests surrounding this bog consist of western white pine, lodgepole pine, Engelmann spruce, subalpine fir, and western larch.

The bog is in a small depression near the summit of a hill about 0.5 km southeast of the confluence of the Middle Fork Flathead River and Bear Creek. Here, a hole was augered to a depth of 400 cm (fig. 14). The Mazama ash (table 2) is a yellowish-brown (10 *YR* 5/4) layer 10 cm thick, at a depth from 160 to 170 cm at the base of a very dark brown (10 *YR* 2/2) peat. Grain-size analysis of the Mazama ash yielded a sand:silt:clay ratio of 8:67:25. The Glacier Peak G ash (table 4) was found at a depth of 352 cm in the upper part of a grayish-brown (10 *YR* 5/2) sand-silt-clay unit. This ash is a white (10 *YR* 8/1) silty layer about 1 cm thick. The ash is underlain by about 50 cm of fine-grained sediment that overlies gravel.

Goat Lick Exposure

The Goat Lick exposure is in the southwest corner of Glacier National Park at an altitude of 1,195 m, about 2 km south of the Walton Ranger Station (fig. 13). This exposure is about 200 m north of the Goat Lick Viewpoint along U.S. Highway 2. The exposure is in the face of a gravel terrace about 30 m thick that overlies bedrock of the Belt Supergroup. During the late Wisconsin, this site, like the Middle Fork bog, was covered by hundreds of meters of ice by the glacier that occupied the valley of the Middle Fork Flathead River.

The Glacier Peak G ash (table 4) is in the terrace face about 30 m above the Middle Fork Flathead River (fig. 17). The ash forms a light-gray (10 *YR* 7/1) silty layer 2–4 cm thick in a sand and gravel layer about 1.5 to 2 m below the terrace surface (fig. 17).

Guardipee Lake Site

Guardipee Lake is a shallow lake, about 0.5 km^2 in area, on the plains east of Glacier National Park, at an altitude of 1,235 m, about 21 km east of the town of Browning (fig. 13). The lake is situated in broad rolling topography, underlain by thin, scattered deposits of late Wisconsin till deposited by the former Two Medicine glacier. Cretaceous sedimentary rock underlies this area (Mudge and Earhart, 1983).

Guardipee Lake was cored by C.W. Barnosky for pollen analysis in August 1984 (Barnosky and others, 1987, fig. 3). The core was more than 6 m long and consisted of gray clays, gray silty clays, interbedded marls and clays, and interbedded silts and sands. Two white (10 *YR* 8/1) ashes about 1 cm thick and 18 cm apart, the Glacier Peak G and Mount St. Helens Jy (tables 4 and 5), were identified in the lower meter of the core (Barnosky and others, 1987).

Discussion

The presence of the Mount St. Helens Jy ash at two of the sites in the Marias Pass area (Marias Pass exposure and Guardipee Lake) and the Glacier Peak G ash at all four sites in this area indicates extensive deglaciation of the area occupied by the former Two Medicine glacier before 11,400 and 11,200 yr. BP.

Figure 17. The Goat Lick exposure; Middle Fork Flathead River in foreground.

The presence of the Glacier Peak G and Mount St. Helens Jy ashes at the Marias Pass exposure indicates that the Continental Divide in this area was deglaciated before 11,400 yr. BP (Carrara and others, 1986). The presence of the Glacier Peak G ash at the Middle Fork bog and Goat Lick exposure indicates that the large trunk glacier that filled the valley of the Middle Fork Flathead River had retreated from this part of the valley by 11,200 yr. BP. Hence, by that time former tributary glaciers heading along the western flank of the Lewis Range and the eastern flank of the Flathead Range, if they still existed, were confined to their own valleys. Furthermore, the stratigraphy at the Middle Fork bog and Goat Lick exposure suggests that the age of the Glacier Peak G ash (11,200 yr. BP) may represent only a very minimum date of deglaciation. At the Middle Fork bog, this ash is underlain by 50 cm of fine-grained sediment (fig. 14). At the Goat Lick exposure the ash is underlain by at least 30 m of fluvial gravels. Hence, the stratigraphy at both sites suggests that they were deglaciated several hundred or possibly several thousand years before the Glacier Peak G ash was deposited. The stratigraphic position of the Glacier Peak G ash at the Middle Fork bog indicates that organic material colonized this area shortly after the deposition of this ash.

Plant macrofossils, pollen, and spores from the Marias Pass exposure suggest that the local vegetation in this area consisted of an open landscape with shrubs, herbs, and scattered conifers when the Glacier Peak G and Mount St. Helens Jy ashes were deposited. At that time there was a larger proportion of spruce and whitebark pine than at present (Carrara and others, 1986). These vegetative data are characteristic of the tundra/subalpine forest boundary (treeline). As the Marias Pass exposure is at an altitude of 1,550 m, whereas present-day treeline is at about 2,050 m, these data indicate that treeline at the time these ashes were deposited was about 500 m lower than present.

The fossil insect data suggest that this 500 m lowering of treeline may be a minimum estimate. Elias (1988) noted the lack of any insect species associated with coniferous forest among the fossil insect assemblage. Insect assemblages from other sites at or above treeline have generally contained considerable numbers of coniferous-bark beetles because of the tendency for these insects to be blown upslope (Elias, 1988). Hence, Elias suggested that the lack of any such insect species in the fossil insect assemblage below the Mount St. Helens Jy ash could be interpreted as indicating that treeline could have been as much as 200 m below the Marias Pass exposure. Therefore, at about 11,400 yr. BP, treeline was probably at least 500 m lower than present, and may have been as much as 700 m lower.

Pollen data from Guardipee Lake indicate that the area around this site has been treeless for the past 12,000 years. During glacial times, conifers may have survived in small populations along the lower mountain slopes bounded by tundra or steppe vegetation at both higher and lower altitudes (Barnosky and others, 1987). The ratio between sagebrush pollen and grass pollen suggests that the area was a grassland before 11,400 yr. BP, as opposed to a sagebrush steppe (Barnosky and others, 1987).

Problems with radiocarbon age determinations in the Glacier National Park region are demonstrated by an attempt to date both the Glacier Peak G and Mount St. Helens Jy ashes at the Marias Pass exposure. Blocks of lake sediment about 10 cm thick were collected from under both ashes. Although the amount of organic matter in these sediments was quite low (<5 percent-lost on ignition), it was concentrated in the laboratory using a technique modified from Kihl (1975), and the <125-μm fraction was submitted for radiocarbon age determination. Ages obtained for the sediments underlying the Glacier Peak G and Mount St. Helens Jy ashes were $>27,000$ yr. BP (W–5166) and $>40,000$ yr. BP (W–5213), respectively (fig. 14). These ages are obviously older than the true age of the sediment. Much of the fine organic material in these lake sediments was probably derived from the Cretaceous mudstones in the area.

These radiocarbon ages underscore the problems of contamination of organic material with older radiometrically inactive carbon in the Glacier National Park region. Other possible sources of error include the presence of reworked lignite from the Kishenehn Formation in the valleys of the North and Middle Forks Flathead River and reworked organic material from black Cretaceous shales on the east side of the Continental Divide beyond the mountain front. These sources of error highlight the importance of the Mazama, Glacier Peak G, and Mount St. Helens Jy ashes in this region, because the ashes are reliable time-stratigraphic markers of known age.

Lower St. Mary Lake Area

Lower St. Mary Lake lies to the east of Glacier National Park on the adjacent Blackfoot Indian Reservation (figs. 1 and 18). The lake is 9 km long and occupies a former glacial valley that is dammed by a large alluvial fan deposited by Swiftcurrent Creek. During the late Wisconsin, this valley was occupied by the St. Mary glacier, that headed on the eastern side of the Lewis Range along the Continental Divide, from the Logan Pass area south to the Blackfoot Mountain area. This glacier flowed down the St. Mary valley, past the mountain front and north across the U.S.-Canada border, where its terminal deposits lie beneath till of a subsequent advance of the Laurentide ice sheet. Tributary glaciers flowed from the valleys of Divide, Red Eagle, Boulder, Swiftcurrent, Kennedy, and Otatso Creeks in Glacier National Park. One site was investigated in this area, the Chewing Blackbones bog (fig. 18).

Chewing Blackbones Bog

The Chewing Blackbones bog is along the eastern side of Lower St. Mary Lake at an altitude of 1,448 m (fig. 18). The bog is in a small valley trending north to south between two late Wisconsin lateral moraines deposited by the former St. Mary glacier. Thick clusters of small aspen containing an occasional Douglas fir dot the surrounding hillsides. The site is underlain by a veneer of late Wisconsin till (deposited by the former St. Mary glacier), which overlies the Two Medicine Formation, an Upper Cretaceous nonmarine mudstone (Mudge and Earhart, 1983).

At the Chewing Blackbones bog a hole was augered to a depth of 350 cm (fig. 19). The Mazama ash (table 2) is 25 cm thick at this site and forms a light-brownish-gray (2.5 *Y* 6/2) layer at a depth from 105 to 130 cm. Grain-size analysis of this ash yielded a sand:silt:clay ratio of 8:70:22. The Glacier Peak G ash (table 4) was found at a depth of 240 cm in an inorganic greenish-gray (5 *GY* 5/1) silty clay. The ash is a light-gray (10 *YR* 7/2) silty layer 1 cm thick and is underlain by 110 cm of fine-grained sediments.

Discussion

The presence of the Glacier Peak G ash at the Chewing Blackbones site indicates that the former St. Mary glacier had retreated substantially from its late Wisconsin position before 11,200 yr. BP. Although the Chewing Blackbones bog is beyond the mountain front, the presence of the Glacier Peak G ash indicates that the St. Mary glacier was confined to its own valley by 11,200 yr. BP and no longer merged with the other glaciers to the north flowing from the valleys of Boulder, Swiftcurrent, Kennedy, and Otatso Creeks; however, it could still have been receiving contributions of ice from glaciers in

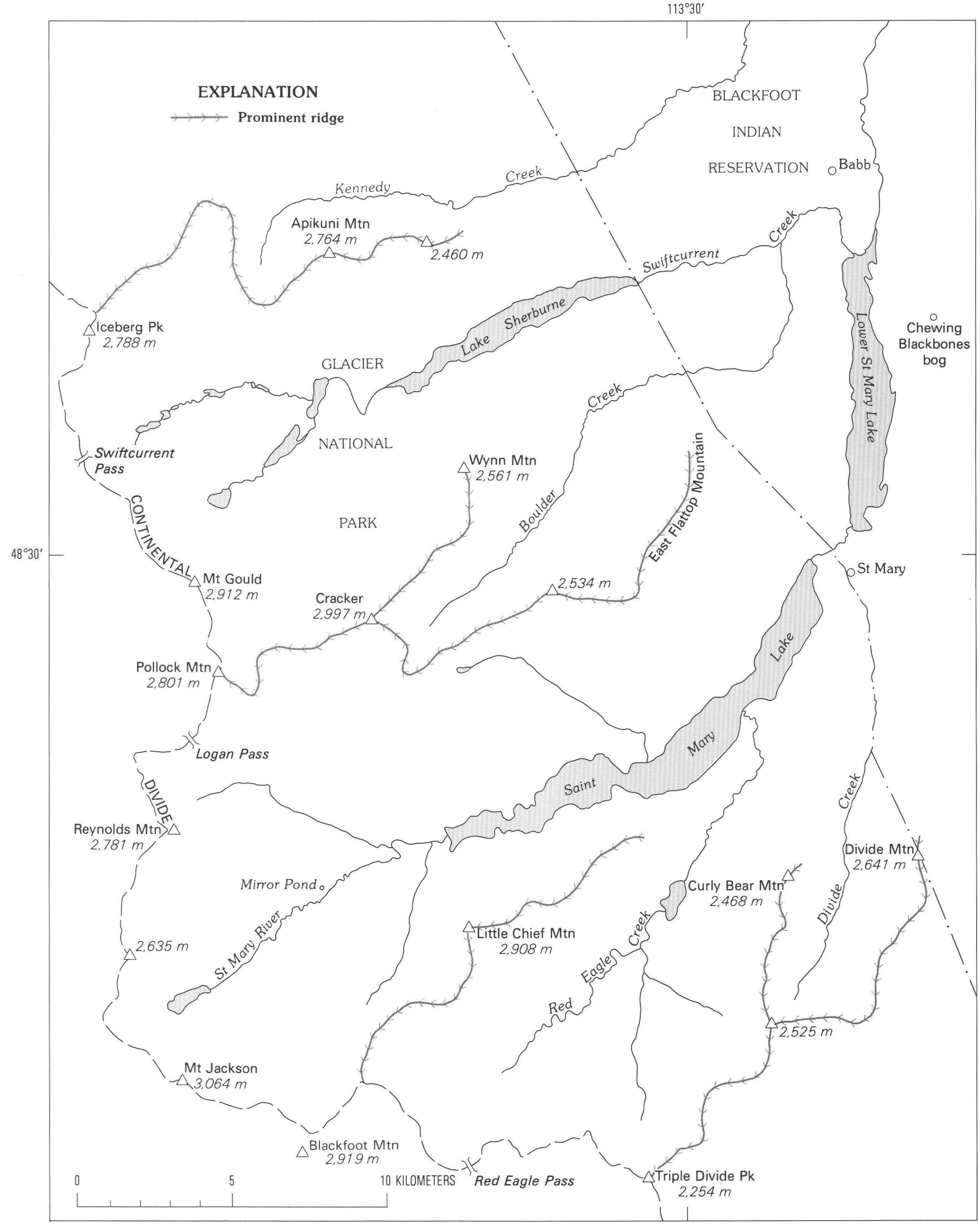


Figure 18. The Lower St. Mary Lake area.

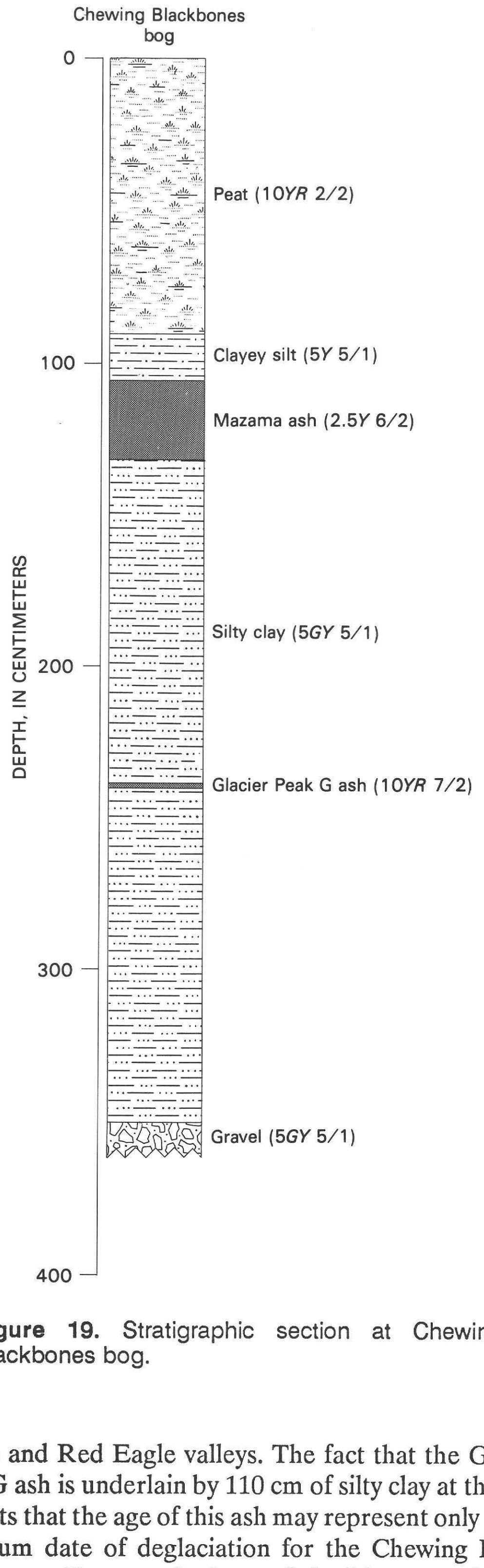


Figure 19. Stratigraphic section at Chewing Blackbones bog.

Divide and Red Eagle valleys. The fact that the Glacier Peak G ash is underlain by 110 cm of silty clay at this bog suggests that the age of this ash may represent only a very minimum date of deglaciation for the Chewing Blackbones area. Because the base of the Mazama ash rests upon this silty-clay unit in which the Glacier Peak G ash occurs (fig. 19), a rate of sedimentation was calculated between these two ashes and extrapolated down to the base of this unit at 350 cm. The extrapolated sedimentation rate yielded an age of about 15,500 yr. BP. This age is of course only an estimate, but indicates that the age of the Glacier Peak G ash is a very minimum date of deglaciation for the area around the Chewing Blackbones bog.

Kootenai Lakes Area

The Kootenai Lakes area lies in the north-central section of Glacier National Park (figs. 1 and 20). The area is underlain by till, in turn overlying rocks of the Belt Supergroup. Forest surrounding this area contains lodgepole pine, Engelmann spruce, and subalpine fir. During the late Wisconsin, this area was overlain by a large glacier several hundred meters thick that headed along the eastern flank of the Livingston Range and the western flank of the Lewis Range. This glacier flowed north down the Waterton valley into Canada. Two sites from the same bog (Mount Cleveland bog) were investigated in this area (fig. 20).

Mount Cleveland Bog-Hole #1

The Mount Cleveland bog, at an altitude of 1,341 m, lies in the Waterton valley about 3 km north-northeast of Kootenai Lakes and northwest of Mount Cleveland (fig. 20). At this bog, hole #1 was augered 540 cm into the underlying sediments (fig. 21). The Mazama ash

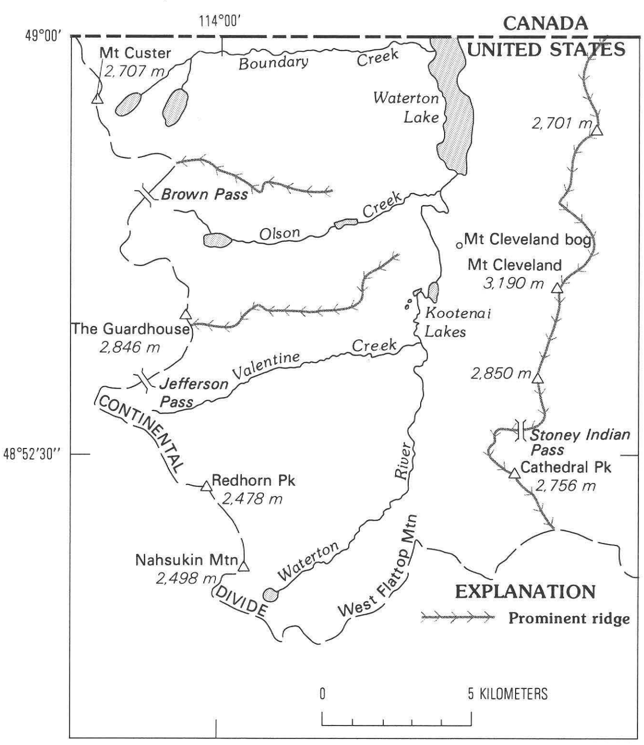


Figure 20. The Kootenai Lakes area.

forms a grayish-brown (10 *YR* 5/2) layer 3 cm thick at a depth between 200 and 203 cm, in a very dark brown (10 *YR* 2/2) peat. Grain-size analysis of this ash yielded a sand:silt:clay ratio of 10:60:30. Wood fragments, collected from a depth of 250 to 290 cm and identified as pine (J.T. Quirk, written commun., 1984), yielded a radiocarbon age of 9,510±350 yr. BP (W–5571) (table 6).

Mount Cleveland Bog-Hole #4

Hole #4 lies about 100 m west of hole #1. Hole #4 was augered to a depth of 380 cm (fig. 21). The Mazama ash forms a brown (10 *YR* 5/3) layer 10 cm thick at a depth from 145 to 155 cm in a dark-reddish-brown (5 *YR* 3/3) peat. Grain-size analysis of this ash yielded a sand: silt:clay ratio of 8:67:25. Organic material, containing willow fragments (J.T. Quirk, written commun., 1984) were recovered from an olive (5 *Y* 4/3) gyttja at a depth of 330 to 380 cm. A radiocarbon age of 9,850±260 yr. BP (W–5577) was obtained on this material (table 6).

Discussion

The radiocarbon ages from the two holes augered in the Mount Cleveland bog provide both minimum dates of revegetation and reforestation and very minimum dates of deglaciation for the Kootenai Lakes area. The age of 9,510±350 yr. BP of pine fragments from hole #1 indicates a minimum date for the reintroduction of lodgepole pine into this area. The fact that these wood fragments are underlain by 160 cm of organic sediment (fig. 21) suggests that revegetation occurred at least several hundred or possibly several thousand years before this date. That this organic sediment is underlain by 90 cm of pale-red (2.5 *YR* 5/2) laminated clayey silt also suggests that a significant amount of time may have elapsed between deglaciation of this site and revegetation. In support of this suggestion, a sedimentation rate was calculated for the interval between the Mazama ash and the radiocarbon age of 9,510±350 yr. BP and then extrapolated down to the base of the organic silty clay at a depth of 450 cm (fig. 21). In this way an age of about 16,600 yr. BP was obtained for the base of the silty clay. Although this age is only a crude estimate, it does suggest that deglaciation of the Kootenai Lakes area may be much older than the radiocarbon age of pine fragments from the bog.

The age of 9,850±260 yr. BP for organic sediment containing willow fragments in hole #4 (fig. 21) is a minimum date for the reintroduction of willow following deglaciation of the Kootenai Lakes area. This radiocarbon age is of material obtained from the lower 50 cm of this hole, but the organic material in the lowermost several centimeters is probably much older than the radiocarbon age, which represents the average age of organic material in this 50-cm interval.

The radiocarbon ages and associated stratigraphy at the Mount Cleveland bog suggest that revegetation in the Kootenai Lakes area was underway before 10,000 yr. BP. Lodgepole pine was reestablished by at least 9,500 yr. BP. Both radiocarbon ages from this area are minimum dates for the reintroduction of lodgepole pine and willow. However, the associated stratigraphy suggests that some time elapsed between deglaciation and revegetation, and therefore deglaciation is tentatively suggested at about 11,000 yr. BP. Hence, this study suggests that by 11,000 yr. BP the glacier that once filled the Waterton valley no longer extended more than about 10 km from its headwall, if it still existed at all.

DEGLACIATION AND REVEGETATION ELSEWHERE IN THE ROCKY MOUNTAINS

The dates of deglaciation and revegetation from the Glacier National Park region are similar to other dates elsewhere in the Rocky Mountains. In the upper Elk River valley of British Columbia, a radiocarbon age of 13,430±450 yr. BP (GX–5599) was obtained from Weary bog (Ferguson and Osborn, 1981). This bog, at an altitude of 1,582 m, is 29 km south of the Mount Joffre Ice Field. This radiocarbon age was determined from gastropod shells and may have been older than the true age of the shells, due to the incorporation of older carbonate (Ferguson and Osborn, 1981). Ages of 11,900±100 yr. BP (GSC–2142) and 12,200±160 yr. BP (GSC–2275) were obtained on material containing wood and unidentifiable organic matter collected along a terrace scarp about 30 km downvalley from Weary bog (Harrison, 1976). During the late Wisconsin, the Elk River valley was filled by a large glacier about 100 km in length; this glacier was tributary to Cordilleran ice in the Rocky Mountain trench that flowed further south into the Flathead Lake area of Montana (Waitt and Thorson, 1983). Hence, the radiocarbon ages from the Elk River valley indicate extensive deglaciation of that valley and also the Rocky Mountain trench by about 12,000 yr. BP and possibly as early as 13,500 yr. BP. In addition, because the Mount Joffre Ice Field lies on the Continental Divide between the Elk River and Kananaskis River valleys, and therefore also supplied ice to the glacier in the Kananaskis River valley, the radiocarbon ages also indicate extensive deglaciation of that valley.

In the Sun River Canyon area of Montana, about 100 km south of Glacier National Park, extensive

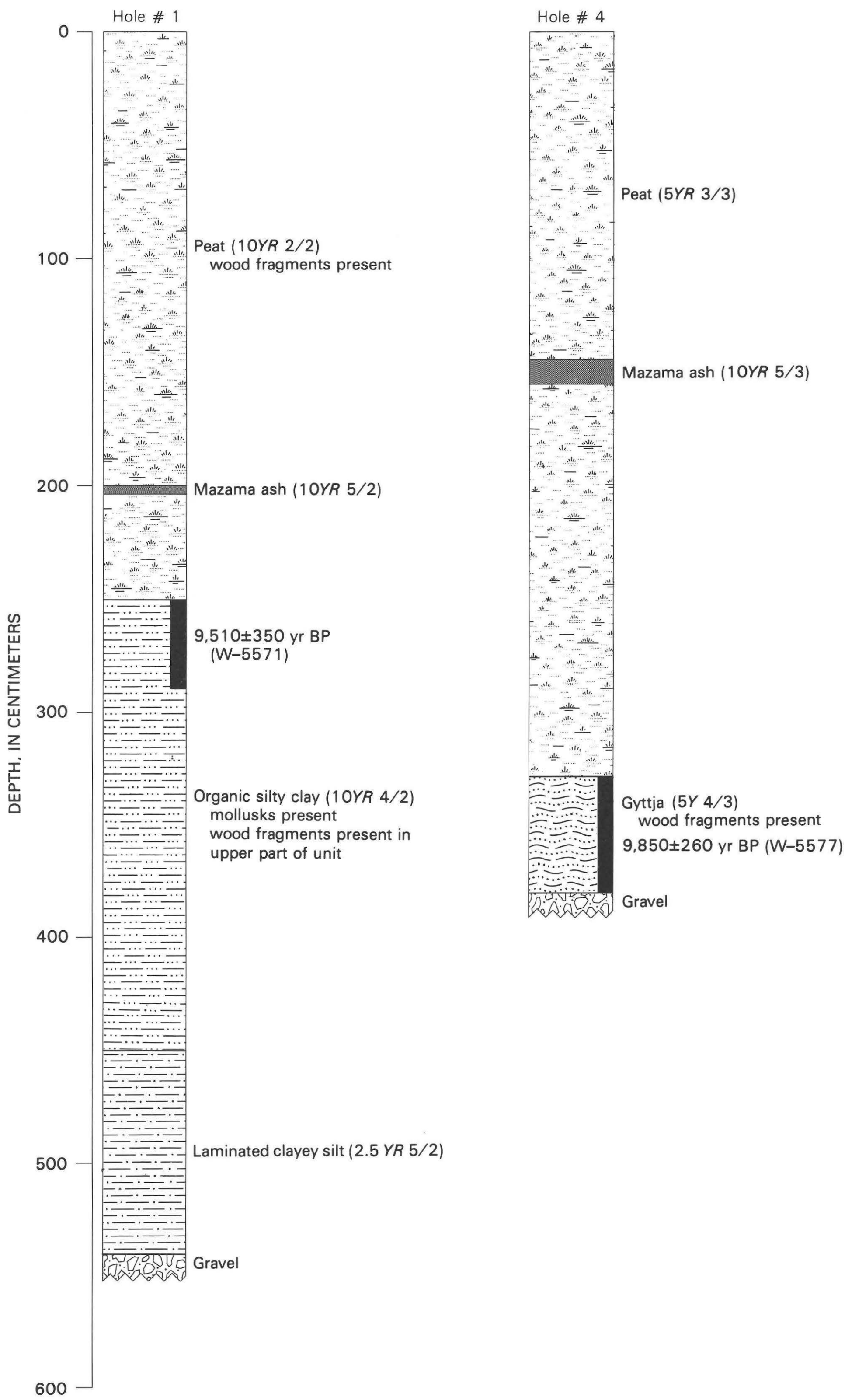


Figure 21. Stratigraphy of the Mount Cleveland bog, holes #1 and #4.

deglaciation occurred before the deposition of the Glacier Peak G ash, 11,200 yr. ago (Lemke and others, 1975; Carrara and others, 1986). This canyon was occupied by a large glacier that flowed beyond the canyon mouth and formed a piedmont glacier 550 km^2 in area on the plains (Calhoun, 1906; Alden, 1932). The presence of the Glacier Peak G ash at several sites in the canyon, including a site 11 km upstream from the canyon mouth, indicates extensive deglaciation of this area before 11,200 yr. BP. Furthermore, stratigraphic relationships suggest that deposition of the Glacier Peak G ash in the Sun River Canyon may postdate deglaciation by several thousand years (Carrara and others, 1986).

In Yellowstone National Park, Wyo., radiocarbon age determinations indicate that deglaciation of the Pinedale ice cap of the Yellowstone Plateau was largely completed by 14,000 to 13,000 yr. BP (Pierce, 1979). This large ice cap covered about 3,400 km^2, averaged more than 700 m in thickness, and produced an outlet glacier that flowed 60 km down the Yellowstone valley (Pierce, 1979).

In Colorado, a radiocarbon age of 13,680±110 yr. BP (W–5028) was obtained from a bog on the Continental Divide in the southern Park Range (Madole, 1986). This area had been covered by a broad ice cap from which large valley glaciers descended. An icefield in the northwestern San Juan Mountains of Colorado, the largest glacier in the southern Rocky Mountains, no longer existed by 15,000 radiocarbon years ago (Carrara and others, 1984).

In the western United States, deglaciation may have been essentially complete by about 11,000 yr. BP. In the North Cascade Range of Washington, cirques were deglaciated before the Glacier Peak ash eruptions, 11,200 yr. ago (Beget, 1984). Beget (1984) further concluded that the glacier cover in the North Cascade Range was similar to that of the present by that time. Pollen records from the western United States can be interpreted as indicating that by about 11,500 yr. BP, Pinedale glaciers had probably disappeared or were about as small as subsequent late Holocene glaciers (Porter and others, 1983).

Similar conclusions regarding the timing of deglaciation have been reached for the Canadian Rockies. In Jasper National Park, Alberta, late Wisconsin ice receded from Tonquin Pass, on the Continental Divide, before 10,000 yr. BP (Luckman and Osborn, 1979; Kearney and Luckman, 1983a). Luckman and Osborn (1979) also concluded that by 10,000 yr. BP glaciers in the middle Canadian Rocky Mountains had receded to positions close to present-day glacial limits. Organic sedimentation began before 9,500 yr. BP in an alpine bog close to its headwall in Watchtower basin, in Jasper National Park (Luckman and Kearney, 1986). In Yoho National Park, British Columbia, late Wisconsin ice receded from the Lake O'Hara basin before 10,060±160 yr. BP (RIDDL–511) (Reasoner and Rutter, 1988). This radiocarbon age determination indicates that the Bow Valley glacier had receded to cirques adjacent to the Continental Divide before 10,000 yr. BP. Because pollen and macrofossil data indicate that a pine-fir forest was also established in this basin by that time, the radiocarbon age is a minimum date of deglaciation.

Evidence from the Rocky Mountains also indicates that revegetation was underway by the time the Glacier Peak ash was erupted 11,200 yr. ago. In the Kootenai River valley, about 150 km west of Glacier National Park, a sediment core from Tepee Lake was collected by Mack and others (1983). This lake is at an altitude of 1,270 m, and the surrounding forest contains grand fir, lodgepole pine, western white pine, ponderosa pine, western larch, Engelmann spruce, and Douglas fir. Five pollen zones were recognized in the sediment core from Tepee Lake. The basal zone (pollen zone 1) contains the Glacier Peak G ash and is dominated by pine pollen (56 percent haploxylon). Other important arboreal pollen includes spruce, fir, alder, and willow. The nonarboreal pollen contribution is relatively small and consists mostly of sagebrush and grass (both ≤ 7 percent). The sunflower family (Compositae) and goosefoot family (Chenopodiaceae) are also represented (Mack and others, 1983). These pollen spectra, similar to those from the Marias Pass exposure (fig. 16), are thought to represent a cool, moist climate. The closest modern pollen rain analog is within stands of whitebark pine and subalpine fir that presently grow at higher altitudes (Mack and others, 1983).

At Sheep Mountain bog, about 150 km south of Glacier National Park, Glacier Peak ash is dominated by pine pollen (about 55 percent) consisting of about equal amounts of haploxylon and diploxylon pine pollens (Mehringer, 1985). This bog is at an altitude of 1,920 m and is presently surrounded by forest containing ponderosa pine, lodgepole pine, whitebark pine, western larch, Douglas fir, subalpine fir, and Engelmann spruce. The distribution of fossil needles and cones in the sampled section indicates that Douglas fir has grown near this site since at least 10,000 yr. BP.

At Lost Trail Pass bog, on the Continental Divide, about 300 km south of Glacier National Park, a core was obtained for pollen analysis (Mehringer, Arno, and Petersen, 1977). This bog lies at an altitude of 2,152 m, and the surrounding forest contains lodgepole pine, whitebark pine, Engelmann spruce, and subalpine fir. Pollen zone 2, containing the Glacier Peak ash, has percentages of pine, spruce, alder, grass, sagebrush, and sedge (Cyperaceae) pollen similar to those of the Marias

Pass exposure. Hence, at about 11,200 yr. BP Lost Trail Pass bog was probably surrounded by an open forest similar to the mixed whitebark pine/lodgepole pine forest currently found at higher altitudes, indicating a climate slightly cooler than present (Mehringer, Arno, and Petersen, 1977).

At Cub Creek Pond, in Yellowstone National Park, about 450 km south-southeast of Glacier National Park, a core was taken for pollen analysis (Waddington and Wright, 1974). The pond lies at an altitude of 2,485 m, surrounded by forest containing lodgepole pine, with some stands of spruce and fir. Pollen zone 1, which immediately overlies the Glacier Peak B ash (Westgate and Evans, 1978), contains relatively high percentages of sagebrush, spruce, juniper, birch, willow, grass, and sedge pollen. This pollen data was interpreted to suggest that upper treeline was about 500 m lower than present (Waddington and Wright, 1974). This estimate of treeline lowering agrees well with that of 500 to 700 m from the Marias Pass exposure.

In the Front Range of Colorado, Benedict (1973) obtained a radiocarbon age of 9,200±135 yr. BP (I–6520) for wood from a bog near Satanta Peak. Basal muck and peat samples from two nearby bogs, in the present krummholz zone, contained spruce cones and wood fragments that gave weighted average ages of 9,590±130 yr. BP (I–12,198 A,B) and 8,885±120 yr. BP (I–12,787 A,B) (J.B. Benedict, written commun., 1983). These radiocarbon ages indicate that treeline in the Front Range had risen to at least its present-day altitude by 9,500 yr. BP.

In the San Juan Mountains of southwest Colorado, 22 radiocarbon ages of coniferous wood (spruce and fir) fragments were obtained from the Lake Emma site, above the present-day krummholz limit (Carrara and others, 1984). These radiocarbon ages indicate that treeline in this mountain range was at least 80 m higher than present by 9,600 yr. BP .

MORAINES IN THE HIGH AREAS OF GLACIER NATIONAL PARK

Moraines of two different age groups have been identified in the high areas of Glacier National Park (Osborn, 1985; Carrara, 1987; Carrara and McGimsey, 1988). Most of these moraines lie at altitudes between 1,900 and 2,400 m and are generally found within a kilometer or two of their cirque headwalls. Both age-groups of moraines lie in front of the present-day glaciers and snowfields, indicating that glacier fluctuations since the late Wisconsin have been small.

Moraines of the Older Group

Location and Description

Moraines of the older group have been identified at 25 sites in Glacier National Park. Many of these sites are in the central part of the Lewis Range, from the Sperry Glacier area east to the Triple Divide Pass area (fig. 1). In the Livingston Range, moraines of this group have been identified at only two sites—one near the former Agassiz Glacier, the other in the Boulder Pass area. These moraines are commonly well vegetated, supporting a dense cover of alpine tundra and krummholz (dwarf, windswept trees above timberline) of subalpine fir and Engelmann spruce. These moraines are rarely more than 10 m high (fig. 22) and generally extend only short distances downvalley from moraines of the younger group (Carrara and McGimsey, 1981, 1988; Carrara, 1987). However, in the Upper Two Medicine Lake area, several snowfields are fronted only by moraines of the older group; moraines of the younger group are not present.

Age

Mazama ash has been identified in the soils of the older moraines at several localities in Glacier National Park (Osborn, 1985; Carrara, 1987; Carrara and McGimsey, 1988). The ash usually forms a light-yellowish-brown (10 *YR* 6/4) to reddish-brown (2.5 *YR* 5/6) silt, 5–10 cm thick at a depth of 10 to 20 cm in small depressions.

In this study the Mazama ash was found in the soils at the following sites: (1) on moraines of the older group fronting the Sperry Glacier, (2) on moraines of the older group fronting the site of the former Red Eagle glacier, (3) immediately downvalley from a moraine of the younger group fronting the Grinnell Glacier, and (4) immediately downvalley from a terminal moraine of the younger group fronting the site of the former Clements glacier (table 3). At the latter two sites, moraines of the older group are not present.

Osborn (1985) also identified Mazama ash on moraines of the older group at several sites in Glacier National Park and adjacent Waterton National Park in Alberta. Those sites in Glacier National Park include (1) immediately downvalley from two small glaciers northwest of Triple Divide Pass, (2) the Iceberg Lake area, (3) the Rose Basin area, and (4) the Ptarmigan Lake area.

Mazama ash, dated at 6,845±50 yr. BP (Bacon, 1983), establishes a minimum age for moraines of the older group. Because organic material has not been found to date these moraines, their actual age has not been determined by radiocarbon dating.

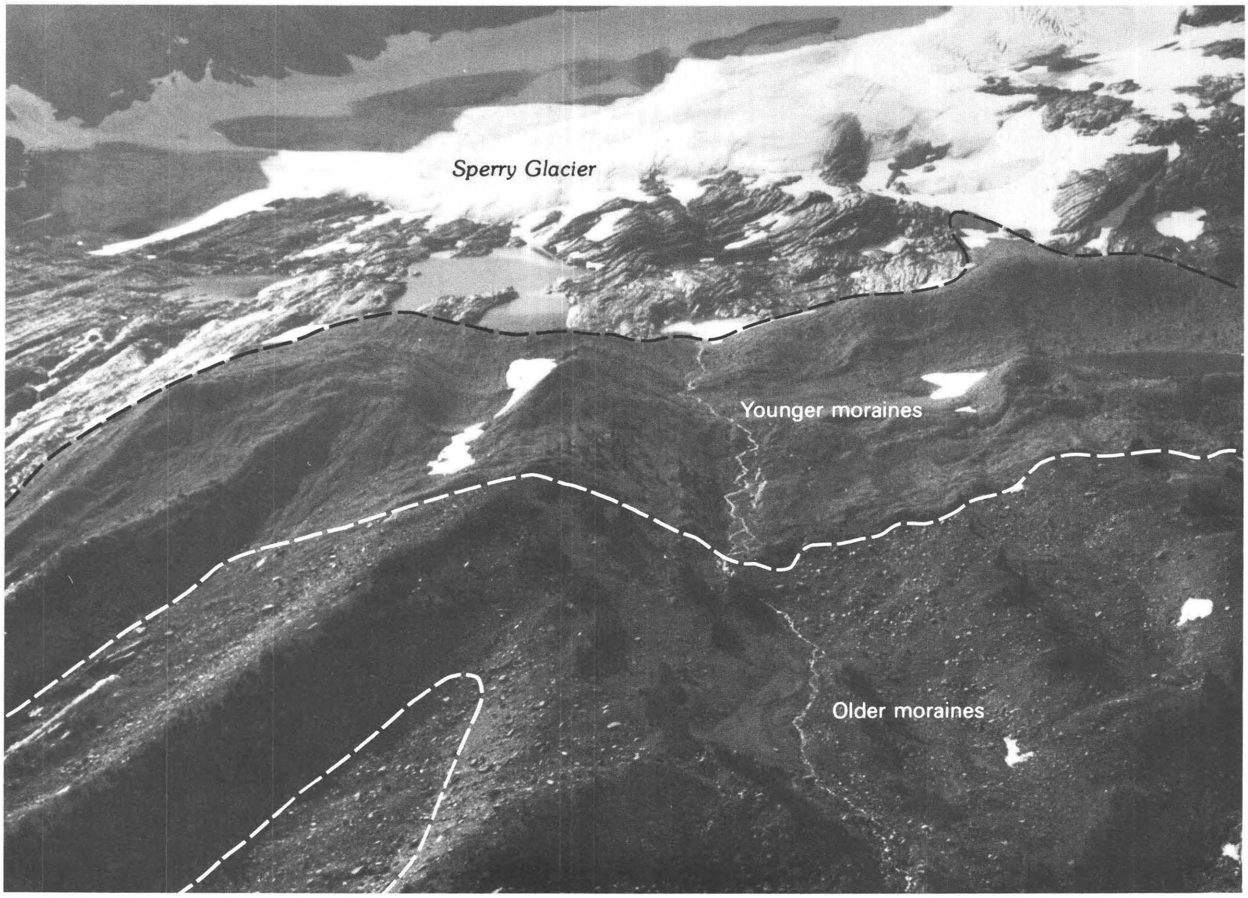


Figure 22. The Sperry Glacier area showing moraines of both the older and younger groups; view is to the south-southeast.

A worldwide early Holocene glacial advance between 8,500 and 7,500 yr. BP has been suggested by Beget (1983). No evidence suggests that moraines of the older group in Glacier National Park date from that period. In fact, evidence elsewhere in the Rocky Mountains suggests that cirque moraines thought to be correlative with moraines of the older group in Glacier National Park are late Pleistocene in age.

In Jasper National Park, Alberta, Crowfoot moraines, also overlain by Mazama ash, are thought to be correlative with moraines of the older group in Glacier National Park (Carrara and McGimsey, 1981; Osborn, 1985; Carrara, 1987). Crowfoot moraines are thought to date from either the period between 9,200 and 8,500 yr. BP or before 9,700 yr. BP (Kearney and Luckman, 1981). The dates of 9,200 to 8,500 yr. BP correspond to a pollen zone in a sediment core from Tonquin Pass, Alberta, thought to represent a cool period. The age of 9,700 yr. BP is the core's basal age. Later studies, including fir:pine pollen ratios and radiocarbon ages of coniferous wood above present-day timberline, indicate that timberline in the Jasper National Park region was at least 100 m higher than present (hence, the climate was correspondingly warmer) by 8,800 yr. BP and generally remained higher until about 5,200 yr. BP (Kearney and Luckman, 1983b; Luckman and Kearney, 1986). During this warm period, glaciers in this region were smaller than present. Wood fragments recovered from the snout of the Athabasca Glacier, Alberta, yielded radiocarbon ages of 8,230±80 (Beta–17373) and 8,000±90 (Beta–20047), indicating that at these times, this glacier was less extensive than at present (Luckman, 1988a). From this evidence, it can be inferred that Crowfoot moraines must predate this warm period and therefore are at least older than 8,800 yr. BP.

In the Wind River Range of Wyoming, a radiocarbon age of 11,770±710 yr. BP (GX–11,772) is thought to closely date the type Temple Lake moraines (Zielinski and Davis, 1986). This age was obtained from Rapid Lake, 1 km downvalley from the Temple Lake moraines, and dates the abrupt transition from clastic to organic sediment. This transition was interpreted as dating the retreat of the former Temple Lake glacier upvalley and deposition of clastic sediment in the basin now occupied by Miller Lake (Zielinski and Davis, 1986). A subsequent radiocarbon age of 11,400±630 yr. BP

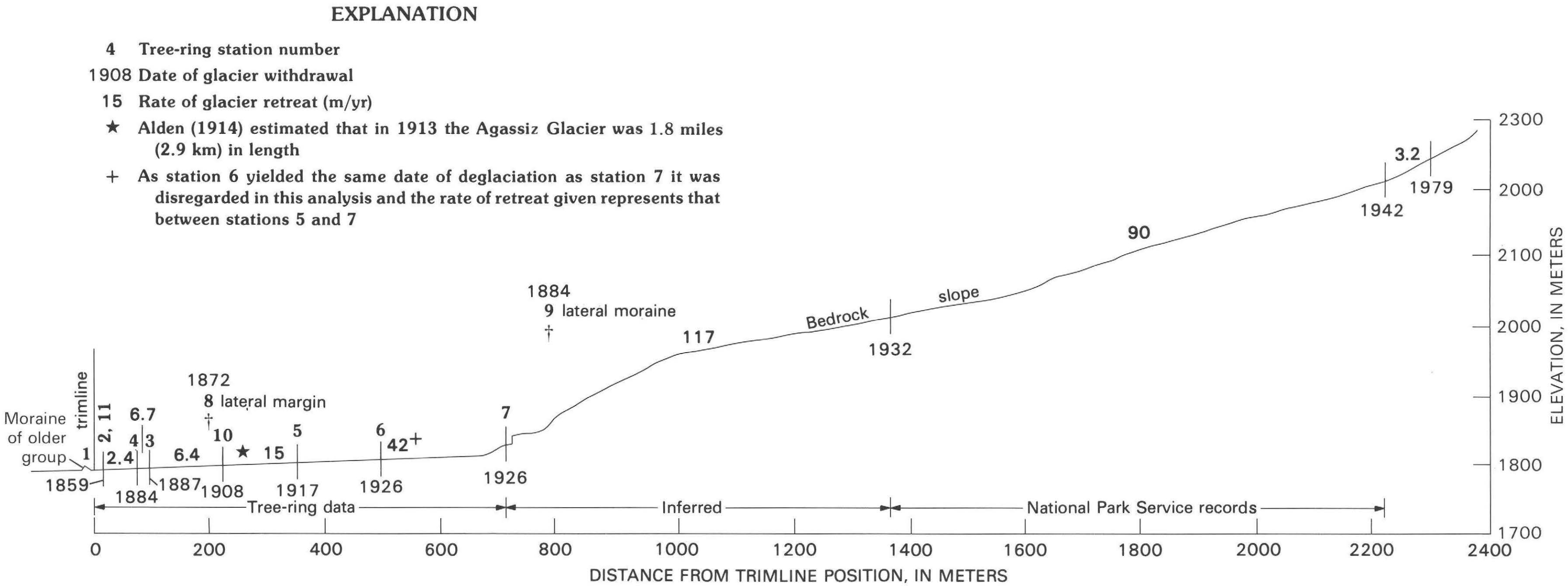


Figure 23. Profile along the central flow axis of the mid-19th century Agassiz Glacier. Profile shows dates of deglaciation, and rates of glacial retreat as determined by tree-ring studies and historical records (Carrara and McGimsey, 1981). Stations 8 and 9 were located above the valley floor on the western margin of the forest trimline and western lateral moraine respectively, associated with the maximum mid-19th century Agassiz Glacier. These stations have been projected over the flow axis, hence their height above the profile (shown by a dagger) represents the ice thickness of the Agassiz Glacier at that time.

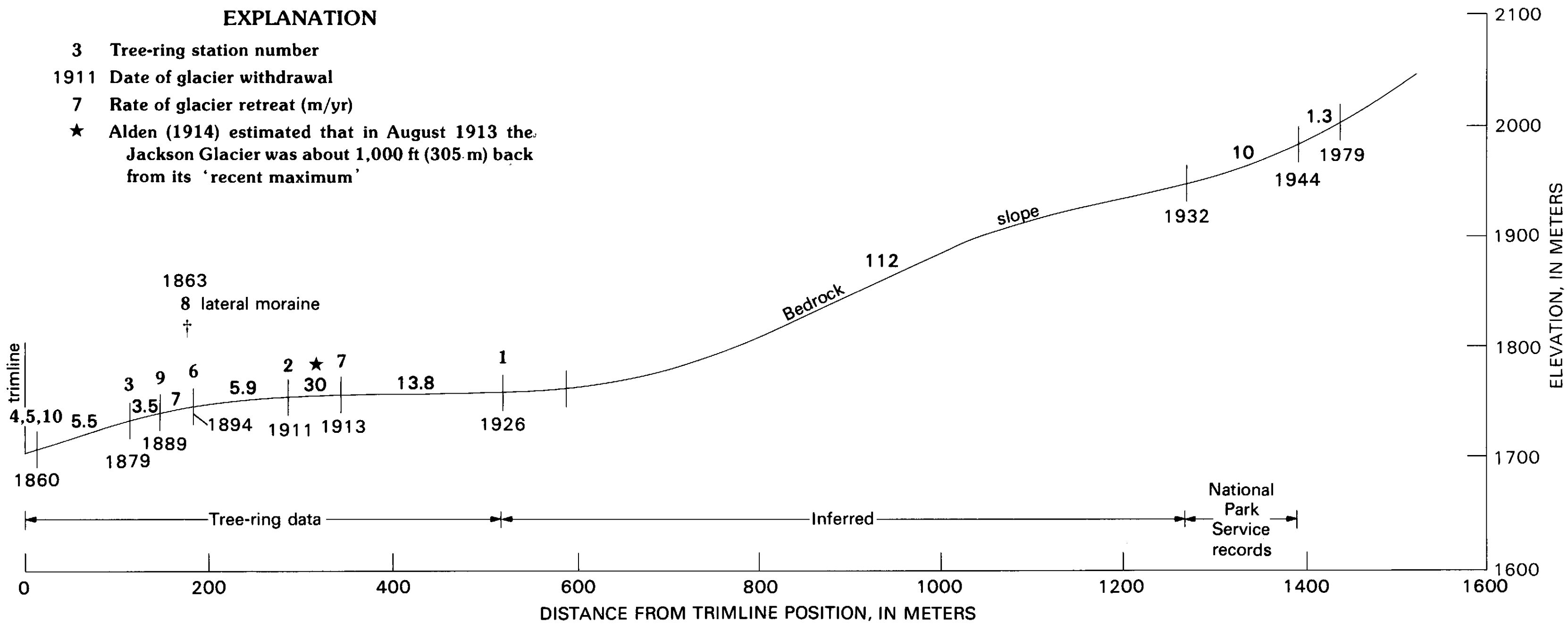


Figure 24. Profile along the central flow axis of the mid-19th century Jackson Glacier. Profile shows dates of deglaciation, and rates of glacial retreat as determined by tree-ring studies and historical records (Carrara and McGimsey, 1981). Station 8 was located on the western lateral moraine, deposited by the Jackson Glacier during its mid-19th century maximum position. This station has been projected over the flow axis, hence its height above the profile (shown by a dagger) represents the ice thickness of the Jackson Glacier at that time.

(GX–12,719) was obtained from a core from Temple Lake, that is inside the inner Temple Lake moraine. Hence, this latter date is a minimum age for the type Temple Lake moraines (Zielinski and Davis, 1987).

In Colorado, minimum ages are also available for cirque moraines similar to the older moraines in Glacier National Park. In the Front Range, a radiocarbon age of 9,915±165 yr. BP provides a minimum age for the Satanta Peak moraine, which is thought to have been deposited between 12,000 and 10,000 yr. BP (Benedict, 1973, 1985). In the San Juan Mountains, spruce:pine pollen ratios and 22 radiocarbon age determinations of coniferous wood fragments from above present-day treeline indicate that by at least 9,600 yr. BP treeline in this mountain range was higher than present and that the climate from 9,600 to 3,000 yr. BP was generally warmer than present (Carrara and others, 1984). Hence, the Yankee Boy and Grenadier cirque moraines in this mountain range must predate 9,600 yr. BP.

As previously discussed, the Glacier Peak G ash has been identified at 10 sites in the Glacier National Park region. Although this ash has not been found in the higher areas of the park, the lack of it on moraines of the older group is difficult to interpret. These moraines may postdate this ash, which dates from about 11,200 yr. BP (Mehringer and others, 1984). However, if the older moraines were already formed by the time this ash was erupted, the likelihood of its preservation would be minimal on a recently deglaciated landscape with sparse vegetation subject to paraglacial processes (Jackson and others, 1982). Furthermore, in the Glacier National Park region, the Glacier Peak G ash is thin compared to the commonly found Mazama ash. The Mazama ash, which has been identified in more than 25 bogs in the Glacier National Park region, is usually 5–20 cm thick; and in some bogs it is more than 50 cm thick. The Glacier Peak G ash is usually only 1–2 cm thick. Hence, the absence of the Glacier Peak G ash on moraines of the older group cannot be taken as conclusive proof that these moraines postdate 11,200 yr. BP.

In summary, although the moraines of the older group are known to predate the Mazama ash (6,845 yr. BP), their actual age is presently unknown. These moraines probably do not represent the early Holocene advance (8,500 to 7,500 yr. BP) suggested by Beget (1983). Moreover, Davis (1984) and Davis and Osborn (1987) have questioned many of the arguments used by Beget to support this proposed early Holocene advance. Moraines of the older group in Glacier National Park are thought to be correlative with other high-altitude cirque moraines elsewhere in the Rocky Mountains. These moraines were probably deposited in the latest Pleistocene. After Osborn and Davis (1985) and Davis and Osborn (1987) reviewed the published radiocarbon ages and paleoclimatic records, they concluded that many of the older, well-vegetated cirque moraines in the American and southern Canadian Cordillera were produced during a roughly synchronous advance around or before 10,000 yr. BP.

Whether moraines of the older group are the product of a separate advance after the end of the late Wisconsin glaciation or are simply the product of the last advance or stillstand of late Wisconsin glaciers before final deglaciation is not known. Because Crowfoot moraines in Banff and Jasper National Parks, Alberta, which are thought to be correlative with the older moraines in Glacier National Park, are found in virtually the same sites as Cavell moraines, which date from the last several centuries, the Crowfoot moraines are thought to represent a separate advance after the end of the late Wisconsin glaciation (Luckman and Osborn, 1979). However, because late Wisconsin glaciers would probably have retreated to those same cirques and well-shaded niches where present-day glaciers and snowfields lie, it seems just as possible that moraines of the older group could simply have been formed during a stillstand or minor advance of late Wisconsin glaciers before final deglaciation.

Moraines of the older group in Glacier National Park are thought to correlate with other cirque moraines elsewhere in the Rocky Mountains: (1) Crowfoot moraines in the middle Canadian Rockies (Luckman and Osborn, 1979); (2) Temple Lake moraines in the Wind River Range of Wyoming (Currey, 1974; Zielinski and Davis, 1986, 1987); (3) Satanta Peak moraines in the Front Range of Colorado (Benedict, 1973, 1985); and (4) Yankee Boy moraines in the San Juan Mountains of Colorado (Carrara and others, 1984).

Glaciers in Glacier National Park during the Early to Middle Holocene Warm Period

During the early to middle Holocene warm period there were probably even fewer glaciers in the park than at present. Several of the larger glaciers in Glacier National Park may have survived this warm period, although many of the park's glaciers probably did not exist at this time. Oxygen-isotope ratios determined by B.H. Luckman on fir fragments from Jasper National Park suggest that the mean annual temperature at 5,900 yr. BP was about 1.9 °C warmer than present (Osborn, 1982). Although a similar amount of warming would certainly eliminate many of the smaller, marginal glaciers in Glacier National Park, the larger, higher glaciers such as the Blackfoot, Harrison, Sperry, Rainbow, Grinnell, and Jackson probably survived the early to middle Holocene warm period. It is interesting to note that as early as 1942, Matthes stated that most of the present-day glaciers in the park were not the shrunken remnants of late Wisconsin glaciers but were new glaciers that probably formed when the climate cooled after the middle Holocene (Matthes, 1942).

Moraines of the Younger Group

Location and Description

Moraines of the younger group have been identified at more than 150 sites throughout Glacier National Park and are commonly the only moraines fronting the many present-day glaciers and snowfields (Osborn, 1985; Carrara, 1987; Carrara and McGimsey, 1988). These moraines lack soils or ash layers, and are generally devoid of vegetation, although pioneering vegetation including scattered krummholz conifers is locally present. They are typified by sharp-crested rubbly moraines, are as much as 50 m high, and slope at the angle of repose (fig. 22). These moraines may contain ice cores.

Some glaciers did not form terminal moraines. During the mid-19th century the Agassiz and Jackson Glaciers extended about 2 km from their cirque headwalls into forested areas. Although the maximum positions of these glaciers are well marked by forest trimlines that merge with fresh, prominent lateral moraines of the younger group, terminal moraines were not formed (Carrara and McGimsey, 1981).

Age

Most of the moraines of the younger group in Glacier National Park are thought to date from advances that culminated in the mid-19th century. Tree-ring studies within the trimlines fronting the Agassiz and Jackson Glaciers (the only glaciers in the park that advanced into well-forested areas during the last several centuries) indicate that retreat began about A.D. 1860 (figs. 23 and 24) (Carrara and McGimsey, 1981). Hence, the glacial advances probably occurred shortly before this time during the mid-19th century. In support of this interpretation, several trees cored just beyond the outermost moraine of the younger group fronting the Sperry Glacier contained very narrow tree rings for the years between A.D. 1840 and 1850, suggesting that this glacier was near its maximum extent during that decade. In addition, Luckman (1988b) dated the maximum Little Ice Age advance of the Athabasca and Dome Glaciers in Jasper National Park, Alberta, at 1843/44 and 1846 respectively.

Some of the moraines of the younger group date from advances other than the mid-19th century. Historical records indicate that several small recessional moraines fronting various ice bodies in the park were deposited in the early 20th century (Carrara and McGimsey, 1988). For instance, two small moraines east of Comeau Pass were deposited in the early 20th century by a western remnant of the Sperry Glacier that no longer exists. Richmond (1986b), who studied glacial deposits in Glacier National Park and the adjacent Blackfoot Indian Reservation during the late 1950's and early 1960's, recognized two ages of moraines within the younger group, both of which he referred to as the Mount Jackson moraines. He thought that the greater amount of vegetation and stability of boulders on the outer of the two moraines present in some cirques suggested that some of them may be "earlier late neoglacial" in age.

In the middle Canadian Rockies, Cavell moraines have been deposited in front of many of the present-day glaciers and snowfields. These moraines have been dated between the 16th and early 20th centuries by dendrochronology, lichenometry, and historical records (Luckman and Osborn, 1979).

In the Wind River Range of Wyoming, moraines in front of present-day glaciers were deposited during the Gannett Peak stade. These bouldery, unvegetated moraines are thought to have been deposited by advances between the 16th and 19th centuries, with the advance of A.D. 1850 being in general the most extensive (Richmond, 1965).

In the Front Range of Colorado, bouldery, unvegetated, ice-cored moraines in front of present-day glaciers and snowfields were deposited by the Arapahoe Peak advance. Lichen measurements suggest that the oldest of these moraines may have been deposited about A.D. 1650, whereas historic records indicate that the youngest of these moraines were deposited by the beginning of the 20th century (Benedict, 1968).

In summary, moraines of the younger group in Glacier National Park are thought to correlate with other "Little Ice Age" moraines elsewhere in the Rocky Mountains: (1) Cavell moraines of the middle Canadian Rockies (Luckman and Osborn, 1979); (2) Gannett Peak moraines in the Wind River Range of Wyoming (Richmond, 1965); and (3) Arapahoe Peak moraines in the Front Range of Colorado (Benedict, 1968, 1973, 1981, 1985).

Climatic Implications of the Younger Moraines

At most sites (>80 percent) in Glacier National Park, only a single moraine of the younger group fronts the present-day glaciers and snowfields. Moraines of the older group are rarely present. Thus, at most sites the climatic cooling of the mid-19th century produced glaciers that probably overran and destroyed moraines of the older group. Hence, the climatic cooling of the mid-19th century, although mild compared to full-glacial conditions, was the most severe in the Glacier National Park region since the end of the late Wisconsin glaciation.

The severity of the mid-19th century climate and the corresponding glacial advance are also indicated by the presence of Mazama ash in the soil immediately downvalley from moraines of the younger group fronting the Grinnell Glacier and the site of the former Clements glacier. Any earlier more extensive glacial advances would have scraped the ash away from these sites; therefore, the presence of this ash indicates that the mid-19th century advances of these two glaciers were their most extensive since the Mazama ash was deposited 6,845 yr. ago.

Similar conclusions regarding the mid-19th century glacial expansion have been reached in the Canadian Rockies. In Jasper National Park, the Cavell advance was generally more extensive than the older Crowfoot advance (Luckman and Osborn, 1979; Osborn, 1982). Based on the relative extent of the Cavell and Crowfoot moraines, Kearney and Luckman (1981) concluded that the most favorable climatic conditions for Holocene glacier development occurred in the last 500 years. In addition, fir:pine pollen ratios from Watchtower basin in Jasper National Park indicate that Holocene timberline reached its lowest altitude (hence the climate was the coolest) in the last 500 years (Kearney and Luckman, 1983).

GLACIAL FLUCTUATIONS SINCE THE MID-19TH CENTURY IN GLACIER NATIONAL PARK

Since their mid-19th century culmination, glaciers in Glacier National Park have shrunk drastically (Alden, 1914; Dyson, 1940, 1948, 1952; Johnson, 1980; Carrara and McGimsey, 1981, 1988; and Carrara, 1987). Field and airphoto inspection of moraines of the younger group indicate that during the mid-19th century there were more than 150 glaciers in Glacier National Park. These glaciers ranged in size from small ice bodies of only a few hectares to the Blackfoot Glacier, which covered about 7.6 km^2 and encompassed the area of the present-day Blackfoot and Jackson Glaciers (fig. 25). During that time the Mount Jackson area of Glacier National Park contained 27 glaciers totaling about 21.6 km^2 (Carrara and McGimsey, 1988).

Observations and photographs from between A.D. 1900 and 1914 indicate that although many glaciers in Glacier National Park had downwasted since the mid-19th century, most still extended to or near to their terminal moraines (Alden, 1914). On the basis of tree-ring studies and historical records, Carrara and McGimsey (1981) concluded that the retreat rates of the Agassiz and Jackson Glaciers were generally modest before about A.D. 1920, but after that time both glaciers began to retreat rapidly (figs. 23, 24). This period of rapid retreat continued until the mid-1940's. This conclusion is supported by climatic records from Kalispell, Mont., that indicate that above average summer temperatures and below average annual precipitation occurred from about A.D. 1918 to 1943 (Carrara and McGimsey, 1981) (fig. 26). Glacier retreat rates slowed markedly after this period (Dyson, 1952; Johnson, 1980; Carrara and McGimsey, 1981).

During the warm, dry period from about A.D. 1918 to 1943, many glaciers in Glacier National Park disappeared. Of the more than 150 glaciers in the park during the mid-19th century, only 50 to 70 remained by 1950 (Dyson, 1952). Only 53 glaciers, comprising 13.8 km^2, remained in 1958 (Meier, 1961).

In 1979, 17 of the 27 glaciers present in the Mount Jackson area during the mid-19th century no longer existed. The remaining glaciers comprised a total area of only about 7.4 km^2, compared with the 21.6 km^2 contained in the 27 glaciers in this area during the mid-19th century (Carrara and McGimsey, 1988).

Most of the 17 former glaciers in the Mount Jackson area were either small or unfavorably oriented with respect to solar radiation. The two largest of these former glaciers, the Logan and Red Eagle glaciers, faced due east and were 0.93 and 0.5 km^2 in size, respectively, during the mid-19th century (fig. 27). These two glaciers probably disappeared by A.D. 1950 (Dyson, 1952). In August 1983, the site of the former Red Eagle glacier was occupied by a freshly striated bedrock surface covered with a veneer of bouldery rubble. A small snowfield lay against the headwall, in several places pierced by bedrock (fig. 28).

The most drastic shrinkage of an ice body during this century in Glacier National Park is that of the Agassiz Glacier, in the northwest corner of the park in the Livingston Range. This glacier occupied a shallow cirque between Kintla Peak (3,080 m) and Mount Peabody (2,808 m) (fig. 29). During the mid-19th century the Agassiz Glacier was one of the largest ice bodies in the park. At that time, it flowed northeast down a bedrock dip slope for 2.4 km before spilling over a small cliff, then advanced across relatively flat forested terrain for another 0.7 km. This glacier had a surface area of 3.38 km^2 and a terminus altitude of 1,793 m (Carrara and McGimsey, 1981).

Retreat of the Agassiz Glacier from its maximum mid-19th century downvalley extent, as shown by its associated forest trimline, began about A.D. 1859

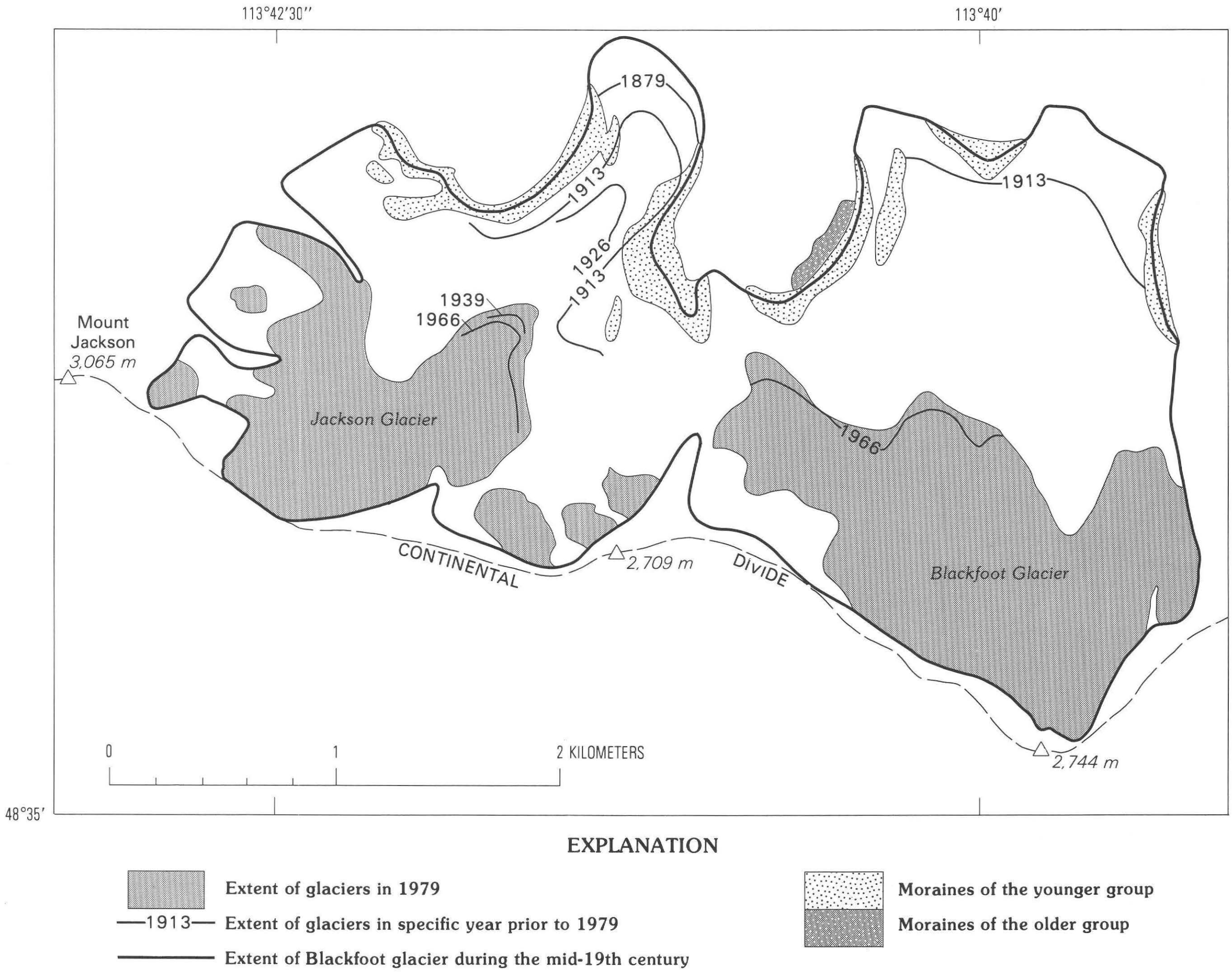


Figure 25. The mid-19th century Blackfoot Glacier showing the disintegration of this glacier into the present-day Blackfoot and Jackson Glaciers. Moraines of the older and younger groups are also shown (modified from Carrara and McGimsey, 1988).

Carrara and McGimsey, 1981). Eleven tree-ring stations within the forest trimline, from the maximum downvalley extent to the base of the bedrock dip slope, indicate that the glacier retreat rates were modest (2.4 to 15 m/yr) until 1917, after which they became quite rapid (>40 m/yr) (fig. 23). By the mid-to-late 1920's this glacier had retreated back onto the bedrock dip slope.

During the next several decades retreat was rapid. From 1932 to 1942 retreat of the Agassiz Glacier was measured by the National Park Service (Matthes, 1944). Retreat was greatly accelerated because the bedrock dip slope under the glacier at that time was 15°. Matthes (1944) stated that recession was caused not only by melting but also by the avalanching of ice down the dip slope. During the period from 1932 to 1942, the Agassiz Glacier retreated nearly 0.9 km. In 1939, Dyson (1941) reported that the Agassiz Glacier had almost separated into two parts, one along the eastern flank of Kintla Peak, the other along the northwest side of Mount Peabody. By 1940, this separation was complete, and although the eastern ice body had a distinct tongue, this tongue disappeared shortly after 1941. At that time Dyson (1941) estimated that these two remnants had a combined surface area of 0.75 km^2.

In 1979, patches of blue ice and projections of bedrock were observed in the western ice body, which had a surface area of 0.75 km^2, but only firn was observed on the eastern ice body, and it is doubtful that either is still a glacier. Volume shrinkage has been drastic due to the tremendous lowering of the ice surface. Field inspection and map interpretation revealed that in places now devoid of ice, the Agassiz Glacier had been 200 m thick. In 1979, ice in the western ice body did not exceed 10 or 15 m in thickness. This western ice body represents less than 3 percent of the volume of the Agassiz Glacier at the time of its culmination in the mid-19th century.

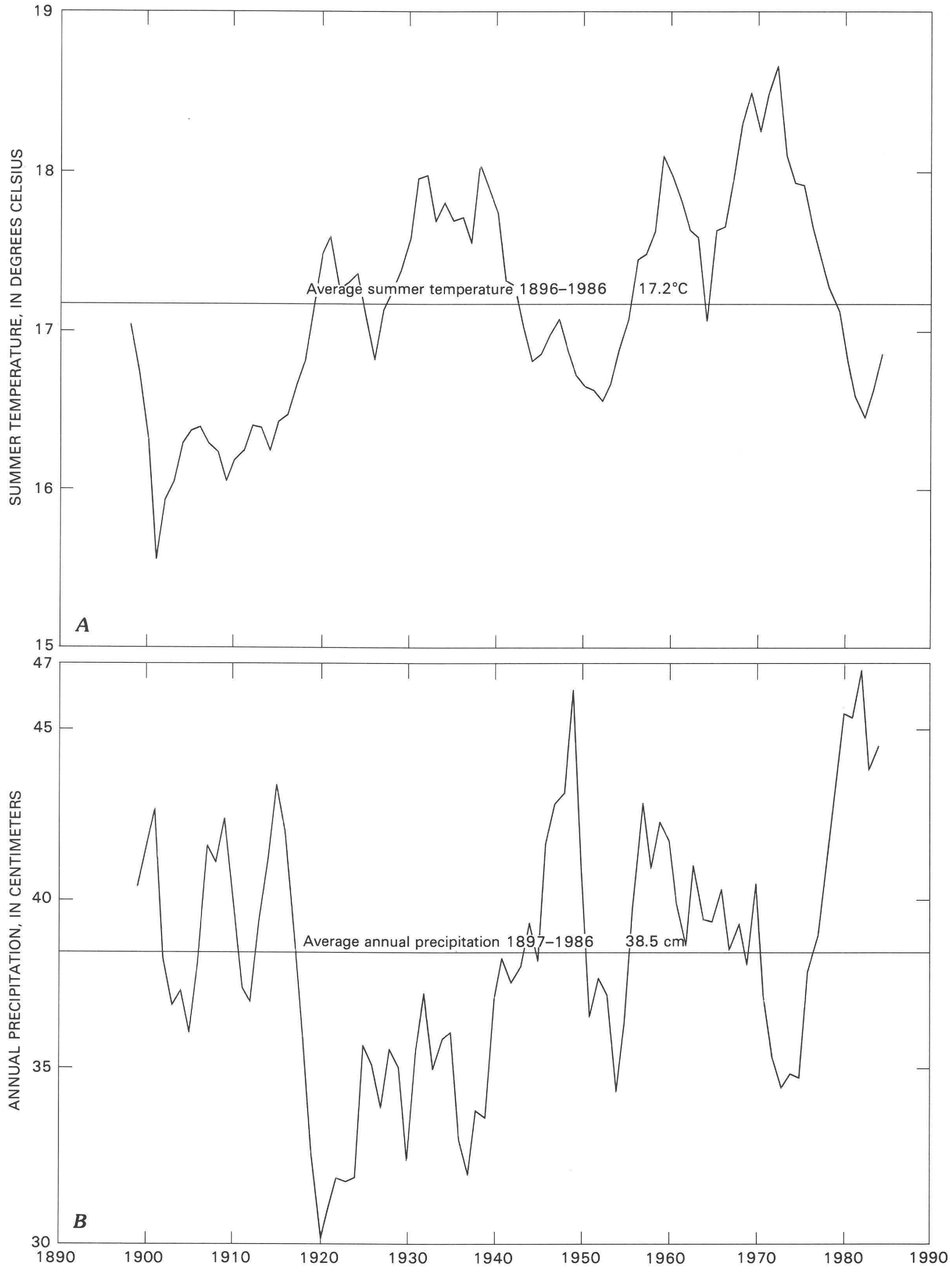


Figure 26. Climatic data, Kalispell, Mont. *A*, Average summer (June, July, August) temperatures; 5-year running mean, 1897–1984. *B*, Average yearly precipitation; 5-year running mean, 1898–1984.

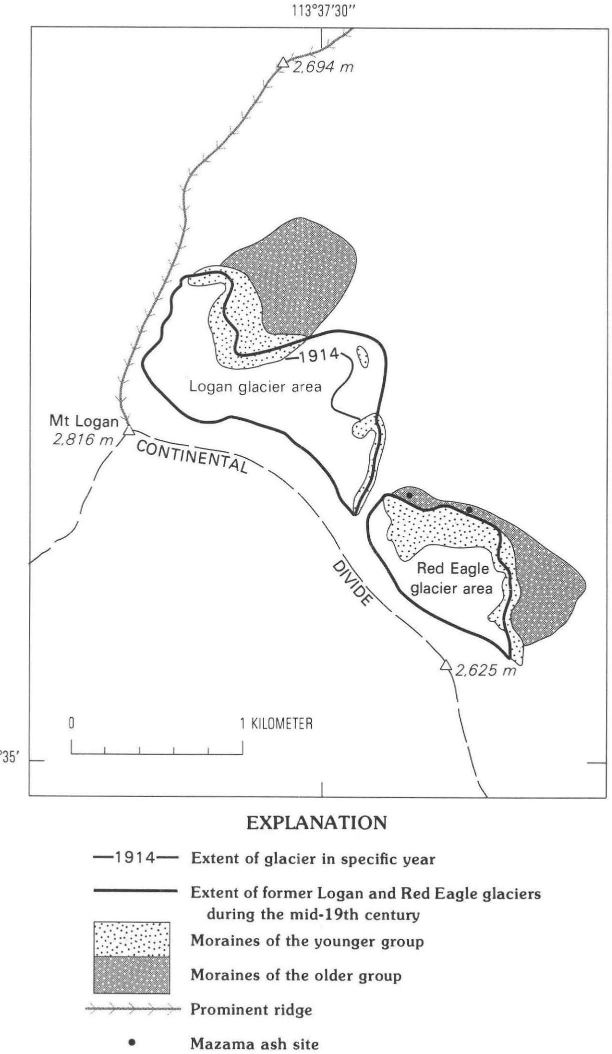


Figure 27. The former Logan and Red Eagle glacier areas, showing moraines of both the older and younger groups and Mazama ash sites (modified from Carrara and McGimsey, 1988).

The Sperry Glacier provides another well-documented example of an ice body that has retreated drastically from its mid-19th century position (Johnson, 1980; Carrara, 1987; Carrara and McGimsey, 1988). During the mid-19th century, this glacier was about 3.87 km^2 in area (fig. 30). By 1927, it had separated into one large and two smaller glaciers (Dyson, 1948). The two smaller glaciers lay against the northeast face of Edwards Mountain and the area southeast of Comeau Pass, and they probably disappeared sometime during the warm, dry period between 1920 and the mid-1940's. In 1979, the Sperry Glacier consisted of a single ice body of about 1.06 km^2, lying against the northern flank of Gunsight Mountain.

Other glaciers have divided into several smaller ice bodies. For example, the Blackfoot Glacier, in the Lewis Range along the Continental Divide between Mount Jackson (3,065 m) and Mount Logan (2,817 m), had two ice lobes and covered about 7.6 km^2 during the mid-19th century. By 1979, this glacier was divided into seven smaller glaciers with a total area of about 3 km^2 (fig. 25). Again, most of this retreat occurred during the warm, dry period between about 1918 and 1943.

During the mid-19th century the western ice lobe of the Blackfoot Glacier (the present-day Jackson Glacier) extended 2.2 km down its steep (15°) bedrock dip slope and another 0.6 km across a relatively flat forested terrain to an altitude of 1,695 m. Tree-ring studies at the downvalley limit of the forest trimline fronting the Jackson Glacier indicate that retreat began about A.D. 1860 (figs. 24, 31). Ten tree-ring stations within the forest trimline indicate that the rate of retreat was modest (<7 m/yr) until about 1911, after which the rate increased. Like the Agassiz Glacier, the Jackson Glacier retreated to its steep bedrock dip slope in the mid- to late-1920's; at that time retreat became rapid. By 1939, the Jackson Glacier had separated from the Blackfoot Glacier by 0.8 km and its terminus was at an altitude of 1,965 m (Dyson, 1941).

Between 1966 and 1979, several of the larger glaciers in the Mount Jackson area advanced slightly (Carrara and McGimsey, 1988). A comparison of 1966 and 1979 airphotos revealed that the ice tongues of the Harrison and Jackson Glaciers (fig. 25) advanced about 100 m during that time. The Blackfoot Glacier also advanced as much as 100 m along its broad western margin at that time, but its eastern margin remained stationary. However, the Sperry Glacier continued a pattern of overall retreat (fig. 30). Climatic records from Kalispell, Mont., suggest that these small advances correspond to an increase in annual precipitation and a decrease in summer temperatures beginning about 1975 (fig. 26).

In 1979, the largest ice body in Glacier National Park, the Blackfoot Glacier, had an area of about 1.74 km^2. Other relatively large glaciers in the park include the Harrison (1.49 km^2), Sperry (1.06 km^2), Rainbow (1.06 km^2), Grinnell (1.03 km^2), and Jackson (1.02 km^2) (Carrara and McGimsey, 1988).

CONCLUSIONS

Deglaciation

This study indicates that large areas of the Glacier National Park region were deglaciated before the deposition of the Glacier Peak G and Mount St. Helens Jy ashes 11,200 and 11,400 yr. BP. Any remaining glaciers at that time were confined to local mountain valleys.

Figure 28. The former Red Eagle glacier area, showing moraines of both the older and younger groups.

Deglaciation was extensive before 11,400 yr. BP in the Marias Pass area. The presence of the Glacier Peak G and Mount St. Helens Jy ashes at the Marias Pass exposure clearly indicates deglaciation of this area before 11,400 yr. BP. As ice flowing east through Marias Pass supplied much of the ice to the Two Medicine piedmont glacier, the disappearance of this glacier from the plains of Montana probably predates the deglaciation of the Marias Pass area by at least several hundred, if not several thousand years.

Extensive retreat of the St. Mary glacier before 11,200 yr. BP is indicated by the presence of the Glacier Peak G ash at the Chewing Blackbones bog. During late Wisconsin time this glacier flowed at least 50 km from its source area along the Continental Divide in the Logan Pass area north into southwestern Alberta. The presence of the Glacier Peak G ash at the Chewing Blackbones bog indicates that by 11,200 yr. BP the St. Mary glacier had retreated at least 20 km from its maximum late Wisconsin position. Because the Glacier Peak G ash is underlain by 110 cm of silty clay at this site, the age of the ash is thought to represent a very minimum date of deglaciation for this area.

On the west side of the park, the valleys of the North and Middle Forks Flathead River, which had contained large trunk glaciers, were extensively, if not completely deglaciated by the time the Glacier Peak G and Mount St. Helens Jy ashes were deposited. The presence of these ashes at the Cabin Creek exposure indicates that by 11,400 yr. BP, the trunk glacier that had filled the valley of the North Fork Flathead River had retreated upvalley from the Cabin Creek exposure. Because at this site the Mount St. Helens Jy ash is underlain by 340 cm of fine-grained sediment that may have required several hundred or thousand years to accumulate, the age of the ash is probably a very minimum date of deglaciation.

Deglaciation of the valley of the Middle Fork Flathead River by 11,200 yr. BP is indicated by the presence of the Glacier Peak G ash at the Middle Fork bog and Goat Lick exposure in the southwestern corner of Glacier National Park. At the Middle Fork bog, this ash is underlain by 50 cm of fine-grained sediment. At the Goat Lick exposure, the Glacier Peak G ash is underlain by about 30 m of gravel. Hence, the stratigraphy at both sites suggests that they may have been

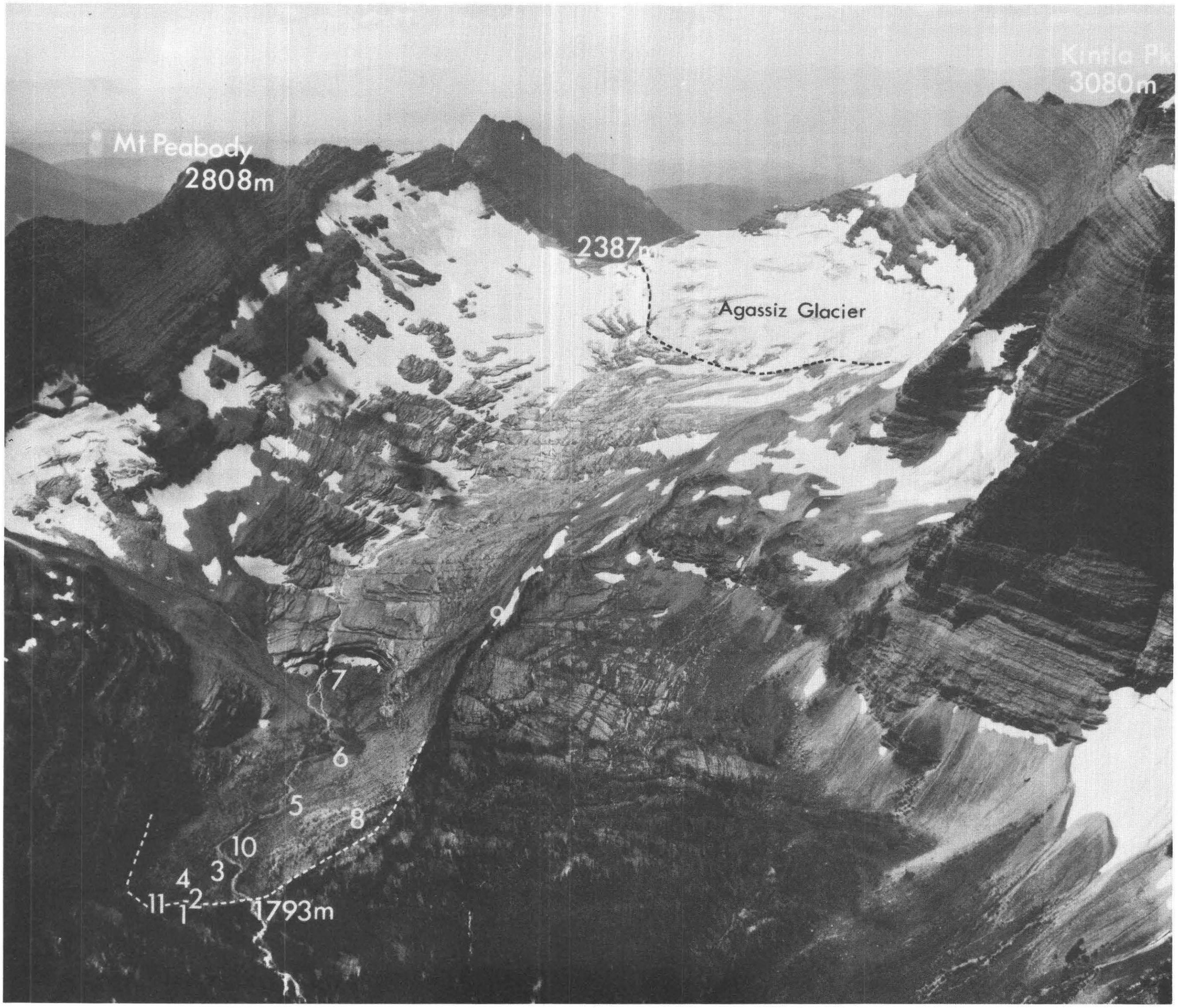


Figure 29. The Agassiz Glacier area, showing forest trimline marking the mid-19th century extent of this glacier and remnant snowfield on eastern flank of Kintla Peak. Numbers within forest trimline represent those tree-ring study sites shown on figure 23 (modified from Carrara and McGimsey, 1981). View is towards the southwest. Photograph by Austin Post, U.S. Geological Survey, Aug. 7, 1961.

deglaciated several hundred or possibly several thousand years before the deposition of the Glacier Peak G ash. Therefore, the large trunk glacier that filled the valley of the Middle Fork Flathead River with hundreds of meters of ice probably had retreated far upvalley into the Lewis and Clark Range by the time the Glacier Peak G ash was deposited.

The former Bowman and McDonald glaciers, which were tributary to the large trunk glaciers in the valleys of the North and Middle Forks Flathead River, were confined solely to their respective valleys before 11,200 yr. BP. This is indicated by the radiocarbon age of 11,150±90 yr. BP from Bowman Lake bog and by the presence of the Glacier Peak G ash at Ben Ryan bog in the Lake McDonald area. Other tributary glaciers along the western flank of the Livingston Range and eastern flank of the Whitefish Range probably were also confined to their valleys by that time. How much farther upvalley from the Bowman Lake and Ben Ryan bogs the Bowman and McDonald glaciers actually lay is unknown. However, data elsewhere in the Rocky Mountains suggest that by as early as about 11,500 yr. BP remaining glaciers may have been as small as subsequent late Holocene glaciers (Porter and others, 1983).

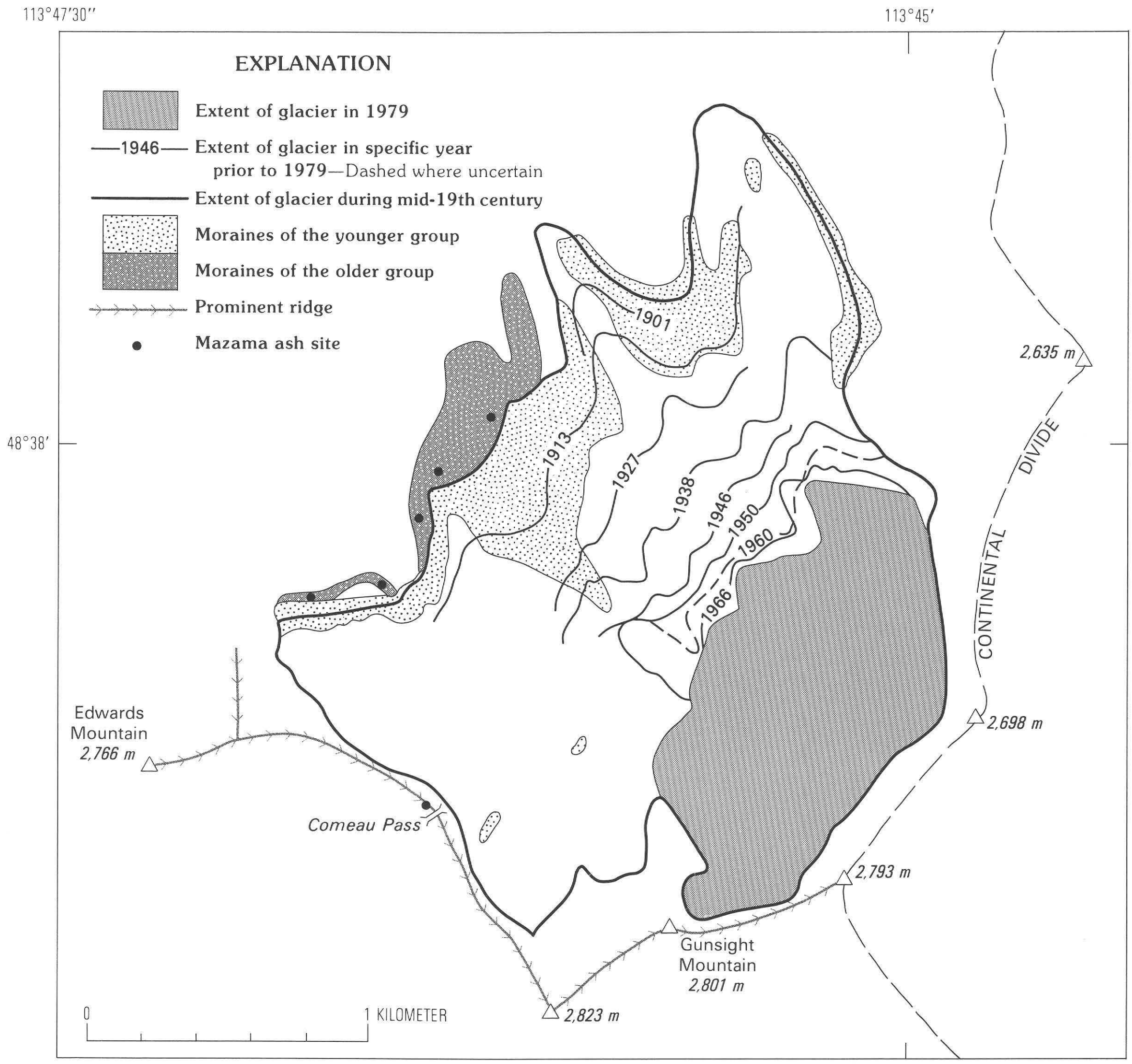


Figure 30. The Sperry Glacier area, showing shrinkage of this glacier from the mid-19th century to 1979, moraines of the older and younger groups, and Mazama ash sites (modified from Carrara and McGimsey, 1988).

The Kootenai Lakes area was also probably deglaciated before 11,000 yr. BP. Although neither the Glacier Peak G nor Mount St. Helens Jy ashes were found in this area, stratigraphic evidence suggests that the time of deglaciation may be older than the radiocarbon ages obtained from the Mount Cleveland bog by several thousand years. This suggestion is based on the fact that pine fragments from hole #1, dated at 9,510±350 yr. BP, were underlain by 250 cm of sediment. The radiocarbon age of 9,850±350 yr. BP on willow fragments from hole #4 is an average radiocarbon age of the lower 50 cm of organic material from this bog. The lower several centimeters of organic material from this hole are probably much older than this date. Hence, the radiocarbon ages from the Mount Cleveland bog are thought to be very minimum dates of deglaciation for the Kootenai Lakes area.

By 10,000 yr. BP, glaciers in Glacier National Park were probably confined to the same cirques and well-shaded niches that shelter the present-day glaciers and snowfields. Radiocarbon ages from sites near present-day glaciers in Banff and Jasper National Parks, Alberta, and Yoho National Park, British Columbia, indicate that by 10,000 yr. BP late Wisconsin glaciers had receded to positions close to those of present-day glaciers (Luckman and Osborn, 1979; Reasoner and Rutter, 1988). Although evidence of that limited glacier extent by 10,000 yr. BP was not found in the Glacier National Park region, a similar amount of deglaciation by that time is suggested.

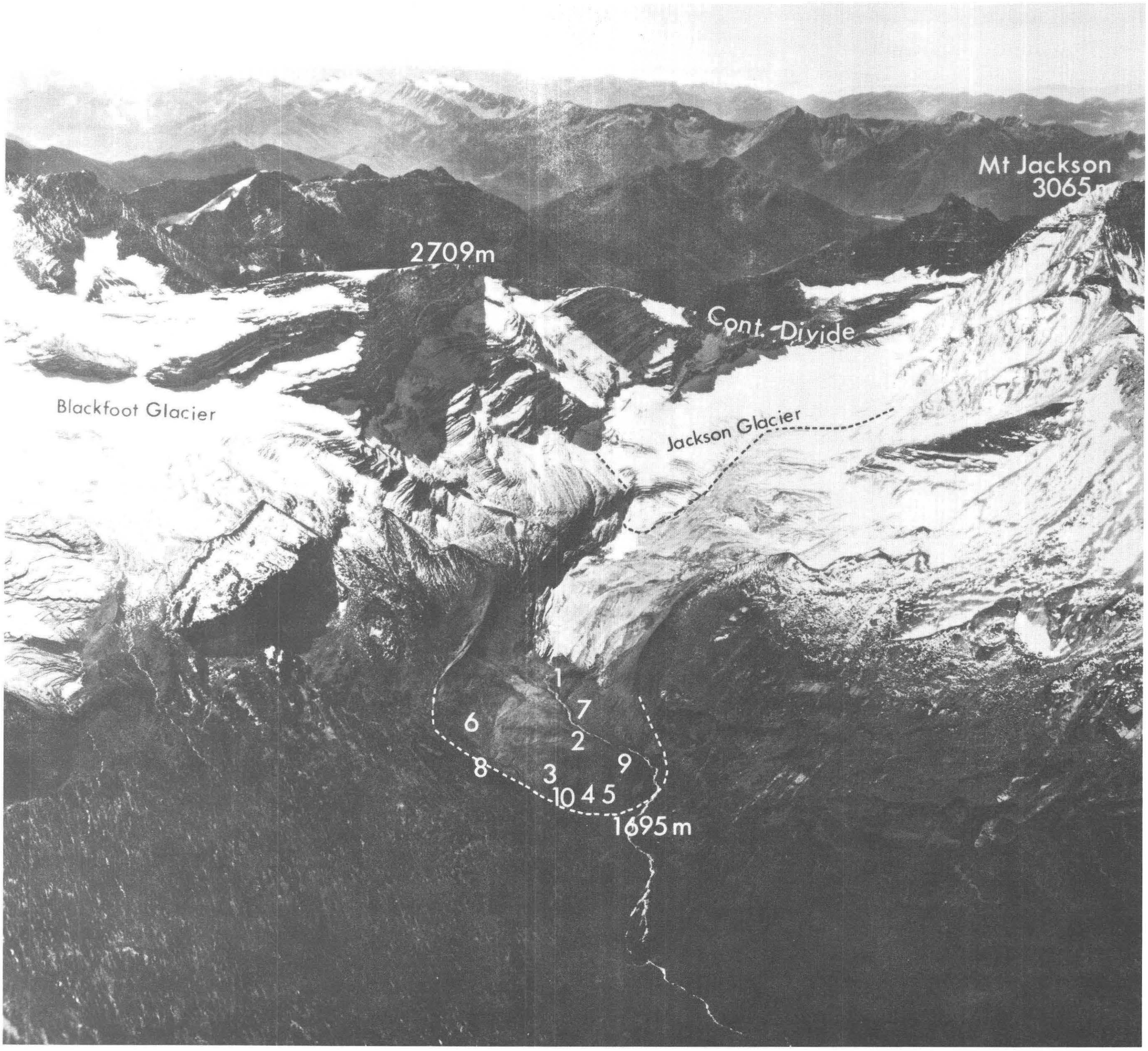


Figure 31. The Jackson Glacier area, showing forest trimline marking the mid-19th century extent of this glacier and remaining glacier. View is toward the south-southwest. Numbers within forest trimline represent those tree-ring study sites shown on figure 24 (modified from Carrara and McGimsey, 1981). Photograph by Austin Post, U.S. Geological Survey, Sept. 8, 1969.

Revegetation

Timberline in the northern Rocky Mountains may have been about 1,000 m lower than at present, during the Wisconsin glaciation (Baker, 1983). Conifers expanded their range rapidly throughout the Rocky Mountains between 11,000 and 9,000 yr. BP (Barnosky and others, 1987). The distribution of conifers during the late Wisconsin is not known with certainty, but they may have survived in small groups along the mountain slopes (Barnosky and others, 1987). With deglaciation, these small groups could have rapidly expanded their range.

Revegetation of deglaciated areas by trees and shrubs was underway by the time the Glacier Peak G and Mount St. Helens Jy ashes were deposited. This is indicated by the macrofossils and pollen found in association with these ashes at the Marias Pass exposure. A conifer needle and alder strobilus found there immediately below the Glacier Peak G ash, and several small willow fragments found below the Mount St.

Helens Jy ash indicate that revegetation of this area was underway between 11,200 and 11,400 yr. BP. From these macrofossils and an analysis of pollen found in the sediment at this exposure, Carrara and others (1986) concluded that at the time the Glacier Peak G and Mount St. Helens Jy ashes were deposited, vegetation in the Marias Pass area consisted of an open landscape of shrubs, herbs, and scattered conifers. At that time there was a larger proportion of spruce and whitebark pine than at present. Because such an assemblage is characteristic of the tundra/subalpine forest boundary (treeline) (presently at an altitude of about 2,050 m), whereas the Marias Pass exposure is at an altitude of 1,550 m, these data may be interpreted as indicating a treeline about 500 m lower than present 11,400 years ago. However, the fossil insect data suggest that a 500-m lowering of treeline may be a minimum estimate. The lack of any tree-related fossil insect species in the sediment immediately below the Mount St. Helens Jy ash led Elias (1988) to suggest that treeline may have been below the Marias Pass exposure by an additional several hundred meters. Therefore, at about 11,400 yr. BP treeline was at least 500 m lower than present and may have been as much as 700 m lower. At that time, the area around Guardipee Lake, on the plains about 60 km northeast of Marias Pass, probably was a grassland (Barnosky and others, 1987).

Evidence of extensive reforestation was obtained from the Lake McDonald area. Radiocarbon ages of wood or peat containing wood fragments of 9,920±120 yr. BP, 9,910±100 yr. BP, and 8,730±80 yr. BP from Glacier Route 7, East McGee, and McGee bogs, respectively, clearly indicate reforestation of the McGee Meadow area by lodgepole pine by about 10,000 yr. BP. In addition, the stratigraphy of the Glacier Route 7 bog suggests that the reforestation of this area by lodgepole pine may have occurred soon after the deposition of the Glacier Peak G ash. The radiocarbon age of 10,630±250 yr. BP of a wood fragment from the Howe Lake bog is a minimum date of reforestation by either spruce or larch in the Lake McDonald area. In addition, the age of 9,650±250 yr. BP of gyttja containing coniferous fragments at the Snyder Ridge bog indicates reforestation of the area surrounding that relatively high altitude (1,351 m) bog by that time. Hence, by as early as 11,000 yr. BP, and certainly by 10,000 yr. BP, the Lake McDonald area supported a forest containing lodgepole pine and probably Engelmann spruce. This forest was probably well established throughout the valleys of the North and Middle Forks Flathead River.

A similar time of reforestation is indicated for the Kootenai Lakes area of Waterton Valley by the radiocarbon ages from the Mount Cleveland bog. A radiocarbon age of 9,510±350 yr. BP from hole #1 is a minimum date for the return of lodgepole pine to this area. The radiocarbon age of 9,850±260 yr. BP from hole #4 is a minimum date for the return of willow. As the major augering effort was concentrated in the valley of the North Fork Flathead River and only a short trip was made into the Waterton Valley, there is no reason to believe that the time of reforestation in Waterton Valley was much different than that of the better documented North Fork. Hence, a forest containing lodgepole pine was probably also well established in the Kootenai Lakes area by 10,000 yr. BP.

Moraines of the Older Group

Mazama ash, which was found to overlie several moraines of the older group in Glacier National Park, provides a minimum age of 6,845 yr. BP for these moraines. This older group of moraines probably do not date from the early Holocene advance (8,500 to 7,500 yr. BP) suggested by Beget (1983). Vegetative evidence as well as correlation with other moraines elsewhere in the Rocky Mountains suggests that these moraines are late Pleistocene in age. Whether these moraines are the product of a separate post-Wisconsin glacial advance or are simply the product of the last advance or stillstand of late Wisconsin glaciers before final deglaciation is not known.

Moraines of the Younger Group

Most of the moraines of the younger group in Glacier National Park were deposited by advances that culminated in the mid-19th century. This is indicated by tree-ring studies in the forest trimlines fronting the Agassiz and Jackson Glaciers (Carrara and McGimsey, 1981) and in the area just beyond the outermost moraine of the younger group fronting the Sperry Glacier. Historical records indicate that several small recessional moraines fronting various ice bodies in the park were deposited in the early 20th century. Richmond (1986b) thought that some moraines of the younger group in Glacier National Park might be "early late neoglacial" in age. Cavell moraines in the middle Canadian Rockies, which are correlative with moraines of the younger group in the park (Carrara and McGimsey, 1981; Osborn, 1985; Carrara, 1987), have been dated between the 16th and early 20th centuries (Luckman and Osborn, 1979).

Most of the present-day glaciers and snowfields in Glacier National Park are fronted only by a single moraine of the younger group. Moraines of the older group are rarely present. Hence, at most sites, the climatic cooling that culminated in the mid-19th century produced glaciers that probably overran and destroyed moraines of the older group. Although mild compared to full-glacial conditions, the climatic cooling of the mid-19th century was the most severe in the Glacier National Park region since the end of the late Wisconsin glaciation.

Glacial Fluctuations since the Mid-19th Century

Glaciers in Glacier National Park have shrunk drastically since the mid-19th century when the park contained more than 150 glaciers. From the mid-19th century until about 1920, glacier retreat rates were slow. However, from about 1920 to the mid-1940's, the Glacier National Park region had above average summer temperatures and below average annual precipitation. During this time the glaciers retreated drastically or disappeared altogether. After the mid-1940's, the overall retreat rates slowed. Between 1966 and 1979, several of the larger glaciers in the Mount Jackson area advanced slightly (Carrara and McGimsey, 1988). By 1979, more than half the glaciers present in Glacier National Park during the mid-19th century no longer existed.

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APPENDIX A

APPENDIX A. PREPARATION AND ANALYSIS OF ASHES BY X-RAY FLUORESCENCE

Collecting and Processing of Ash Samples

Most of the ash samples discussed in this paper were obtained with a bucket auger from bogs, at depths ranging from 0.3 to 4 m. Some samples of Mazama ash were collected from soils. Hence, in many cases, these ash samples were heavily contaminated with detrital sand, silt, clay, and organic material. Several relatively clean ash samples were collected directly from exposures, although these samples also contained small amounts of detrital material. Because of this contamination, all samples were pretreated before microscopic examination and chemical analysis.

Ash samples collected in this study commonly contained about 80–90 percent silt and clay. Because the silt- and clay-size fraction of the ash samples could not be separated from the contaminating detrital silt and clay, the entire silt- and clay-size fraction was discarded leaving only the sand-size fraction. Hence, all analyses were performed on the sand-size fraction, which commonly comprised only about 10–20 percent of the ash samples. Therefore, when collecting ash samples, a relatively large amount (300–500 g) was needed to assure that after elimination of the silt- and clay-size fraction, enough would be left for analyses.

In order to obtain a relatively clean sample of the glass fraction, the ash samples were initially put in a weak solution of dispersing agent (commercial laboratory detergent) and agitated for about 5 minutes daily with an ultrasonic probe to disperse any clay and organic material present in the sample. After several days, organic material would rise to the surface of the liquid and could then be decanted off. The samples were then placed on a 270-mesh sieve and washed with water to remove the silt and clay. If the sample was heavily contaminated with silt and clay, it was again put into a weak dispersing solution for several days and the procedure was repeated. Next, the samples were centrifuged in heavy liquid (specific gravity of 2.42) to separate out the glass fraction (specific gravity <2.42). Although the >2.42 specific gravity fraction contained some detrital rock fragments, it also contained the various accessory minerals used to help identify the ash; hence, it was also saved. Finally, the glass fraction was passed through a magnetic separator to eliminate those glass shards that contained mafic crystals. The remaining glass fraction, at least 0.8 g, was then analyzed for major elements by wavelength-dispersive X-ray fluorescence. A complete explanation of the principles of wavelength-dispersive X-ray fluorescence spectrometry, sample preparation, instrumentation, and precision is given in Taggart and others (1981, 1987).

Comparison of X-ray Fluorescence Data with Electron-microprobe Data

It has been suggested that major-element values determined on bulk samples may be affected by contamination and weathering (Lemke and others, 1975; Smith and others, 1977). Because of these concerns, the use of the electron microprobe has been favored to obtain chemical data from the various ash fractions commonly analyzed. However, Cormie and Nelson (1983) used energy-dispersive X-ray fluorescence on glass separates to obtain trace-element data that allowed them to distinguish between the Mazama, Bridge River, and Mount St. Helens Yn ashes. Results from this study indicate that most major-element values on glass separates determined by wavelength-dispersive X-ray fluorescence for the Mazama, Glacier Peak B and G, and Mount St. Helens Jy ashes are comparable to values of other investigators using electron microprobes.

Mazama Ash

Major-element values of Mazama ash determined by X-ray fluorescence in this study are compared to values determined by electron microprobe by different investigators (table 7). As can be seen, values obtained by X-ray fluorescence for samples of Mazama ash from both bogs and soils in the Glacier National Park region are comparable to those obtained by electron microprobe. In fact, except for the SiO_2 values, which are lower than the other SiO_2 values listed in table 7 by less than 1 percent, all other major-element values determined by X-ray fluorescence are intermediate to values in table 7 that were determined by electron microprobe. This comparability is also demonstrated by the K:Fe ratios. Ratios determined for 11 samples of Mazama ash from bogs in the Glacier National Park region (table 2) averaged 1.46:1, whereas ratios determined for 5 samples of Mazama ash from soil samples (table 3) averaged 1.49:1. Both ratios are intermediate between those derived from the major-element values of Westgate and Gorton (1981), 1.37:1; Sarna-Wojcicki and others (1983), 1.55:1; and Lemke and others (1975), 1.61:1, determined by electron microprobe. In fact, table 7 suggests that individual preparation and operation procedures may be more significant than the type of equipment used to obtain the major-element values for the Mazama ash.

Table 7. Major-element values of glass in the Mazama ash, as reported by various investigators

Reference--	S and others[1]	W and G[2]	L and others[3]	C[4]	C[5]
SiO_2	73.6	72.6	73.0	72.8	72.7
Al_2O_3	14.4	14.4	15.7	14.6	14.6
[6]FeO	1.98	2.11	1.85	2.01	1.95
MgO	.37	.56	.46	.54	.48
CaO	1.56	1.66	1.91	1.84	1.89
Na_2O	4.70	5.18	4.17	4.99	5.17
K_2O	2.87	2.70	2.79	2.75	2.73
TiO_2	.44	.48	.38	.43	.41
Total--	99.92	99.69	100.26	99.96	99.93
K:Fe	1.55:1	1.37:1	1.61:1	1.46:1	1.49:1

[1]Modified from Sarna-Wojcicki and others (1983, table 5-2); values determined by electron microprobe.

[2]From Westgate and Gorton (1981, table 2); values determined by electron microprobe.

[3]From Lemke and others (1975, table 2, sample 60-7-29R); values determined by electron microprobe.

[4]Carrara, this study, average of values in table 2; values determined by X-ray fluorescence.

[5]Carrara, this study, average of values in table 3; values determined by X-ray fluorescence.

[6]Total iron as FeO.

Glacier Peak B and G Ashes

Major-element values determined on samples of Glacier Peak B and G ashes collected at the Trinity Mine site in Washington (Porter, 1978) are presented in tables 8 and 9. This site is the type locality for ashes erupted from Glacier Peak about 11,200 yr. BP. These tables compare major-element values determined by electron microprobe by various investigators to major-element values determined by X-ray fluorescence in this study.

Results indicate that several of the major-element values determined by X-ray fluorescence differ significantly from values determined for those same elements by electron microprobe. For the Glacier Peak B ash, values obtained by X-ray fluorescence were significantly lower than values obtained by electron microprobe for SiO_2, K_2O, and TiO_2, whereas significantly higher values were obtained for Al_2O_3, MgO, and CaO (table 8). For the Glacier Peak G ash, values determined by X-ray fluorescence were significantly lower than those determined by electron microprobe for SiO_2, whereas significantly higher values were determined for MgO and CaO (table 9). Although the FeO value of 1.09 determined by X-ray fluorescence for the Glacier Peak G ash is the highest value on table 9, it is close to the values of 1.04 and 1.07 determined by electron microprobe by Westgate and Evans (1978) and

Table 8. Major-element values of glass in the Glacier Peak B ash from the Trinity Mine, Washington, as reported by various investigators

References-	W and E[1]	S and others[2]	M and others[3]	C[4]
SiO_2	77.2	77.8	77.1	74.3
Al_2O_3	12.6	13.1	13.2	15.0
[5]FeO	1.14	1.12	1.21	1.20
MgO	.38	.26	.30	.43
CaO	1.37	1.33	1.38	2.30
Na_2O	3.97	3.00	3.26	3.92
K_2O	2.91	3.08	2.96	2.69
TiO_2	.23	.27	.23	.20
Total--	99.80	99.96	99.64	100.04
K:Fe	2.72:1	2.93:1	2.61:1	2.39:1

[1]From Westgate and Evans (1978, table 2, sample UA 147); values determined by electron microprobe.

[2]Modified from Sarna-Wojcicki and others (1983, table 5-2); values determined by electron microprobe.

[3]Modified from Mehringer and others (1984, table 1); values determined by electron microprobe.

[4]Carrara, this study; values determined by X-ray fluorescence.

[5]Total iron as FeO.

Table 9. Major-element values of glass in the Glacier Peak G ash from the Trinity Mine, Washington, as reported by various investigators

Reference--	L and others[1]	W and E[2]	S and others[3]	M and others[4]	C[5]
SiO_2	77.1	77.2	77.4	77.7	75.7
Al_2O_3	13.7	12.7	12.8	13.1	13.7
[6]FeO	.95	1.04	.93	1.07	1.09
MgO	.29	.33	.20	.27	.40
CaO	1.15	1.17	1.11	1.21	1.69
Na_2O	3.21	3.89	3.81	3.04	3.94
K_2O	3.40	3.36	3.50	3.17	3.27
TiO_2	.17	.22	.21	.19	.19
Total--	99.97	99.91	99.96	99.75	99.98
K:Fe	3.82:1	3.45:1	4.02:1	3.16:1	3.20:1

[1]From Lemke and others (1975, table 2, sample 68W48); values determined by electron microprobe.

[2]From Westgate and Evans (1978, table 2, sample UA 148); values determined by electron microprobe.

[3]Modified from Sarna-Wojcicki and others (1983, table 5-2); values determined by electron microprobe.

[4]Modified from Mehringer and others (1984, table 1); values determined by electron microprobe.

[5]Carrara, this study; values determined by X-ray fluorescence.

[6]Total iron as FeO.

Mehringer and others (1984). In fact, these three values of FeO are significantly different from those values of Sarna-Wojcicki and others (1983) and Lemke and others (1975), 0.93 and 0.95 respectively. Reasons for this difference in FeO values are not readily apparent, but they may result from different preparation procedures. In addition, values reported for Na_2O in both the Glacier Peak B and G ashes vary among the different investigators (tables 8, 9). Those for the Glacier Peak B ash range from 3.00 to 3.97, and those for the Glacier Peak G ash range from 3.04 to 3.94.

The differences in major-element values between those determined by X-ray fluorescence and those determined by electron microprobe for the Glacier Peak B and G ashes from the Trinity Mine site may be due to the proximity of the site to Glacier Peak, and the cleaning procedure used in this study. The Trinity Mine is only about 20 km from Glacier Peak (Porter, 1978). These ash samples contained a large nonglass fraction that the cleaning procedure used in this study may not have adequately removed. In the process of preparing the Glacier Peak B and G samples from the Trinity Mine site for analysis, it was found that the <2.42 specific-gravity fraction comprised only 7.6 percent and 12.2 percent respectively of the sample weight. This same specific-gravity fraction in samples of the Glacier Peak G ash from the Glacier National Park region comprised about 40–45 percent.

With increasing distance from the volcano, the ash became increasingly lighter as the heavier minerals and rock fragments settled out. Hence, about 40–45 percent of samples of the Glacier Peak G ash in Glacier National Park had a specific gravity <2.42. Therefore, major-element values determined by X-ray fluorescence may become increasingly accurate and hence more comparable to values determined by electron microprobe, with increasing sample distance from the volcano.

At Sun River Canyon, Mont., a site containing the Glacier Peak G ash has been sampled and analyzed (table 10) by several investigators (Lemke and others, 1975; Westgate and Evans, 1978; and Carrara and others, 1986). Most major-element values determined by X-ray fluorescence (Carrara and others, 1986) are similar to those values determined by electron microprobe (Lemke and others, 1975; Westgate and Evans, 1978) (table 10). However, the CaO value determined by X-ray fluorescence (1.49) is significantly higher than those CaO values determined by electron microprobe (table 10). Also, the Na_2O value determined by X-ray fluorescence is somewhat low compared to those determined by electron microprobe. Inasmuch as Na_2O values on tables 8 and 9 range widely, this difference is not considered significant. Moreover, some of the electron microprobe data seem suspect. For example, the MgO value (0.22) of Lemke and others (1975) is quite low compared to the other MgO values in table 10. Also, the K_2O value (3.40) of Westgate and Evans (1978) is somewhat low compared to the other K_2O values in table 10. Hence, it appears that as the Glacier Peak G ash becomes more glass-rich with increasing distance from the source, major-element values determined by X-ray fluorescence become more comparable to values determined by electron microprobe, except for CaO values.

CaO values determined for the Glacier Peak B and G, and Mount St. Helens Jy ashes by X-ray fluorescence were significantly higher than values determined by electron microprobe. In addition, CaO values determined for the Glacier Peak G ash by X-ray fluorescence on the nine samples listed in table 4 range rather widely, from 1.53 to 2.41. Many of the glass shards in the Glacier Peak ashes contain small nonmagnetic inclusions. The cleaning procedures used in this study to obtain bulk samples of the glass fraction may not have eliminated all those glass shards containing these inclusions.

Table 10. Major-element values of glass in the Glacier Peak G ash from Sun River Canyon, Montana, as reported by various investigators

Reference--	L and others[1]	W and E[2]	C and others[3]
SiO_2	76.6	77.0	76.6
Al_2O_3	13.2	12.7	13.0
[4]FeO	.99	1.06	1.03
MgO	.22	.39	.38
CaO	1.26	1.18	1.49
Na_2O	3.94	3.94	3.66
K_2O	3.62	3.40	3.63
TiO_2	.18	.22	.19
Total--	100.01	99.89	99.98
K:Fe	3.90:1	3.42:1	3.76:1

[1]From Lemke and others (1975, table 2, sample HS-87); values determined by electron microprobe.

[2]From Westgate and Evans (1978, table 2, sample UA 647); values determined by electron microprobe.

[3]From Carrara and others (1986, table 1); values determined by X-ray fluorescence.

[4]Total iron as FeO.

Comparison of major-element values for Glacier Peak ashes found in the Glacier National Park region (table 4) to values obtained for the Glacier Peak B and G ashes collected from the Trinity Mine site, Washington (tables 8 and 9), indicate that the Glacier Peak ash present in the Glacier National Park region is the G layer. This correlation is based primarily on the FeO and K_2O values and is also reflected in the K:Fe ratios, which are also characteristic of the G ash.

Mount St. Helens J Ashes

The ash of Mount St. Helens set J, at the three sites in the Glacier National Park region, is believed to be the

Table 11. Major-element values of glass in the Mount St. Helens Jb, Jy, and Js ashes (from Westgate and Evans, 1978)

Ash--	Jb[1]	Jy[2]	Js[3]	Jy[4]
SiO_2	72.4	75.7	77.3	75.8
Al_2O_3	14.8	13.8	13.5	13.8
[5]FeO	1.97	1.31	.92	1.31
MgO	.75	.33	.14	.41
CaO	2.02	1.36	.97	1.78
Na_2O	5.40	4.88	4.34	4.48
K_2O	2.27	2.41	2.60	2.34
TiO_2	.33	.20	.20	.19
Total--	99.94	99.99	99.97	100.11
K:Fe	1.23:1	1.96:1	3.02:1	1.91:1

[1]From Westgate and Evans (1978, table 3, sample UA 482); values determined by electron microprobe.

[2]From Westgate and Evans (1978, table 3, sample UA 483); values determined by electron microprobe.

[3]From Westgate and Evans (1978, table 3, sample UA 484); values determined by electron microprobe.

[4]Carrara, this study, average of the Marias Pass and Cabin Gulch samples (table 5); values determined by X-ray fluorescence.

[5]Total iron as FeO.

Jy layer. Westgate and Evans (1978) analyzed the glass fraction of various set J ashes collected by D.R. Mullineaux. Samples UA 482, UA 483, and UA 484 (Westgate and Evans, 1978, table 3) are from layers Jb, Jy, and Js, respectively (D.R. Mullineaux, oral commun., 1984). Comparison of major-element values of glass in these three set J layers (table 11) with values of the ashes from the three sites in the Glacier National Park region (table 5) indicate that the Mount St. Helens ash in the Glacier National Park region is Jy.

Mount St. Helens J ashes from the Marias Pass and Cabin Creek exposures were analyzed by X-ray fluorescence. Values of the eight major elements analyzed for the Marias Pass ash are similar to the Jy ash analyses of Westgate and Evans (1978) except for Na_2O, whose value of 4.40 is closer to that of Js (4.34). Values determined for the Mount St. Helens J ash at the Cabin Creek exposure are also most similar to values given for the Jy ash by Westgate and Evans (1978), although the CaO and K_2O values (1.88 and 2.31) are closer to those of the Jb ash. Some of the major-element values determined for the Mount St. Helens J ash from Guardipee Lake are similar to values given by Westgate and Evans (1978) for the Js ash (SiO_2, Al_2O_3, and Na_2O). However, considering that the Js layer is of limited extent, the ash from Guardipee Lake is also probably the Jy ash. In addition, all three J ashes from the Glacier National Park region have K:Fe ratios (Marias Pass 1.91:1; Cabin Creek 1.91:1; Guardipee Lake 2.14:1) similar to that of the Jy (1.96:1) ash analyzed by Westgate and Evans (1978) (tables 5 and 11).

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