1-1-2006

Geologic Framework and Glaciation of the Central Area

Christopher L. Hill
Boise State University

This document was originally published in Handbook of North American Indians.
During the Late Pleistocene, the Laurentide ice sheet extended over the western interior Plains and Great Lakes region in the central region of North America. This central area generally encompasses the northwestern interior Plains of North America, extending from the Rocky Mountains in the west to the western Great Lakes and Hudson Bay in the east (figs. 1-2). It includes parts of the Mackenzie River, Missouri River, and Mississippi River systems. Deglaciation of this region led to the development of landscapes that were inhabited by Rancholabrean faunal communities including human groups.

Three major ice centers formed in North America: the Labrador, Keewatin, and Cordilleran. The Cordilleran covered the region from the Pacific Ocean to the eastern front of the Rocky Mountains ("Geological Framework and Glaciation of the Western Area," this vol.) while the Labrador and Keewatin ice fields combined to form the Laurentide Ice Sheet. The glacial geology of the central region of North America includes evidence of the lobes and sublobes of the Keewatin ice as well as lobes from the Labrador ice source that expanded from the Lake Superior and Lake Michigan basins. The Keewatin ice covered the western interior plains of North America from the Mackenzie River to the Missouri River and the headwaters region of the Mississippi River. The Labrador ice covered much of eastern Canada and the northeastern part of the United States ("Geological Framework and Glaciation of the Eastern Area," this vol.), overlapping in the western Great Lakes and Mississippi valley with Keewatin ice.

The central glaciated area of North America contains evidence of successive phases of glacial advance and melting of lobes and sublobes of the Laurentide Ice Sheet. Glacial lakes formed where drainages were blocked by the ice sheet. Deglaciated landscapes contain evidence of landscape evolution and changing biotic systems. Biotic evidence, including plant remains, vertebrate fossils, and artifacts also provide an indication of the landscape and its viability for habitation.

The Late Pleistocene geologic framework of the region is the product of research since the 1800s (Chamberlin 1883). Prior to the 1950s lithostratigraphic and geomorphic data were the primary basis for interpretation and correlation. Starting in the 1950s chronometric techniques, chiefly radiocarbon dating, were applied to test the developing framework. This led to the abandonment and reorganization of much of the earlier terminology (Willman and Frye 1970). For example, the Iowan, Tazewell, Cary, and Mankato Wisconsin glaciations were all placed into the Woodfordian. The Woodfordian substage is followed...
by the Twocreekan, based on a buried fossil forest bed in eastern Wisconsin. Prior to the 1970s it was thought that the till overlying the lake sediments that buried the Two Creeks forest was associated with the Valders advance (C.V. Haynes 1964, 1969). It now appears likely that the Valders type locality is older than the Two Creeks forest (Evenson et al. 1976). The till overlying the Two Creeks forest, deposited as part of the Two Rivers advance, represents the beginning of the Greatlakean, starting around 11,600 B.C. Thus, landscapes that can be presumed to be potentially associated with human groups using Clovis and Folsom artifacts (Stanford 1999a; V.T. Holliday 2000) are post-woodfordian and chiefly associated with the early part of the Greatlakean chronozone.

The record of glacial and interstadials reflected in the stratigraphic sequences and physiographic features in the central region of North America can be compared to the climate record in the Greenland ice cores and in northern Europe. The Woodfordian chronozone encompasses the interval of time since the Farmdalian. It includes glacial advances prior to the beginning of the Twocreekan chronozone (Kaiser 1994; W.J. Johnson et al. 1997). The beginning of the time interval of the Two Creeks forest in the western Great Lakes may be correlated with the Older Dryas between the Bolling and Allerod in northwestern Europe. The time range from the Bolling to the beginning of the Younger Dryas, which would include the time of the Two Creeks forest, forms the Last Glacial Interstadial (about 12,700-11,600 B.C.). The Younger Dryas, dated to about 11,000-9500 B.C., is followed by the Pre-Boreal chronozone in Northern Europe. Clovis and Folsom artifacts would be expected to be associated with landscapes developed after the Twocreekan, from about the end of the Last Glacial Interstadial until about the end of the Younger Dryas. In the Central Area the physical landscapes potentially available for Late Pleistocene biotic populations are chiefly the product of surficial processes related to glacial activity. The spatial and temporal distributions of radiocarbon ages can be used to estimate the extent of the Laurentide ice during the Middle Wisconsin (fig. 1), as well as the dynamics of Late Wisconsin ice advances and deglaciation (figs. 2-6) (Dyke et al. 2002; Dyke, Moore, and Robertson 2003; Dyke 2004). The geologic framework of the glaciated central area of North America can be examined using stratigraphic, geomorphic, and chronologic evidence from three regions associated with the northwest, southwest, and south-central margins of the Laurentide Ice Sheet.

Northwest Margin

The northwest part of the Late Pleistocene Laurentide ice sheet covered the present-day Mackenzie River system. This includes parts of the drainages of the Athabasca and Peace rivers, and Lake Athabasca, Great Slave Lake, and Great Bear Lake basins (vol. 6:5-18). The Mackenzie River system presently drains northward into the Beaufort Sea.

The spatial distribution of radiocarbon dates indicates that portions of the Mackenzie drainage system were ice-free during the Middle Wisconsin (Dyke et al. 2002; Harington 2004). The most informative radiocarbon dates are based on measurements of vertebrate fossils and fragments of wood in stratigraphic context. To the north, on the Arctic coastal plain east of the Mackenzie River, wood with an age of about 37,600 B.C. may be an indication of nonglaciated landscapes during the Middle Wisconsin (Harington 2004). West of the Mackenzie River delta, plant remains have been dated to 37,800 B.C. South of the Peel River, wood below the Hungary Creek till may reflect slightly earlier interstadial conditions at 40,200 B.C. South of the Peace River, near Watino, Alberta, wood under Late Wisconsin glacial deposits ranges in age from 44,500 to 30,400 B.C. Woolly mammoth fossils (Mammuthus primigenius) dating to 30,300 B.C. were recovered from near the Peace River at Taylor, British Columbia (Churcher and Wilson 1979; Catto et al. 1996). It has been proposed that, in contrast to other regions of the glaciated area of central North America, the maximum of the last advance of the northwest margin of the Laurentide ice sheet may be Middle Wisconsin, with an age of perhaps 33,400 B.C. (Lemmen, Duk-Rodkin, and Bednarski 1994). Mammoth remains under deposits of glacial Lake Old Crow have ages of 28,100 and 27,700 B.C. (Morlan et al. 1990), suggesting the presence of ice in the region by the early part of the Late Wisconsin.

The northwest margin of the Laurentide ice sheet extended west of the present-day Mackenzie River until about 11,800 B.C. (Duk-Rodkin and Hughes 1991; Duk-Rodkin and Lemmen 2000; Dyke 2004). At the time of maximum advance, continental ice covered the Mackenzie and Peel plains and reached the Mackenzie and Richardson Mountains. The maximum Laurentide advance does not coincide with major glacial activity in the mountains themselves; at the time of the maximum Late Pleistocene continental advance montane glaciers were restricted to icecaps on the Mackenzie Mountains. The Laurentide ice did, however, block the mountain drainages, forming glacial Lake Hughes by damming the Peel River in the Richardson Mountain region, and forming glacial Lake Nahanni in southern Mackenzie Mountains. Lobes of the Laurentide ice sheet also blocked the Peace River drainage west of Lake Athabasca (Mathews 1980).

During the Katherine Creek phase of glaciation (about 24,600-19,700 B.C.) expansion of montane glaciers in the Mackenzie Mountains led to coalescence of Cordilleran and Laurentide ice in several river valleys (Duk-Rodkin and...
Hughes 1995; Duk-Rodkin and Lemmen 2000). There were no glaciers in the Richardson Mountains. Melting of the Laurentide and Cordilleran ice sheets from the Katherine phase margins led to an open region between the glaciers. Peat overlying the Mackenzie lobe dated to 15,700 B.C. indicates the timing of deglaciation. By 14,000 B.C., during the Tutieta Lake phase, the Laurentide margin was still west of the present-day Mackenzie River. Glacial Lake Nahanni was still present in the southern Mackenzie Mountains. By about 12,700 B.C. the Laurentide margin had melted away from the present-day Mackenzie delta region and glacial Lake Travaillant had formed east of Fort McPherson, along the northwest margin of the ice in the Mackenzie River valley. Continued melting led to the formation of glacial Lake Ontaratue, southwest of glacial Lake Travaillant, by about 12,200 B.C. As the continental ice sheet continued to melt, glacial Lake Ontaratue disappeared and glacial Lake Mackenzie developed. The ice position was at the Kelly Lake moraine prior to the development of a delta in the Mackenzie River that contains wood dated to 11,600 B.C.

The early stages of glacial Lake Mackenzie in the Mackenzie valley west of present-day Great Bear Lake are associated with radiocarbon ages of 11,600-11,100 B.C., indicating it was present during the time associated with Clovis artifacts in other regions (Stanford 1999a; V.T. Holliday 2000). This is approximately the time for the final draining of glacial Lake Nahanni in the Mackenzie Mountains (Duk-Rodkin and Lemmen 2000; Duk-Rodkin et al. 2004). By about 11,000 B.C. the ice margin had melted east of the Mackenzie River to the western edge of the Great Bear Lake basin, forming glacial Lake McConnell (Hare Indian River phase), while glacial Lake Mackenzie expanded southward as the ice front melted. To the south, glacial ice still extended west of the present-day Mackenzie River, blocking the drainages and forming glacial Lake Laird (fig. 3), north of Fort Laird. The Laird delta was deposited adjacent to the Laird moraine west of present-day Great Slave Lake around 11,300 B.C. from meltwater along the southern margin of the Mackenzie lobe of the Laurentide ice sheet. The Laird moraine and delta, therefore, may be
parts of a landscape contemporaneous with early Clovis artifacts.

Wood dated to 11,500-11,400 B.C. from delta deposits is associated with the Great Bear basin phase of glacial Lake McConnell. After 11,000 B.C. the Laurentide ice sheet was completely east of the Mackenzie River. Glacial lake McConnell expanded to include most of the Great Bear Lake basin and glacial Lake Mackenzie expanded to include the Mackenzie River valley southward towards the Great Slave Lake basin, which was still covered by the Laurentide ice. By 9500 B.C., the Laurentide ice sheet had melted to east of the Great Bear Lake basin and was along the east side of the Great Slave Lake basin (Bednarski 1999). This led to the expansion of glacial Lake McConnell, which eventually covered the Great Bear Lake, Great Slave, and Athabasca basins (Lemmen, Duk-Rodkin, and Bednarski 1994; D.G. Smith 1995). Several landforms appear to be contemporaneous with the time associated with late Clovis, Folsom, and other late-Pleistocene artifacts (figs. 3-4). The Peace delta formed between 10,800 and 7900 B.C. in the northwest Lake Athabasca basin. The Athabasca delta also formed between 10,800-7,900 B.C. west of the Cree Lake moraine in the southern Lake Athabasca basin; it was partially formed by floodwaters from glacial Lake Agassiz around 9400 B.C. Glacial Lake Agassiz, enclosed on the northeast by the Cree Lake moraine, drained into the Clearwater River valley and into glacial Lake McConnell. The Cree Lake moraine marks the position of the Laurentide Ice Sheet around 9500 B.C.

While the melting ice front was leading to the formation of glacial Lake McConnell, the glacial margin of the Athabascan lobe of the Laurentide Ice Sheet was initially west of present-day Lake Athabasca in northern Alberta. Proglacial lakes formed along the margin of the Laurentide Ice Sheet as it melted eastward. By 11,000 B.C. a lobe of ice extending into the Peace River lowland was surrounded by ice-free uplands to the north and south (fig. 3) (Mathews 1980; Bednarski 1999). A late stage of glacial Lake Peace fronted the lobe toward the west. Charlie Lake Cave, in British Columbia, was covered initially covered by a glacial lake. A deglaciated landscape at the site around 10,700 B.C. may have been contemporaneous with the Clayhurst stage of glacial Lake Peace (J.C. Driver 1988). Laurentide ice also blocked the Athabasca River valley to the south forming glacial Lake McMurray. Melting of the ice front led to the expansion of glacial Lake Peace, eventually forming glacial Lake Athabasca (or Lake Tyrrell) in the Lake Athabasca basin. By about 9500 B.C. the Athabasca lobe was at the Slave moraine, east of Slave River and west of present-day Lake Athabasca.
been correlated with the Cree moraine. Lake McConnell appears to have merged with the lake in the Athabascan basin by 9500 B.C. (Duk-Rodkin and Lemmen 2000). By about 7500 B.C., lowering lake levels led to the separation of glacial Lake McConnell into present-day Lake Athabasca, Great Slave Lake, and Great Bear Lake. If humans were present in the Mackenzie River system during the time when Clovis and Folsom artifacts were being used in other parts of North America, they would have been present in a recently deglaciated landscape associated with proglacial lakes and an eastward melting ice margin (figs. 3-4). Yet no evidence has been found in these landscapes.

**Southwestern Margin**

The southwestern margin of the Laurentide ice sheet advanced across the western interior plains of North America, altering the landscape of the present-day Saskatchewan River system, principally in Alberta and Saskatchewan, as well as the upper and middle Missouri River system in parts of Montana and northwestern North Dakota. The Saskatchewan River system presently drains into the Hudson Bay while the Missouri River is a major tributary to the Mississippi River.

The region that contains the Missouri River and Saskatchewan River basins appears to have been mostly ice-free and inhabited by Rancholabrean faunal communities during parts of the Middle Wisconsin (fig. 2) (Burns 1990, 1996; Dyke et al. 2002). Ages on vertebrate remains provide an estimate for the duration of this nonglacial interval. In the Hand Hills region west of Drumheller, Alberta, bone collagen ages indicate ice-free conditions from 37,300 to 24,900 B.C. associated with a mammoth steppe biome (R.R. Young et al. 1999). At Medicine Hat, Alberta, along the South Saskatchewan River, Middle Wisconsin sediments contain mammoth (*Mammuthus*), ground sloth (*Nothrotherium*), horse (*Equus*), camel (*Camelops*), and helmeted musk-ox (*Stalker* and *Churcher* 1982). Deposits overlying the main bone bed date to 41,100 and 40,700 B.C. at the Galt Island Bluff Section. At Evilsmelling Bluff, the next younger deposit has radiocarbon ages from 31,500 to 27,500 B.C., and it is overlain by two tills (Stalker 1977a; Harington 2004). Floodplain and paludal deposits between the two glacial tills contain toad (*Bufo*), wolf (*Canis*), saber-toothed cat (*Smilodon*), *Mammuthus*, *Equus*, and *Camelops*. Thus, it appears that landscapes capable of supporting a diverse group of animals were present prior to the onset of the last major glacial advance in eastern Alberta.

There is evidence for nonglacial conditions to the east, in Saskatchewan, and to the south, in Montana. Along the Saskatchewan River in the north Saskatchewan area, *Mammuthus*, *Equus*, bison (*Bison*), and *Camelops* fossils have been recovered from the Riddell Member of the Floral Formation (Skwara Woolf 1981; Morlan et al. 2001). The Riddell Member is overlain by an upper till of the Floral Formation and the late Wisconsin Battleford Formation (Christiansen 1968, 1968a). Wood between the Floral and Battleford Formations has been dated to 40,600 B.C. (Christiansen 1971). A paleosol found within exposures of the South Saskatchewan valley in eastern Saskatchewan has been attributed to nonglacial conditions ending about 20,000 years ago (David 1966). South of Alberta and Saskatchewan, in the upper Missouri basin, *Mammuthus* and *Equus* fossils found near Box Creek, Montana, that date to around 36,800-36,000 B.C. appear to reflect nonglacial conditions in a region containing both Illinoian and Wisconsin till (Hill 2004). Remains of *Mammuthus* and *Equus* have been found in gravels near Frazer, Montana, under Woodfordian till (Jensen and Varnes 1964).

Interpretations of the extent and timing of glaciation along the southwest margin of the Laurentide ice sheet include support for a maximum glacial advance into northern Montana (Christiansen 1979; Clayton and Moran 1982; Fullerton and Colton 1986) and, in contrast, support for an ice position marked by the Lethbridge moraine in southern Alberta (B. Reeves 1973; Stalker 1976, 1977a). Based on luminescence dating of glacial lake deposits near Great Falls, Montana, it appears that at least part of the Laurentide ice sheet (the Havre lobe) impounded the Missouri River during the late Wisconsin (Woodfordian) (Hill and Feathers 2002; Hill 2006). The age of glacial Lake Great Falls may provide an approximate age for the maximum advance of other glacial lobes to the east that impounded the Missouri, Yellowstone, and Little Missouri rivers forming glacial lakes Musselshell, Circle, Jordan, Lambert, Glendive, and Mikkelsen (A.D. Howard 1960; R.B. Colton and D.S. Fullerton 1986).

A lobe of the maximum late Wisconsin Laurentide ice sheet in central Alberta advanced southeastward into west and central Montana and southwestern Saskatchewan, leaving the Cypress Hills and Boundary Plateau unglaciated. As the Laurentide ice crossed the area of the present-day Milk River valley in southern Alberta, it was divided into two lobes by the Sweetgrass Hills, which also remained unglaciated. The western lobe, the Shelby lobe (Mickelson et al. 1983; Fullerton and Colton 1986), advanced southward to the Missouri River, in the vicinity of Great Falls, Montana. The lobe that extended east of the Sweetgrass Hills, the Havre lobe, moved in two directions. The Loma sublobe, advanced across the Missouri River to north of the Highwood Mountains. The Malta sublobe expanded southeastward along the present-day Milk River, bounded to the north by the Boundary Plateau and to the south by the Little Rocky Mountains and the region of the Musselshell River.

The maximum advance of the Laurentide ice sheet blocked the drainages of north- and east-flowing rivers, forming glacial lakes along the margin of the ice (Calhoun 1906; Alden 1932; R.B. Colton, R.W. Lemke, and R.M. Lindvall 1961; R.B. Colton and D.S. Fullerton 1986; Hill and Valppu
The Lethbridge moraine and its equivalents west of the Cypress Hills in Alberta may coincide with the advance of a lobe eastward across southern Saskatchewan. By 15,400 B.C. the ice margin may have been at the Thompson Lake moraine fronted by glacial Lake Kincaid. There may have been some stagnant ice to the west in the Swift Current and Frenchman River drainages (Christiansen and Sauer 1988). As the ice margin melted to the northeast, glacial Lake Gravelbourg was formed adjacent to the Maxwellton moraine. The lobe that expanded across southern Saskatchewan appears to have expanded into northwestern Montana forming the Medicine Lake moraine, which may correlate with the Grenora, Alamo, and Lostwood moraine margins in North Dakota.

In western North Dakota deglaciation from the Late Wisconsin Laurentide margin to the Grenora, Alamo, and Lostwood moraines by 15,400 B.C. resulted in landscapes associated with glacial lakes. When the ice margin dammed the Missouri River, Lake Crows Fly High was formed, in present-day Lake Sakakawea (Clayton, Moran, and Bluemle 1980). Glacial Lake McKenzie formed when an ice lobe blocked the Missouri River farther downstream.

In Alberta, successive proglacial lakes were created by the melting of the lobe from the Lethbridge moraine. Retreat from the Lethbridge moraine allowed water from glacial Lake Taber to drain into glacial Lake Medicine Hat. Continued melting led to glacial lake’s Taber, Tilly, Patricia, and Bassano. Water from glacial lakes Gleichen, Beiseker, and Drumheller drained into the basins. By the time the lobe was north of the Red Deer River, glacial Lake Patricia covered much of the area to the south. Meltwater from the Kincorth moraine formed glacial Lake Bigstick around 14,700-12,400 B.C.

Deglaciation around 11,800 B.C. in Saskatchewan and Manitoba led to the formation of glacial lakes Leduc, Red Willow, Unity, Saskatchewan, and Regina, bounded on the north and east by Laurentide ice. Ice margins were associated with the Condi, Qu’Appelle, Petlura, and Cowan moraines (Klassen 1994, 1989). Meltwater drained into Lake Agassiz through the Assiniboine channel and through Lake Souris (Christiansen 1979). The abandonment of the Qu’Appelle spillway channel occurred before 11,900 B.C., as determined from radiocarbon dated wood in alluvium. North of the Saskatchewan River, the Kyle mammoth, dated to about 11,900 B.C., is associated with glacial lake deposits (M.C. Wilson and J.A. Burns 1999; Morlan et al. 2001; Harington 2004); the landscape inhabited by the mammoth would have been south of glacial Lake Saskatchewan. In one paleogeographic model, by 11,000 B.C. Lake Saskatchewan had merged with Lake Agassiz, and the ice position was at The Pas moraine (Christiansen 1979). This ice margin is correlated with the Beaver River moraine and glacial Meadow Lake, which flowed north-eastward from northwest Saskatchewan into the Athabasca River drainage and Lake McConnell in northern Alberta. By 9500 B.C. melting of the ice position to the Cree moraine had led to the northward expansion of Lake Agassiz.

There is some information on the viability of the Late Pleistocene deglaciated landscapes of Saskatchewan (M.C. Wilson and J.A. Burns 1999; Beaudoin and Oetelaar 2003). The Niska site overlies deposits of glacial Lake Old Wives and has a charcoal radiocarbon age of 10,900 B.C.
on a possible Folsom component similar to assemblages at the Charlie Lake or Vermilion Lake sites in the Canadian Rockies. This is overlain by a Cody component. The Sjovold site overlies glacial and postglacial deposits north of the Missouri Coteau along the South Saskatchewan River, south of Saskatoon, near the outlet channel that drained from Lake Saskatchewan into Lake Agassiz (Dyck and Morlan 1995). South of the Saskatchewan River, the Heron Eden site contains an Agate Basin point possibly dated to about 9900 B.C. North of Saskatchewan River, the Wiseton imperial mammoth (Mammuthus imperator) is dated to about 10,500 B.C. It was found in lake sediments (M.C. Wilson and J.A. Burns 1999; Morlan et al. 2001).

Deglaciation in the vicinity of the Rocky Mountain foothills of Alberta between the Bow and South Saskatchewan rivers, southwest of Calgary, appears to have begun by 17,100 B.C., based on wood overlying a till from Cartwright Lake (Beierle and Smith 1998). Inhabitable landscapes in the foothills region northwest of Lethbridge, Alberta, can be inferred by the age of 11,100 B.C. on a mammoth tooth (L.E. Jackson et al. 1999). The mammoth appears to have been deposited in glaciolacustrine sediments overlying the Laurentide till. Other vertebrate evidence provides an indication of the types of landscapes present after deglaciation within the time interval potentially associated with Clovis artifacts (L.E. Jackson and M. Pawson 1984; Burns 1996; C. Cambell and J.A. Cambell 1997). Helmeted musk-ox (Bootherium), Equus, Bison, and caribou (Rangifer) were present around 11,300 B.C., along with Clovis points that contain horse blood residue (Kooyman et al. 2001). The Cochrane gravels along the Bow River contain Equus, Bison, Rangifer, and cervids with ages of 11,200 and 10,700 B.C. (Churcher 1968). At the Gallelli gravel pit along the Bow River Camelops has been dated to 11,200 B.C. (M.C. Wilson and C.S. Churcher 1978). At the Bindloss gravel pit Columbian mammoth (Mammuthus columbi) has been dated to 10,900 B.C., and the gravels also contained lion (Panthera) and Equus fossils. Fossils of Equus recovered from the Pasheley gravel pit have an age of 10,900 B.C. (Hills and Harington 2003). At Medicine Hat, delta and river channel deposits overlying the uppermost till contain Mammuthus and Equus and predate river terrace deposits with Canis, Mammuthus primigenius, Equus, Camelops, and Bison. At Lindoe Bluff, Bison and Equus have been dated to 11,100 B.C. (L.E. Jackson and M. Pawson 1984). In southeast Alberta, the Milk River Ridge contains Equus dated to 11,100 B.C. (Burns 1996). Thus Alberta supported a diverse late Pleistocene fauna associated with deglaciated landscapes by the time people were using Clovis artifacts elsewhere on the Plains.

Geologic evidence associated with the Woodfordian ice margin and deglaciated landscapes in the Upper Missouri basin in Montana and western North Dakota also reflect Late Pleistocene habitats after the maximum advance of the Laurentide ice sheet. Stratigraphic sequences occasionally contain Late Pleistocene and early Holocene tephras, buried soils, as well as Rancholabrean fauna. For instance the Marias River sequence, in northern Montana, contains late Wisconsin till overlain by a volcanic ash (correlated with the Glacier Peak tephra, and dated by nearby mammal bone to 11,200-11,100 B.C.) and a buried soil, both underlying the Mazama tephra (Hill 2002, 2005, 2006). In the deglaciated region of the Milk River, in eastern Montana, the Hinsdale mammoth was found in gravels and sands of a glacial outwash channel (Hill 2005). In western North Dakota, mammoth (Mammuthus primigenius?) have been found in the region bounded by the Yellowstone and Little Missouri rivers between the Last Glacial Maximum margin and the Missouri River (Haraldson 1952). Farther north, fossils of Columbian mammoth are known from the Garrison Reservoir.

In northern and eastern Montana and western North Dakota, late Wisconsin deposits associated with ice-marginal and deglaciated landscapes (above the Coleharbor Group in North Dakota) have been assigned to the Aggie Brown Member of the Oahe Formation (Clayton, Moran, and Bluemele 1980; Huber and Hill 2003). Buried soils developed in sediments of the Aggie Brown Member are termed the Leonard paleosol. In Montana, between the Missouri and Yellowstone rivers, the Deer Creek mammoth dating to about 12,500 B.C. underlies silts and buried soils (Hill 2003, 2006). South of the Yellowstone River, radiocarbon ages indicate that the silt-paleosol-silt sequences date to the Pleistocene-Holocene transition. Similar contexts are documented in the Little Missouri and Knife River drainage basins in North Dakota (Kuehn 1993; Tipson 2000; Rawling, Fredlund, and Mahan 2003). Artifacts have been found in association with the Leonard paleosol at several localities (Artz 1995). These include stratigraphic sequences associated with eolian-colluvial sediments overlying Wisconsin terraces along the Missouri River, and alluvial deposits. For example, in the Knife River drainage a stratigraphic sequence comprising the Leonard paleosol in alluvium overlying a late Wisconsin channel lag gravel contains Goshen and Hell Gap artifacts. At the Big Black site, which contains a Folsom assemblage, soil humates from the lower section of a paleosol date to 11,100 B.C. (Tipson 2000). Bone from a Folsom component at the Bobtail Wolf site was dated to 9400 B.C., while average ages on humate fractions from the Leonard paleosol range from 10,600 to 7600 B.C. (Root 2000).

**South-Central Margin**

The south-central margin of late Wisconsin Laurentide glaciation is bounded by the middle Missouri River valley in North and South Dakota and extends to the western Great Lakes (figs. 5-6). Some of this area is linked to landscapes associated with glacial Lake Agassiz that formed in the Red River basin and parts of the Mississippi drainage. The region includes the Assiniboine and Souris river basins in the...
Manitoba, the James River basin in South Dakota, the Minnesota River (glacial River Warren) and Saint Croix Rivers in Minnesota, and the Iowa River in northern Iowa. It also includes Lake Superior basin and the west side of the Lake Michigan basin (vol. 13:44-60).

Middle Wisconsin nonglacial conditions are reflected by radiocarbon dated materials under till in this region (fig. 1). East of the Assiniboine River at Zelena, Manitoba, charcoal in a silt between two tills has been dated to 26,800 B.C. (Morlan, McNeely, and Nielsen 2000). In north-central North Dakota and southeastern Saskatchewan wood recovered under two glacial deposits indicates nonglacial Middle Wisconsin landscapes at 31,300 and 30,800 B.C. Along the north flank of Turtle Mountain, in Manitoba, a mammoth tusk has been dated to 38,100 B.C. (Fulton 1995). In northeastern North Dakota and southwestern Minnesota wood buried by glacial deposits has been radiocarbon dated to 23,000 B.C. (Matsch, Rutford, and Tipton 1972; Clayton and Moran 1982). In eastern South Dakota, wood has been dated to 29,000 and 23,000 B.C. Charcoal incorporated in advancing ice and redeposited in younger sediments implies nonglacial conditions at 42,500 and 38,700 B.C. in central Minnesota (H.E. Wright et al. 2004). Wood dated to around 40,200 B.C. found below lake deposits and overlying till in Minnesota appears to represent a forest at the onset of the Late Wisconsin period buried by Rainy Lobe Wadena phase till (Meyer 1997). In eastern Minnesota, redeposited wood has been dated to 43,000 and 35,800 B.C. (Meyer 1989). Charcoal fragments from near the base of the Roxana silt south of the Rost River in southeastern Minnesota date to 40,000 B.C. (Mason, Nator, and Hobbs 1994). In Iowa, wood under glacial till dates to 22,100 B.C. (Ruhe 1969). In west-central Wisconsin there are two wood dates of 33,900 and 31,800 B.C., while in southeastern Wisconsin wood dates range from 35,400 to 31,800 B.C. Thus, there are many examples implying a nonglacial episode in central North America prior to the Woodfordian glaciation.

The Middle Wisconsin loess, the Roxana Silt, overlies Sangamon deposits and is buried by the Late Wisconsin Peoria loess (Muhs and Bettis 2000). Radiocarbon dates from Minnesota and Wisconsin are 39,300 and 29,400 B.C. (Leigh and Knox 1993). There is evidence of at least two glacial advances from about 40,000 to 26,000 B.C. associated with the Tazewell till (Sheldon Creek Formation, Bettis 1997). Wood dates for the Tazewell margin are 28,000 and 22,100 B.C. (Ruhe 1969). These glacial deposits are overlain by the Late Wisconsin Dows Formation or loess of the Peoria Formation (Bettis, Quade, and Kemmis 1996). Loess designated as the Pisgah Formation was deposited about 45,100-27,000 B.C. The Peorian Formation consists of loess in eastern Iowa dated about 23,300-11,800 B.C.; it is overlain by the southern part of the Des Moines Lobe. In southeastern Minnesota the Farmdale soil just below the Peoria loess has been dated to about 26,700 B.C. (Lively et al. 1987).

The maximum advance of the south-central portion of the Laurentide ice sheet during the Late Wisconsin is associated with two source areas and several lobes and sublobes. One lobe of ice extended from the Lake Winnipeg region in Manitoba southward into North and South Dakota. This advance, the Napeolian phase of glaciation, is estimated to date to about 22,000 B.C. and may have extended west of the present-day location of the Missouri River in North Dakota (Clayton, Moran, and Bluemle 1980; Fullerton 1995a; Bluemle, Reid, and Mitchell 2004). Ice also expanded southward from the Red River lowland as the James lobe. At about the same time another lobe, associated with the Tazewell phase position, advanced in a southwestward direction into eastern South Dakota.

Several glacial lobes had important advances that date to about 15,400 B.C. (Hallberg and Kemmis 1986; Richmond, Fullerton, and Christiansen 1991; Fullerton 1995a; Bettis, Quade, and Kemmis 1996). The maximum advance of the Souris lobe, which expanded from the Winnipeg region, is associated with the Long Lake glacial phase. The James Lobe expanded from the Red River valley into South Dakota during the Cat Tail Creek phase. The Des Moines lobe also expanded from the Red River Valley, reaching its maximum extent marked by the Bemis moraine, in Iowa (fig. 2).

The next phase of glaciation for the Des Moines Lobe is marked by the Altamont moraine and dates to about 14,500 B.C. (Ruhe 1969). To the west, the Souris and James ice lobes may have been at about the Zeeland phase location at this time (Fullerton 1995a). There are several younger glacial phases associated with these lobes. The Algona phase of the Des Moines lobe dates from about 13,900-12,500 B.C., and there are some equivalent phases associated with the ice margins of the Souris and James lobes. The deglaciation of the Algona phase (Mankato substage) is constrained by radiocarbon ages on wood of about 11,800 B.C. associated with the New Ulm till in eastern Minnesota (Meyer 1989) and wood in overlying sediments dated to 10,800 B.C. in southern Minnesota (Matsch 1972). The youngest margin associated with the James lobe is the Bowdie phase position dated to about 13,600 B.C., while the youngest margin of the Des Moines lobe is the Big Stone moraine dated to about 11,900 B.C.

The margins of the Laurentide ice sheet continued to recede northward. By 11,800 B.C. the James lobe had disappeared, and the north and east regions of North Dakota were covered by the Leeds lobe. Several phases of glaciation are associated with the Leeds lobe; the position at about 11,600 B.C. was at the Mylo moraine in north-central North Dakota (Fullerton 1995). To the west, the temporally equivalent margin of the Souris lobe is marked by the Martin moraine. To the east, as ice melted back from the Big Stone moraine, there were several phases of glaciation associated with the Red River lobe. The Erskine phase is dated to about 11,800 B.C. while the Edinburg phase marks the approximate position of the Red River lobe in eastern North Dakota and western Minnesota around 11,200 B.C. and thus may be
a landscape setting contemporaneous with humans using Clovis artifacts elsewhere in North America. Beaches of the Lockhart and Emerson phases of Lake Agassiz formed on the Edinburg moraine (K.L. Harris, M.R. Luther, and J. Reid 1986). In Saskatchewan and Manitoba the ice margin of the Lake Manitoba lobe formed the Belleau Brook, Harvey Lake, Griffon, Petlura, and Darlingford moraines at the same time as the Red River lobe was at the Edinburg moraine (Teller and Fenton 1980; Klassen 1989).

There was an advance into northern Minnesota from the northwest to the Culver and Sugar Hills moraines by the Saint Louis sublobe (Alborn phase), probably before 11,800 B.C. (H.E. Wright 1972; Richmond and Fullerton 1983). The Saint Louis sublobe advanced over a region that had been previously glaciated by the Rainy lobe (Knaeble, Meyer, and Mooers 2004). It may have advanced while the Rainy lobe was still active in the area; the Saint Louis sublobe margin could have been at the Rabideau moraine at about the same time that the Rainy lobe was at the Big Rice moraine (Mooers, Larson, and Marlow 2005). Glacial Lake Norwood was formed as the Rainy lobe melted away from the Vermilion moraine, perhaps around 12,700 B.C. (Leverett 1932; Fullerton 1994). Melting of the Saint Louis sublobe led to the formation of glacial Lake Upham. In northwestern Minnesota and southwestern Ontario, in the vicinity of Lake of the Woods, the Sugar Hills phase of the Koochiching sublobe has been correlated with the Big Stone moraine in the Red River Valley. By about 11,200 B.C. the position of the Koochiching lobe was southwest of Lake of the Woods; it can be correlated with the Edinburg phase of the Red River lobe (Fullerton 1995, 2000). The Edinburg phase may be associated with landscapes contemporaneous with Clovis artifacts based on radiocarbon dating in other regions of North America. This region, which would become the eastern part of the Lake Agassiz basin, contained glacial Lake Koochiching along the east margin of the Koochiching lobe.

In northern North Dakota and southern Manitoba, the next major position of the Souris lobe after the Martin moraine is associated with the Minot phase (Clayton, Moran, and Bluemle 1980; Fullerton 2000). The Martin moraine and slightly younger ice margins were parts of landscapes that were contemporaneous with Clovis artifacts. This Martin margin can be correlated with the Cartwright moraine of the Assiniboine lobe dating to around 11,600 B.C. In western Manitoba by about 11,300 B.C. the Assiniboine margin was at the Pipestone and Oak River moraines and had reached the Qu’Appelle moraine sometime after 11,100 B.C. (Morlan, McNeely, and Nielsen 2000).

Glacial lake or pond deposits near Buffalo Creek, in southeast North Dakota, contained remains of Mammutthus (the Millarton mammoth) under a buried soil (the lower Aggie Brown Member). The age of the mammot is estimated at 11,800-9500 B.C. (Harrington and Ashworth 1986). Late Wisconsin ice-contact lake deposits east of Napoleon, North Dakota, contain proboscidean remains (Clayton 1962). In southeastern North Dakota, southwest of Fargo, outwash deposits formed by the melting of ice blocks of the Kensal-Oakes moraine of the Des Moines lobe at Moon Lake contain wood dated to 11,600 B.C. (Valero-Garces et al. 1997; Laird et al. 1998; J.S. Clark et al. 2001). In northeast South Dakota, in the upland between the James River to the west and the Minnesota River to the east, wood from sediments overlying glacial deposits at Medicine Lake indicates the presence of vegetated landscapes by 10,900 B.C. (Valero-Garces, Kelts, and Ito 1995). Woodfordian (Cary) outwash in eastern South Dakota contains the bones of Rangifer and Equus (Flint 1955). Outwash associated with younger Woodfordian (Mankato phase) end moraines in northeastern South Dakota contain Mammutthus remains.

Two sublobes of the Red River–Des Moines Lobe advanced eastward covering older Woodfordian deposits (Leverett 1932; H.E. Wright 1972; Richmond and Fullerton 1984). The Grantsburg sublobe extended northeastward from the Mississippi Valley region around Minneapolis, Minnesota, into eastern Wisconsin, while the Saint Louis sublobe extended southeastward over the drainage of the Saint Louis River. The Grantsburg sublobe of the Des Moines lobe overlies Saint Croix phase deposits and, based on radiocarbon ages on wood, may date 11,900-11,700 B.C. (H.E. Wright and M. Rubin 1956; H.E. Wright, C.L. Matsch, and E.J. Cushing 1973; Meyer 1989). Glacial Lake Lind formed as a result of the melting of the Superior lobe, while Lake Grantsburg developed in front of the Grantsburg sublobe (Leverett 1932; W.S. Cooper 1935; M.D. Johnson et al. 1999). Glacial Lake Anoka developed after the melting of the Grantsburg sublobe. The Saint Louis sublobe advance may have occurred prior to 11,800 B.C. (Mooers, Larson, and Marlow 2005) or perhaps as late as 11,600 B.C. or 10,900 B.C. (H.E. Wright and W.A. Watts 1969). Deposits of the Saint Louis sublobe have been reported to overlap the Rainy-Patrician lobe Vermilion moraine (Leverett 1932); this relationship would imply that the Rainy lobe was at the Vermilion margin before the advance of the Saint Louis sublobe, perhaps sometime before 11,800 B.C. It has also been proposed that the Rabideau moraine of the Saint Louis sublobe and the Big Rice moraine of the Rainy lobe were contemporaneous (Mooers, Larson, and Marlow 2005; Mooers, Marlow, and Phillip 2005). Melting of the Saint Louis sublobe led to the development of glacial lakes Aitkin, Upham, and Saint Louis (Upham 1899; Winchell 1901; Leverett 1932; Hobbs 1983). Wood on the top of a paleosol buried by Lake Aitkin dates to 11,600 and 11,500 B.C., suggesting the lakes may have been part of a landscape contemporaneous with Clovis. Although mammoth remains have been reported from this area (Stauffer 1945), the earliest direct indicator of human presence is a Clovis point from the Cloquet River drainage (Romano and Johnson 1990) in a region that had been deglaciated since the melting of the Rainy lobe to the vicinity of the Big Rice and Vermilion moraines (Hill 1994, 1995; Hill and Huber 1996; Mooers, Larson, and Marlow 2005).
During the Late Wisconsin, lakes were a critical component of the landscape that had been previously covered by glaciers during the time range associated with Clovis artifacts (fig. 5). Around 11,500 B.C., melting of the James lobe resulted in the formation of Lake Dakota in northeastern South Dakota and southeastern North Dakota. Lake Dakota was a proglacial lake within the James River until the ice location in North Dakota stood at the Oakes moraine (Hard 1929). At about this time the Des Moines lobe disappeared by melting back to the Red River valley, and the remaining ice formed the Red River lobe. Glacial Lake Minnesota developed with the melting of the Des Moines lobe to the Big Stone moraine, while Lake Agassiz developed in front of the Red River lobe. Farther north, the Weyburn lobe of the Laurentide began to melt in southeastern Saskatchewan around 11,100 B.C., leading to the presence of Lake Regina (Klassen 1989, 1994). Water from Lake Regina drained southward into Lake Souris, which formed in the region that previously contained the Souris lobe, in southwestern Manitoba and north-central North Dakota. Lake Souris first drained southward into Lake Dakota and Lake Agassiz. After the Leeds lobe had receded into Manitoba, Lake Souris water drained northward into Lake Hind and then along the ice margin into Lake Agassiz (Lockhart stage) within the Red River valley around 11,100 B.C. (Elson 1958; Lemke 1958; Teller 1987, 2004).

Deposits of glacial Lake Hind are found along the present-day Souris River in southwestern Manitoba. The regression of this proglacial lake is associated with a flood event that originated in glacial Lake Regina and flowed into glacial Lake Souris, eroding the Pembina spillway channel (later containing the Souris River) and draining the southern region of Lake Hind (Kehew 1982; Teller 1987; Kehew and Teller 1994; Sun and Teller 1997; Boyd, Running, and Havholm 2003). Seeds dated to 10,300 B.C. are associated with the regression of the north part of glacial Lake Hind and flooding events through the Qu’Appelle and Assiniboine spillway channels. Folsom-Midland artifacts (fluted and unfluted) have been found within the basin of glacial Lake Hind, indicating human presence after the drainage of the lake. Folsom points have also been found in southwestern Manitoba west of the Moorhead phase shoreline of Lake Agassiz (fig. 6) (Buchner and Pettipas 1990).

Glacial Lake Agassiz initially formed as ice melted back from the Big Stone moraine. Mammoth (Mammuthus primigenius) remains were recovered from the highest major (Herman) strandline of Lake Agassiz near Emden, North Dakota (Harington and Ashworth 1986). Mammoth remains are also known from under gravels of the Herman beach near Ripon, North Dakota (Upahm 1895) and from the Herman strandline near Absaraka, North Dakota (Harington and Ashworth 1986). Since the Herman beach dates to about 11,500-11,000 B.C. (K.L. Harris, M.R. Luther, and J. Reid 1986; Leverington, Mann, and Teller 2000), it was part of a landscape that in other regions of North America contains evidence for the presence of Clovis artifacts.

The Lockhart phase of Lake Agassiz (Herman, Norcross, Tintah, and Campbell 1 shorelines; K.L. Harris, M.R. Luther, and J. Reid 1986), which ended possibly as late as 10,800 B.C. (T.G. Fisher 2003), may have been contemporaneous with humans using Clovis artifacts. Surface finds of Clovis artifacts are distributed west of the Lockhart phase shoreline in southwestern Manitoba (Buchner and Pettipas 1990). During the Moorhead phase, water levels dropped because glacial melting to the east in the Superior basin created a new outlet. Dates on the Moorhead range from 10,800-9800 B.C. in North Dakota (Clayton and Moran 1982; Yansa and Ashworth 2005), indicating this low-water stage was contemporaneous with the Younger Dryas (Teller and Leverington 2004). The Moorhead regression (associated with the McCauleyville, Blanchard, Hillsboro, Emerado, and Oaketa shorelines) (K.L. Harris, M.R. Luther, and J. Reid 1986) may be associated with the younger range for Clovis artifacts, but it is probably the landscape that was contemporaneous with the presence of Folsom artifacts (fig. 6). The Marquette advance of the Superior lobe blocked the drainage outlet of Lake Agassiz and led to the Emerson phase, with shorelines at the level of the Campbell II beach. On the western side of the Lake Agassiz basin, along the Assiniboine River in southern Manitoba, west of Winnipeg, there are wood dates of 8900 B.C. from a lagoon deposit behind the Campbell beach that formed after the Moorhead low water phase (Teller 1989, 2001). Early Emerson phase ages range about 9500-9300 B.C. (Bajc et al. 2000; Teller et al. 2000; Leverington and Teller 2003; Yansa and Ashworth 2005). The landscape associated with melting ice of the Rainy and Wadena lobes in central Minnesota appears to have been a open spruce forest around 9500 B.C. (H.E. Wright et al. 2004). The end of the Emerson phase is around 8500 B.C. (Teller et al. 2000; Teller and Leverington 2004).

In the western Great Lakes, the Woodfordian maximum is correlated with the 7500-22,000 B.C.), and the Saint Croix phase (19,700-16,200 B.C.) of the Superior and Rainy lobes in Minnesota (M.D. Johnson and H.D. Mooers 1998; Phillips and Hill 2004). These lobes combined to advance south and west from the Superior basin and the region to the west reaching the area of Minneapolis and Saint Paul in Minnesota. The Saint Croix moraine merges to the west with the Itasca moraine of the Wadena lobe. These may correlate with the Johnstown, Saint Johns, and Sagola margins of the Green Bay and Michigan bavine lobe of Michigan and the West Chicago margin of the Michigan lobe (Clayton and Moran 1982; Clayton 1984; W.L. Peterson 1986; Attig, Clayton, and Mickelson 1985; Clayton, Attig, and Mickelson 2001). The Harrison moraine of the Chippewa lobe from the Superior basin is slightly younger than deposits correlated with the Johnstown moraine that is likely older than 14,200 B.C. (R.F. Black 1976). Glacial
lakes formed west of the Green Bay lobe (Clayton and Attig 1989).

During deglaciation after the Saint Croix phase, the Rainy or Patrician lobe melted back to the Vermilion moraine in northern Minnesota, and the Superior lobe melted back to the Superior basin (H.E. Wright and W.A. Watts 1969; Matsch and Schneider 1986). The ice margin associated with the Vermilion moraine may date to about 13,100 B.C. Proglacial lakes Upham and Aitkin I developed in the region left deglaciated by the Rainy lobe during this interstadial. While the Rainy lobe remained stable in the vicinity of the Vermilion moraine, the Superior lobe readvanced to the Mille Lacs and Highland moraines and the Frederic ice-margin position in Wisconsin (Leverett 1932; H.E. Wright 1972; M.D. Johnson and H.D. Mooers 1998). Another melting of the Superior lobe led to the formation of a proglacial lake in the southeast end of the Superior basin.

The Superior lobe at the time of the advance of the Grantburg sublobe of the Des Moines lobe may have been at the Mille Lacs–Highland moraine margin in northeastern Minnesota and may correlate with the Tiger Cat and Summit Lake margin in northern Wisconsin (Clayton 1984; Attig, Clayton, and Mickelson 1985; Matsch and Schneider 1986). In southeastern Wisconsin the stratigraphic equivalent may be the Wadsworth Member (Willman and Frye 1970; W.H. Johnson 1976).

Ice appears to have withdrawn from the Superior basin with the glacial margin located along the north shore of Lake Superior, in Ontario (Phillips and Hill 2004; Anderton, Regis and Paquette 2004). The Superior basin contained a lake that received water from Lake Agassiz during the Moorhead low-water stage (most likely contemporaneous with Folsom artifacts). In upper Michigan, this interval is correlated with the Gribben forest interstadial (Lowell et al. 1999; Karrow, Dreimanis, and Barnett 2000). Ice again advanced southward from the Superior basin and buried the Gribben forest under lake and outwash sediments, during the Marquette phase. This ice advance blocked the eastward outlet of Lake Agassiz and led to the Emerson high water stage. The maximum of the Marquette phase is dated to about 9600 B.C., at the end of the Younger Dryas (Lowell et al. 1999). While the Marquette ice was present in the central and eastern Superior basin and adjacent areas of northern Wisconsin and Michigan, glacial Lake Duluth formed in the west part of the basin. A till correlated with the Marquette phase is stratified between glacial Lake Duluth sediments in northeastern Minnesota; glacial ice expanded from the southwest part of the Superior basin reaching the inner margins of the Nickerson and Thomson moraines (Mooers, Larson, and Marlow 2005). The lake may have extended northward to the Marks moraine, the Marquette phase position in Ontario. If humans entered the Great Lakes region during the Gribben interstadial, before the Marquette advance, evidence for their presence would be expected to be associated with features of abandoned high shorelines along the west side of Lake Superior (Phillips and Hill 2004). Agate Basin artifacts may be about the same age as the Gribben interstadial, while Hell Gap artifacts may be contemporaneous with the subsequent Marquette advance. Shorelines would also have been present during the Marquette advance if the ice margin extended from the Marks moraine southeastward to the Whitefish moraine and then into the present-day Lake Superior basin. To the west in the region deglaciated by the Rainy lobe, Agate Basin and Hell Gap artifacts have been recovered along the Cloquet River drainage (Hill 1994, 1995; Hill and Huber 1996).

The Rainy lobe melted northward from the Vermilion moraine in northern Minnesota probably sometime after 11,800 B.C. to the Steep Rock moraine in southern Ontario and then to the Brule Creek moraine (Bjork 1985). A proglacial lake, glacial Lake Norwood (Winchell 1901), formed as the ice melted from the Vermilion moraine. It is possible that Upper Herman stage of Lake Agassiz extended to the Steep Rock moraine around 11,100 B.C. (Fullerton 1995a). Human groups using Clovis artifacts elsewhere in North America would coincide in time with landscapes associated with the ice position at the Steep Rock moraine, and the Brule Creek moraine, dated to about 11,000 B.C. (fig. 5) (Fullerton 1994). The Brule Creek moraine was eroded by, and thus older than, the Marquette advance of the Superior lobe which formed the Marks moraine. The Marks moraine is contemporaneous with Mackenzie interlobate moraine and the Dog Lake moraine (Phillips and Hill 2004), which indicate the location of the Laurentide ice margin around 9500 B.C., immediately northwest of present-day Lake Superior.

Evidence for a “post-Folsom” advance of glacial ice from the Superior basin in northern Wisconsin includes several radiocarbon dates associated with moraines or glaciogenic sediments. For example, wood under the Saxon margin till ranges 9700-9100 B.C. and may correlate with the Porcupine margin dated to 9900 B.C. and the Marquette margin dated from 9600 B.C. to possibly as late as 9400-9300 B.C. (Brubaker 1975; R.F. Black 1976; Richmond and Fullerton 1984; Lowell et al. 1999). Agate Basin and Hell Gap artifacts may be contemporaneous with the Gribben interstadial phase and Marquette advance while Scottsbluff-Eden artifacts may be associated with the Marquette deglaciation and the lowering of lake levels in the Superior basin to the Houghton stage by about 8100 B.C. (Anderton, Regis, and Paquette 2004).

Melting of the Laurentide ice sheet north of the Lake Superior basin is associated with a series of moraines that reflect the location of the ice margin. By about 8900 B.C. the Rainy lobe position was at the Whitewater, Nipigon, and North Shore moraines (Nielsen, McKillop, and McCoy 1982). The Cochrane phase margin marks the position of the Rainy lobe around 7400 B.C. (Fullerton 1994). By about 7000 B.C. the Laurentide ice margin was at the Agutua-Nakina moraine. Retreat from this moraine led to the
drainage of water from Lake Agassiz into Lake Ojibway (Breckenridge et al. 2004).

During the early part of the Woodfordian (Late Wisconsin) along the west side of the Lake Michigan basin, in present-day Wisconsin, the Green Bay lobe advanced southwestward, in the north merging with the ice advancing from the Superior basin, and in the south advancing over the Fox River valley to a maximum position in the vicinity of Madison marked by the Johnstown moraine (McCartney and Mickelson 1982; Colgan 1999; Syerson and Colgan 2004). The initial Woodfordian advance occurred sometime after 28,800 B.C. Radiocarbon dates of 13,900 B.C. from the Valders Quarry and about 12,800 B.C. from Devils Lake help provide an estimate for the melting of the ice (Maher and Mickelson 1996; Maher et al. 1998). Lakes formed as part of the landscape west of the Green Bay lobe (Clayton and Attig 1989).

The initial Late Wisconsin advance of the Lake Michigan lobe formed the Darien moraine in Wisconsin (Alden 1918; Eschman and Mickelson 1986; Syerson and Colgan 2004). Recession of the ice led to the development of glacial Lake Milwaukee. The next advance is marked by the Valparaíso moraine. As ice receded from this margin several other moraines were formed including the Tinley moraine and the Lake Border moraines in southeast Wisconsin. The Schaef er, Heboir, Mud Lake, and Fenske mammoths are situated on these landforms. Ages on bone and wood associated with these mammoth fossils range from 14,800-10,900 B.C. indicating the presence of inhabitable landscapes at least by the time elsewhere associated with Clovis artifacts (Overstreet and Kolb 2003). The presence of Clovis, Folsom, and younger artifact forms as well as vertebrate fossils in the region west of the Green Bay and Michigan lobes may be an indicator of post-Woodfordian environmental conditions in the western Great Lakes region (Dudzik 1991; R.J. Mason 1986; Kuehn 1998; Stoltman 1998; Holman 2001).

As ice from the last Woodfordian advance receded, a forest developed on lake sediments that formed part of the deglaciated landscape. In eastern Wisconsin the presence of this forest is used to define the Two Creeks interstadial and the Twocreekan chronzone. The duration of the forest was about 11,900-11,600 B.C. (Broecker and Farrand 1963; Fullerton 1980; Schneider 1990; Kaiser 1994). The end of the interval appears to coincide with the earliest possible presence of human groups using Clovis artifacts, and well before the Younger Dryas. It is also associated with the rising waters of glacial Lake Chicago, which flooded the Two Creeks forest as the Green Bay lobe advanced again from the Lake Michigan basin. The Calumet shorelines may be associated with the ice advances before and after the Two Creeks interval (Schneider and Hansel 1990). At its type locality, the Calumet shoreline appears to have formed after the Two Creeks interval and is associated with the advance of the ice to the Two Rivers moraine. At the Two Rivers type locality wood under the glacial deposits is from the Two Creeks forest and dates to 11,700 B.C. (Schneider 1990). Shoreline features of glacial Lake Chicago in Illinois contain wood with a range of 11,700-10,900 B.C. (Schneider and Hansel 1990), thus indicating they were part of a landscape associated with human groups using Clovis artifacts elsewhere in North America.

Glacial Lake Middleton was formed as the Green Bay lobe melted from the Johnstown moraine and the recessional Milton moraine (Ripley 1998). Continued melting of the Green Bay lobe led to the development of glacial Lake Yahara, probably prior to 11,800 B.C. with lowered lake levels by 11,000 B.C. Loess deposition and soil development in the uplands began by about 11,800 B.C. and buried the glacial lake sediments. The surfaces of landforms in this area contain Gainey, Folsom, Agate Basin and Cody artifacts.

In Illinois, the maximum limit of the Lake Michigan lobe is associated with a series of advances and retreats dated about 28,800-14,700 B.C. (Hansel and Johnson 1999; Curry and Baker 2000; Stiff and Hansel 2004; Mickelson and Colgan 2004). The Marengo moraine marks the earliest advance around 28,500-28,100 B.C. based on wood and peat dates. The Shelbyville and Bloomington morainic systems mark the ice position at about the Last Glacial Maximum with dates of 22,400-21,700 B.C. Proglacial lakes formed along the margin of the ice as it melted back into the Lake Michigan basin. By about 14,600 B.C. the ice margin was north of Chicago, based on wood from shore-zone sediments of glacial Lake Chicago. Loess deposits (the Peoria Silt) overlie the glacial deposits and date about 28,100-12,700 B.C.

There was an advance of the Michigan lobe around 12,500-11,800 B.C. to about Milwaukee (Port Huron phase) (Krist and Lusch 2004). Following a retreat, there was an advance to Two Rivers around 11,500 B.C.; this is the advance that buried the Two Creeks forest. The Greatlakes advance maximum dates to about 11,600 B.C. It melted from Upper Michigan before 11,000 B.C., forming glacial Lake Ontonagon within the Superior basin (fig. 5). The lake may have persisted until about the Marquette advance, which reached its maximum around 9500 B.C. As the Marquette ice melted, Lake Ontonagon was present along the margin, until it merged with glacial Lake Duluth around 9100-9000 B.C. (Hack 1965; Syerson and Colgan 2004). Because ice was still in the eastern portion of the Superior basin, the lake drained into Lake Chippewa in the Lake Michigan basin through the Au Train–Whitefish channel in Upper Michigan. As the ice melted completely from the Superior basin, Lake Minong formed, after 9000 B.C. (Julig, McAndrews, and Mahaney 1990; W. Ross 1997).

Conclusion

Radiocarbon dates on wood and vertebrate remains suggest that nonglacial landscapes existed in the central region of
North America during the Middle Wisconsin. The nonglaciated region likely extended from the Mackenzie River delta in the north to the western Great Lakes in the south. The Laurentide ice expanded to the west and to the south as a set of individual lobes during the late Wisconsin (Woodfordian) (figs. 2-6). The advance and melting of these various lobes led to the presence of glacial lakes. The largest were glacial Lake McConnell, along the northwest margin of the Laurentide ice, and glacial Lake Agassiz, along the southern margin. Both were present by about 9500 B.C.

In the western Great Lakes region, the Woodfordian ends with the beginning of the Twocreekan chronozone, defined by radiocarbon dates on wood from the Two Creeks forest. Landscapes that developed after the burial of the Two Creeks forest are associated with the time interval linked to Clovis, Folsom, and Plano artifacts elsewhere. The landscapes along the margin of the Laurentide ice sheet at about 11,000 B.C. were characterized by stagnant, melting ice and glacial lakes. In the recently deglaciated regions to the west and south of the ice margin, landscapes appear to have contained biomes that were inhabited by Rancholabrean fauna.