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Geologic Framework and Glaciation of the Eastern Area

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Late Pleistocene landscapes in glaciated eastern North America included changing ice margins, fluctuating lake and sea levels, and deglaciated physical settings that were inhabited by a variety of extinct (Rancholabrean) fauna. The glaciated East of North America consists of the midcontinent from Hudson Bay to south of the Great Lakes and extends eastward to the Atlantic coast. Glaciers were present along the Atlantic coast from southern New York north to Labrador.

Some of this region appears to have been ice-free during parts of the Middle Wisconsin; the interstadial ice margin around 33,400-29,400 B.C. may have been situated within the northern parts of most of the Great Lakes basins, with ice covering Ontario, Quebec, Labrador, and Newfoundland (fig. 1). During the Late Wisconsin, glaciers extended southward from the Great Lake basins and eastward onto the Atlantic continental shelf. Geomorphic features and sedimentary deposits record a series of glacial advances and retreats that are connected with ice-margin lakes and changes in relative sea level (figs. 1-9). The deglaciated landscapes were able to sustain some elements of the Rancholabrean fauna.

The phases of glaciation represent the advance of lobes of the Laurentide ice sheet and their subsequent retreat or stagnation. There is also evidence for independent, local glaciers in northern New England and the Canadian Maritime provinces. Typically these glacial events are reflected by the presence of end moraines or recessional moraines and other types of till (fig. 10), as well as lacustrine and fluvial deposits associated with the glacial outwash or the blocking of drainages by glacial ice. Multiple recessional moraines at times formed as the ice melted back from its maximum advance. Thus evidence of the glacial phases includes sedimentary deposits, stratigraphic sequences, and physiographic (geomorphic) features. The timing and duration of the glacial phases can be inferred based on stratigraphic relationships and radiocarbon measurements. Radiocarbon measurements made on plant materials (e.g., wood, charcoal) and mammal bones serve as the principal means of developing and interpreting the temporal framework, although less reliable measurements on shells are still the basis for developing a chronologic framework in coastal-marine regions.

In the Great Lakes region, the Wisconsin stage has been divided into a series of chronostratigraphic units (W.H. Johnson et al. 1997; Karrow, Driemanis, and Barnett 2000). The Altonian substage dates to before 30,500 B.C., while the Farmdalian substage ranges in age from about 30,500 to 28,000 B.C. The Woodfordian substage ranges in age from about 28,000 to about 12,800 B.C.; it is associated with extensive glacial activity and subsumes previously used terminology such as Tazeville, Cary, and Mankato (Willman and Frye 1970). The Twocreekan substage is a short interval after the Woodfordian and before the Greatlakean, generally ranging 12,800-11,800 B.C. The Greatlakean designates the interval after the Twocreekan to about 5900 B.C.; this interval had previously been designated the Valderian (Evenson et al. 1976). It partly coincides with the Younger Dryas, about 11,000-9600 B.C.

Several alternative approaches to classification have been applied in attempts to provide a realistic framework that reflects the time-transgressive nature of paleoenvironmental events and their relationship to geomorphic features and sedimentary deposits. Thus, for example, the Wisconsin Episode has been subdivided into several diachronic units in various parts of the Great Lakes region. The time transgressive Alton and Farmdale phases in the area influenced by the Lake Michigan lobe generally coincide with the Port Talbot, Brimley, and Farmdale phases in the eastern and northern Great Lakes. In this conceptual framework the late Wisconsin is divided into phases reflecting the advance and recession of various glacial lobes (e.g., Shelby phase, Port Bruce phase, Port Huron phase, Mackinaw phase, Two Creeks phase, Two Rivers phase, Gribben phase, Marquette phase). Thus, several methods of classification (time-stratigraphic or chronostratigraphic, climate or event, and diachronic) have been applied in the area.

Proglacial lakes developed along the ice fronts, and relative sea levels were influenced by glaciation and deglaciation. Shorelines of glacial lakes and inferred coastlines have been used to identify potential landforms associated with the presence of Late Pleistocene human groups and to develop a relative chronology for artifact forms found in the region. Thus, Gainey and Bull Brook artifact assemblages have been inferred to be regional variants or temporal
equivalents of Clovis (with an age range estimated at about 11,000-10,500 B.C.); Parkhill, Barnes, and Debert assemblages may roughly coincide in time with Folsom artifacts; and Crowfield and Holcombe artifact forms are associated with later landscape contexts.

The glaciated East includes the glacial record and related postglacial landscapes of three major regions: the Great Lakes; the area east of the Great Lakes to the Atlantic coast, but south of the present Saint Lawrence River (including parts of Pennsylvania, New Jersey, and southern Quebec as well as New York, New England, New Brunswick, and Nova Scotia); and parts of Ontario, Quebec, and Labrador that remained glaciated after 9600 B.C. The western boundary of the Great Lakes region includes the Superior and Michigan basins and adjacent areas of western Ontario, Minnesota, Wisconsin, and Michigan. The southern boundary includes northern Illinois, Indiana, Ohio, Pennsylvania, and New York. As the Laurentide margin receded northward from the Great Lake basins, it left a deglaciated landscape that extended into Ontario. Laurentide ice also advanced southward from the Saint Lawrence River region into New York and the northern parts of Pennsylvania, and New Jersey, as well as southeastward across New England to the Atlantic coast; these have been designated as the Champlain-Hudson and New England lobes (Mickelson et al. 1983; Mickelson and Colgan 2004). Besides the Laurentide ice sheet, local ice centers appear to have formed in northern New England as well as in the Canadian Maritime provinces. These local ice centers persisted after the Laurentide ice sheet had receded from the region (Dyke 2004). Moraines and other geomorphic features and sediments in parts of Ontario, Quebec, and Labrador indicate the extent and timing of the Laurentide ice after 9600 B.C.

**Great Lakes Region**

The geological framework associated with the dynamics of glaciation and deglaciation in the Great Lakes region is the product of several lobes of the Laurentide ice sheet as it advanced southward from Ontario (figs. 1 and 10). Glacial ice extended from the Superior basin southwestward into northern and central Minnesota as the Superior lobe, and into northern Wisconsin as the Chippewa, Wisconsin, and Langlade sublobes (“Geological Framework and Glaciation of the Central Area,” this vol.). Two major lobes of ice advanced from the Michigan basin. The Green Bay lobe extended into northeastern Wisconsin, while the Michigan lobe advanced into eastern Wisconsin, northeastern Illinois, northwestern Indiana, and western Michigan. The Lake Huron lobe advanced southward across Michigan and Ontario into northern Illinois, Indiana, and Ohio. The west side of the Erie lobe appears to have coalesced with parts of the Huron lobe (forming the Erie-Huron lobe); it advanced into northeastern Ohio, northwestern Pennsylvania, and western New York. Several sublobes extended eastward from the Lake Ontario basin and adjacent Saint Lawrence lowlands.

**Middle Wisconsin**

Along the shore of Lake Michigan, fragments of wood and peat between two tills have dates of 41,000 and 40,300 B.C. (Eschman and Mickelson 1986). Illinois was ice free until about 28,000 B.C. (Stiff and Hansel 2004). Mammoth remains from the Saginaw Bay area of central Michigan, dated to about 26,400 B.C., may also indicate Middle Wisconsin ice-free conditions (Farmlandian or Plum Point interstadials, Kapp 1970). In Ohio, a paleosol developed in loess contains wood dated to 30,900 B.C. (Fullerton 1986). Wood recovered from till in Ontario ranges 31,200-29,800 B.C., and deposits underlying a till at the Woodbridge locality near Toronto, Ontario, contain fossils of grizzly; Churcher and Morgan 1976). Based on a radiocarbon date on bone of 32,500 B.C. from this unit, the bear and mammoth bones probably reflect Middle Wisconsin landscapes (Karrow et al. 2001). Logs found below the Kent Till south of the Lake Ontario basin range 27,700-24,800 B.C., and in New York wood fragments recovered from a till have been dated to 27,500 B.C. (Fullerton 1986). Thus, prior to the Late Wisconsin glacial advance, nonglacial conditions appear to have prevailed at least south and west of the Superior basin, while the southern parts of the Michigan, Huron, and Ontario basins were ice-free, as was the entire Erie basin (fig. 1) (Dyke et al. 2002).

**Late Wisconsin**

The extent and timing of Late Wisconsin glaciation had a direct effect on the physical landscapes in the Great Lakes region. The margins of the glacial lobes and sublobes are often indicated by the presence of moraines (fig. 10). In some instances lake strandlines and other physiographic or depositional features can be related to particular ice margins or the postglacial development of the landscape. Lacustrine, paludal, fluvial, and eolian sediments overlying the glacial till also contain evidence of dynamic changes in the landscape. These deposits contain plant and animal material that can be dated and serve to constrain the glacial chronology of the region, as well as to provide indications of the potential landscape conditions present during the Late Wisconsin.

- **Glaciation in the Lake Superior and Michigan Basins**

There were several phases of glaciation associated with the Superior basin (Phillips and Hill 2004). The age of the earliest Woodfordian ice advance could be in the range of 28,000-22,000 B.C. In contrast to the other Great Lake basins, there is evidence that the Superior basin was filled with ice around 11,000 B.C. (fig. 6) and again, during the Marquette phase of glaciation, around 9600 B.C. (fig. 8). By this time the rest of the Great Lakes appear to have been entirely ice-free, and the Laurentide ice margin seems to...
have been north of the Saint Lawrence River. Ice may have persisted in the northern part of the Lake Superior basin until about 8900 B.C. Continued northward recession of the ice margin in Ontario eventually led to the formation of glacial Lake Ojibway starting around 8100-7500 B.C. (fig. 9). With the separation of the Laurentide ice into two ice centers around 6500 B.C., Lake Ojibway became part of the Hudson Bay drainage.

There was a succession of eight major glacial phases of the Lake Michigan lobe (Hansel and Johnson 1992). The Michigan lobe extended southward into Illinois between 28,000-14,800 B.C. (Stiff and Hansel 2004). The Marengo phase marks an advance around 28,000 B.C. The maximum advance of the Michigan lobe was during the Shelby phase; the ice-margin formed the Shelbyville and Bloomington moraines. The advancing ice led to the formation of glacial Lake Milan around 22,400 B.C. (Curry 1998). There were several additional advances into southern Wisconsin, northern Illinois, and adjacent Indiana and Michigan by the Michigan lobe. A major geomorphic feature is the Valparaiso moraine system, which is the result of an advance that occurred around 16,800 B.C. (Stiff and Hansel 2004).

An early phase of glacial Lake Chicago (fig. 2) formed as the ice receded from the southern part of the Michigan basin around 15,400 B.C. (Hansel and Johnson 1992), but the Tinley moraine and younger Lake Border moraines are the results of subsequent advances of the Lake Michigan lobe. Interstadial (nonglacial) conditions associated with the Twocreekan ended around 11,800 B.C. with the Two Rivers phase advance along the western margin of the Michigan basin (fig. 5). By 10,500 B.C., the Michigan basin and surrounding area appears to have been entirely free of active glacial ice (fig. 7).
Pleistocene plant and animal remains from deposits in the region of the Shelbyville complex of moraines provide an indication of the deglaciated landscapes in northern Illinois (Schubert et al. 2004). Pond deposits contain Jefferson’s ground sloth (Megalonyx jeffersonii) and stag-moose (Cervalces scotti) fossils both dated to about 11,400 B.C. At the margin of the Cropsey moraine in Illinois, wood from lake sediments has been dated to 11,100 B.C. (Pasenko and Schubert 2004). A mammoth tooth was recovered from the lake deposits.

The Valparaiso moraine system of the Michigan lobe is present in southwestern Wisconsin, northeastern Illinois, northwestern Indiana, and adjacent southern Michigan. Outwash deposits of the Valparaiso ice margin are overlain by peat that has been dated to 16,500 B.C. as well as mammoth bones dated to 14,100 B.C. (Springer and Flemal 1981; W.H. Johnson 1986). South of the Valparaiso moraine, in outwash deposits between the older Maxinkuckee moraine and the Packerton moraine of the Huron lobe (in north-central Indiana), wood found beneath the Wells mastodon was radiocarbondated to 12,300 B.C. (Gooding and Ogden 1965). The Powers mastodon (Mammut americanum), dated to 11,200 B.C., was found in gray clay and peat deposits between the Valparaiso and Kalamazoo moraines of the Michigan lobe (Garland and Cogswell 1985). These deposits overlie Woodfordian river and proglacial sediments.

Several moraines were formed by the Michigan lobe between the Valparaiso moraine complex and the Lake Chicago Plain in southeastern Wisconsin. The Tinsley moraine is east of, and younger than, the Valparaiso moraine system. Near the margin of the Valparaiso and Tinsley moraines, the Mud Lake mammoth has been dated to 14,900-12,700 B.C. (Overstreet and Kolb 2003). The Lake Border moraine complex lies between the Tinsley moraine and the Lake Chicago Plain. Mammoth bones at the Fenske locality situated on the Lake Border moraine have ages of 14,800 B.C. Mammoth bones found in a glacial pond on the Lake Border moraine at the Schaefer locality have radiocarbon ages ranging 13,000-12,400 B.C. (Joyce 2006). The oldest wood date, inferred to indicate the presence of a deglaciated landscape, is 13,900 B.C. The nearby Heboir mammoth has been dated to 13,000-12,800 B.C. Wood at
Wisbog bog south of the Schaefer mammoth provided an age of 13,000 B.C. (Huber and Rapp 1992).

Three mastodons have been found west of the Michigan basin in southern Wisconsin associated with postglacial lake sediments. Wood from these deposits was dated to 13,700 B.C. and 10,900 B.C., while the mastodon bones were dated to 9000 and 8900 B.C. (R.M. West and J.E. Dallman 1980). There are several examples of the remains of mammoths in deposits associated with postglacial lakes, and bison (Bison occidentalis) remains have been found with artifacts.

**GLACIATION IN THE HURON, ERIE, AND ONTARIO BASINS**

The Huron lobe expanded southward across Michigan developing several sublobes (Mickelson et al. 1983; Eschman and Karrow 1985; Fullerton 1986; Krist and Lusch 2004). The maximum late Wisconsin glaciation ice position is marked in most of northern Indiana by deposits of the East White sublobe, in eastern Indiana and western Ohio by sediments related to the Miami sublobe, and in central Ohio by till of the Scioto sublobe (Fullerton 1980, 1986; Szabo and Chanda 2004). The maximum advance of the Miami sublobe around 22,000 B.C. extended almost to Cincinnati, north of the Ohio River, in southwest Ohio and southeast Indiana (Lowell 1995; Huesser et al. 2002). Radiocarbon dates indicate the advance began prior to 27,000 B.C. (Lowell 1995). Northeastern Ohio contains till associated with the Killbuck and and Grand River sublobes of the Lake Erie lobe. The maximum advance of the Huron lobe in Indiana and western Ohio is correlated with the Shelbyville, Hartwell (Miami sublobe), and Cuba (Scioto sublobe) moraines. The Hartwell moraine dates to about 21,400 B.C., while radiocarbon measurements indicate the Cuba moraine ranges 26,000-22,400 B.C. (Lowell, Hayward, and Denton 1999). In eastern Ohio, northwest Pennsylvania, and southwest New York the maximum advance of the Erie lobe is indicated by the Kent moraine. The maximum advance is estimated at 25,900 B.C.

Huron lobe margins in northeast Indiana, northwest Ohio, and southern Michigan are marked by a series of younger moraines. Some of these include the Union City-Powell, Mississinewa, Saint Johns, Wabash, Fort Wayne, Defiance, Packerton, and Kalamazoo moraines (fig. 10). The moraines of the Huron lobe reflect glacial positions...
associated with the post-Last Glacial Maximum Port Bruce phase and later Port Huron phase glacial advances. The Union City and Powell moraines mark the position of the Huron lobe at about 16,800 B.C. and may be generally contemporaneous with the Valparaiso moraines of the Michigan lobe (Mickelson et al. 1983). A kettle lake formed in the Cartersburg Till situated just south of the Union City moraine has bog sediments as old as 15,500 B.C. that are overlain by mastodon bones. The deposit containing the mastodon fossils ranges in age from 14,300 to 12,100 B.C. (Graham, Holman, and Parmalee 1983). Lake sediments between the Union City and Bloomer end moraines in west-central Ohio contain Jefferson’s ground sloth and stag-moose; wood from the lake deposit has been dated to 12,400 B.C. (Schubert et al. 2004).

The minimum age of recession from the Powell moraine is about 16,100 B.C., based on radiocarbondated wood north of the moraine. Between the Saint Johns moraine and the younger Wabash moraine in Ohio, intermorainal lake deposits contain mammoth bones; wood from these deposits dates to 10,200 B.C. (M.C. Hansen et al. 1978).

As ice of the East White sublobe of the Huron-Erie lobe receded, it formed the Packerton moraine. During the Lagro phase, the Miami sublobe of the Huron-Erie lobe advanced into the region and covered the Packerton moraine, before receding back to the Fort Wayne moraine in eastern Indiana. Remains of mastodon dated to 11,100 B.C. have been found in the kettle basins formed as part of this glaciated landscape (Hunt and Richards 1992). In Michigan, the Pleasant Lake mastodon was recovered in peat above a lacustrine deposit that overlies the till associated with the Fort Wayne moraine (D.C. Fisher 1984). Wood above the lake deposit but below the mastodon fossils has been dated to 13,700 B.C., while wood within the cavities of the mastodon tusks yielded an age of 10,400 B.C.

Another glacial advance associated with the Huron lobe Port Bruce phase is dated to about 15,800-15,400 B.C. (fig. 2); this advance led to the deposition of the Elma Till. As the Huron lobe receded, melt-out of ice in tills and glacio-lacustrine and glacial fluvial sediments was deposited over the Elma Till. Depressions created by the melt-out were filled with silt and gravel and eventually marls and peats. An example of these postglacial processes is at the Rostock site in southwestern Ontario, where plant debris recovered from the marls dated to 12,900 B.C. helps to constrain the timing of deglaciation (Pliny, Morgan, and Morgan 1987). The peat overlying the marl contains bones of woolly mammoth radiocarbondated to 10,800 B.C.

The Huron-Erie lobe Port Bruce phase advance is associated with the Defiance moraine in southeastern Michigan (Menzies 2001). At the Shelton mastodon locality Woodfordian deposits are overlain by deposits containing wood and faunal remains. The wood ranges in age 12,500-11,700 B.C. (Shoshani 1989). Mammoth remains found near Flint, Michigan, date to 11,400 B.C. (Oltz and Kapp 1963).

Immediately south of the Huron basin in northern Michigan is the Port Huron moraine. This major physiographic feature marks a glacial advance dated to about 14,000 B.C. (fig. 3) (Schubert and Packerton 2001; Krist and Lusch 2004) and has been correlated with the Whitehall and Manistee moraines in western Michigan and the Wyoming moraine in Ontario. After the Port Huron advance, the Laurentide ice margin began to recede into the Lake Huron and Georgian Bay basins. By 11,000 B.C., the ice position may have been north of Lake Huron and Georgian Bay, possibly at the Whiskey Lake and Cartier moraines.

To the east, different phases of the Erie lobe are marked by the Kent, Defiance, and Lake Escarpment moraines south of Lake Erie. The Kent moraine marks the maximum position of the Erie lobe as it extended generally southward into northeast Ohio, northwest Pennsylvania, and southwestern New York (Braun 2004). Wood beneath the Kent till dates to about 25,900 B.C. When the Erie lobe margin was at the Defiance moraine, Lake Maumee was present in northwestern Ohio and adjacent parts of Michigan and Indiana (Calkin and Feenstra 1985; Laub 2003).

The ice margin melted northward from Ohio into the Erie basin about 15,800-15,200 B.C. South of present-day Lake Erie, several glacial lakes were formed that appear to have been contemporaneous with stages of Lake Maumee and Lake Arkona in the Erie basin (Lakes Killdeer, Wharton, Vanlue, and Carey). This region contains the stratigraphic sequence at Sheridan Cave with radiocarbon dates of 13,700-12,800 B.C. on stag-moose, 11,600-11,400 B.C. on short-faced bear (Arctodus simus), and 10,400 B.C. on caribou (Rangifer tarandus). The sequence also contains two bone artifacts and an artifact that resembles a Gainey point in deposits radiocarbondated to 11,000-10,400 B.C. (Tankersley 1997, 1999; Tankersley, Redmond, and Grove 2001; Redmond and Tankersley 2005).

The last major Laurentide ice advance in southern Ontario is associated with the Port Huron phase (Fulton et al. 1986; Karrow 2004). Ice advanced into the eastern part of the Erie basin and completely filled the Ontario basin. The correlatives of the Port Huron moraines for the Lake Erie and Lake Ontario lobes may be the Paris moraine or the Hamburg or Marilla moraines in New York. The Girard moraine marks the position according to Fullerton (1980). By 12,000 B.C., active glacial ice was still present in the northern Huron basin and was north of the Ontario basin. By 11,000 B.C. the ice had receded north of the Huron basin.

The Port Huron advance is marked by the Hamburg moraine south of Lake Ontario in western New York. As the ice withdrew from the Alden, Buffalo, and Niagara moraines, Lake Warren developed and expanded (fig. 4). By about 12,000 B.C. the ice margin was at the Batavia moraine. Recession from the Batavia moraine led to an ice position marked by the Barre moraine. Lake Tonawanda formed south of the Barre moraine. Situated between the
Batavia and Barre moraines, the Hiscock site, New York, contains antler, bone, and twigs dated 11,400–11,200 B.C. (Laub 2003). Nearby, twigs at Divers Lake date to 11,500 B.C. Mastodon fossils at the Hiscock site range in age from 11,000 to 10,500 B.C. Hiscock contains fluted points that resemble Gainey points as well as bone tools dated from 11,000 to 10,800 B.C. The Carlton moraine lies immediately south of Lake Ontario and north of the strandline of Lake Iroquois (Laub 2003).

In western New York south of the Ontario basin, Lake Tonawanda appears to have disappeared by 11,600 B.C., based on wood from above lacustrine deposits at Whitney Creek (Tankersley et al. 1997). In this vicinity, wood ages of 10,300 B.C. constrain the minimum age of artifacts that resemble Clovis and Gainey forms from the Arc site assemblage. Near King Ferry, north of the Valley Heads moraine, wood imbedded in marl containing a mastodon has been dated to 11,400 B.C. (Deevey, Gralenski, and Hoffren 1959).

Melting of the Huron-Erie lobe led to the formation of glacial Lake Maumee at around 15,300 B.C., contemporaneous with glacial Lake Chicago to the west (Krist and Lusch 2004). Continued melting northward of the ice margin to the Port Huron moraine in northern Michigan around 14,000 B.C. led to the presence of glacial Lake Whittlesey and glacial Lake Saginaw (G.J. Larson, T.V. Lowell, and N.E. Ostram 1994; G.J. Larson and R.J. Schaeztl 2001). Retreat of the ice front led to the formation of Lake Warren in the Lake Huron basin. This may coincide with the Glenwood II phase of Lake Chicago in the Lake Michigan basin dated to about 13,600 B.C. The Greatlakean ice advance (equivalent to the Two Rivers phase advance in Wisconsin that led to the a burial of the Two Creeks forest around 11,800 B.C.) is associated with Lake Oshkosh in eastern Wisconsin, the Calumet stage of Lake Chicago in the Michigan basin, and Early Lake Algonquin in the Huron basin (Krist and Lusch 2004).

Strandline features associated with glacial lakes have played an important role in geoarcheological studies that have examined the relative ages of archeological occurrences in this region (Fitting, Devischer, and Wahl 1966; Fitting 1970; L.J. Jackson 1983, 2004; L.J. Jackson et al. 1966, 1970).
2000; Karrow 2004). Early Lake Algonquin in the Huron basin drained southward to Early Lake Erie and then into Lake Iroquois in the Ontario basin. Renewed melting led to a coalescence of the waters in the Michigan and Huron basins and the main phase of Lake Algonquin, around 11,000 B.C. (contemporaneous with Lake Ontonogan in the Superior basin). Shoreline features of this age have not been observed in southern Michigan; they may be submerged or eroded away (G.J. Larsen and R.J. Schaetzl 2001). A similar situation may have occurred along the northwest shore of the Superior basin where shorelines may have been eroded by lake transgression (Phillips and Hill 2004). If this is so, these strandlines and artifacts potentially contemporaneous with Clovis are not visible; the strandlines are present in southwestern Ontario.

The Holcombe localities in southeastern Michigan have been related to landscape features associated with Lake Algonquin (Fitting 1970). The localities are west of the Mount Clemens moraine and lie slightly above the shoreline of glacial Lake Clinton (Fitting, Devisser, and Wahla 1966). Holcombe artifact assemblages may date to about 10,200 B.C. (L.J. Jackson 2004). Barren-ground caribou remains have been recovered from the Holcombe site (Cleland 1965).

With continued recession of the Laurentide ice margin, meltwater from Ontario drained eastward to Lake Minong in the eastern part of the Superior basin, Lake Chippewa in the Michigan basin, and Lake Stanley in the Huron basin. After the maximum of the Marquette phase advance in the Superior basin around 9600 B.C. (Lowell, Hayward, and Denton 1999), Lake Duluth formed. Melting of the Marquette phase ice led to the expansion of Lake Minong after 8900 B.C. within the Superior basin (Phillips and Hill 2004).

In Ontario, the Thedford II site lies on glaciolacustrine deposits of Lake Warren, immediately southwest of
the Wyoming moraine, which has been attributed to the Port Huron phase advance (A.J. Cooper 1979; Deller and Ellis 1992). The artifact assemblage contains fluted points assigned to the Parkhill complex (Barnes points), possibly dating to around 10,600 B.C. Proglacial lake strandlines are present between the Wyoming moraine and present-day Lake Huron. These include shorelines of an early proglacial lake, and Lakes Lundy or Grassmere dated to about 12,800-12,600 B.C. (Fullerton 1980; Eschman and Karrow 1985). Features of glacial Lake Algonquin (12,000-10,400 B.C.) may have been destroyed by the middle Holocene Lake Nipissing transgression (Karrow 1980).

The Fisher site is situated along the Lake Algonquin shoreline in south-central Ontario, west of Lake Huron and south of Georgian Bay. It contains artifacts attributed to the Parkill complex (Nolin and Gwyn 1997). The site lies above the Newmarket Till, which was deposited about 16,200-15,400 B.C., and is older than the Port Huron phase advance, which is expressed locally as the Kettleby Till. An early stage of glacial Lake Algonquin may have formed 12,600-11,400 B.C., during the Two Creekan, with the melting of the ice associated with the Port Huron phase. The low-water “Kirkfield” stage, estimated to date to about 11,400 B.C., was followed by the main lake stage estimated to date 11,100-10,400 B.C.

In the region southeast of Lake Huron and north of the Lake Erie basin there are many examples of mastodon fossils and a few examples of mammoth (McAndrews and Jackson 1988). In contrast, mammoth fossils are more common to the north (east of Lake Huron and the western region of Lake Ontario). Some of these have been dated using radiocarbon measurements and provide an indication of when this region was deglaciated and inhabitable by large mammals. Immediately north of Lake Erie, wood associated with the Verbeke mastodon was dated to 9600 B.C. North toward Lake Huron, plant remains associated with the Perry mastodon provide an age range of 12,100-11,900 B.C., while collagen from the Thamesville mastodon has been dated to 11,300 B.C. (McAndrews and Jackson 1988). The Rostock mammoth, situated on the Elma Till plain between Lakes Huron, Erie and Ontario, has been dated to 10,800 B.C. (Pliny, Morgan, and Morgan 1987). The mastodon sites in southwestern Ontario are found in upland areas associated with the initial retreat of Port Bruce phase ice or are younger than strandlines of various glacial lakes. The maximum potential age for the fossils on landscapes associated with the Port Bruce phase deglaciation would be around 15,400 B.C. Mastodon remains are found below the Whittlesey, Warren, Grassmere, Lundy, Rouge, and Dana shorelines, indicating the presence of inhabitable landscapes from perhaps 13,200-10,400 B.C. (McAndrews and Jackson 1988). In southwestern Ontario, mammoth remains found above the Ridgeway Island moraine may date to about 13,200 B.C., while another was adjacent to the Orangeville moraine west of the Niagara Escarpment. East of the Niagara Escarpment there are several mammoths that occur on landscapes associated with the deglaciation of Port Huron phase ice.

In south-central Ontario several mammoth and mastodon remains can be related to glacial lake strandlines. One mastodon was found below the level of Lake Warren and thus is younger than about 12,800 B.C., while the Hamilton mastodon is the same age or younger than Lake Iroquois and thus has a maximum potential age of about 11,400 B.C. Several mammoths can be related to Lake Iroquois (McAndrews and Jackson 1988). A mammoth found near Egypt, Ontario, is associated with the strandline of Lake Algonquin, present about 11,400-10,400 B.C. (fig. 6). Glacial Lake Iroquois deposits in the Toronto region contain bison, musk-ox (Ovibos), and mammoth (Bensley 1923; Karrow 1967). Holcombe points are found below Lake Algonquin shorelines. They are considered, on the basis of this geomorphic relationship, to be younger than 10,400 B.C.

The Ontario lobe advanced southward into New York to the Valley Heads moraine, possibly reflecting the ice margin position during the Port Bruce phase. The Ashtabula moraines in Ohio and Pennsylvania and the Lake Escarpment and Valley Heads moraines in New York may be about the same age at 16,200-15,500 B.C. (Mickelson et al. 1983; Ridge 2004). However, mastodon remains recovered from sediments associated with the Valley Heads ice have been dated to 14,400 B.C. (Coates, Landry, and Lipe 1971). Wood associated with the King Ferry mastodon north of the Valley Heads moraine has been dated to 11,400 B.C. (Muller 1960). The Ontario lobe receded into the Ottawa Valley by about 11,500 B.C. (Dyke 2004). This led to the replacement of Lake Iroquois first by Lake Vermont and then later by the Champlain Sea. Lake Vermont has been dated to about 11,600-11,400 B.C. (Ridge et al. 1999).

Glaciation East of the Great Lakes to the Atlantic Coast

The northeastern region of North America is associated with several lobes of the Laurentide ice sheet and also appears to have contained several local centers of ice during the Wisconsin glaciation (Clark, Knight, and Gray 2000). The geologic framework for this region provides information on landscape evolution during the Middle Wisconsin and Late Wisconsin.

Middle Wisconsin

Radiocarbon measurements provide one line of evidence indicating the presence of nonglacial landscapes in New York, New England, and the Canadian Maritimes during parts of the Middle Wisconsin. Western New York radiocarbon dates of 50,300-28,800 B.C. (Young and Sirkin 1994) may correlate with Plum Point and Port Talbot interstadials in Ontario. In the Genesee Valley, south of Lake Ontario,
wood indicates ice-free conditions at around 27,500 B.C. (W.J. Brennan 1988). Nonglacial conditions associated with the Middle Wisconsin on Long Island may be reflected by redeposited wood dated to 41,000 B.C. (Sirkkin and Stuckenrath 1980; Stone and Borns 1986). Within the Saint Lawrence valley near Quebec City, deposits containing insects and plant macrofossils have yielded radiocarbon ages of 31,014 B.C., 40,100 B.C., and older (Cummings and Occhietti 2001). A fragment of caribou bone found in the gravels of the Saint Antonin moraine in southeastern Quebec dated 42,300 B.C. (Harington 2003). A regional glacial recession may have occurred in the New Brunswick area prior to 44,000 B.C., but several local ice caps began to develop in Atlantic Canada; there may have been an independent ice cap in New Brunswick during the Middle Wisconsin (Foisy and Prichonnet 1991; Seaman 2004). In Nova Scotia, wood in till dates to 36,000 and 26,700 B.C., indicating a Middle Wisconsin ice recession. Proboscidean fossils from Nova Scotia radiocarbondated to 34,400 B.C. may also imply a Middle Wisconsin ice-free interstadial (D.R. Grant 1994).

Late Wisconsin

- CHAMPLAIN-HUDSON AND NEW ENGLAND LOBES The Lake Champlain-Hudson River lobe and the eastern part of the Ontario lobe advanced southward through New York to the Olean moraine in Pennsylvania and New Jersey (Connally and Sirkkin 1973; Stone, Stanford, and Witte 2002; Braun 2004). The Champlain-Hudson River lobe and the New England lobes (Connecticut Valley, Buzzards Bay, Cape Cod, Georges Bank lobes) also advanced to the southeast toward the Atlantic coast where moraines were formed on Long Island and off the coast of Massachusetts. The initial advance of Late Wisconsin (Woodfordian) ice is estimated to have reached its maximum during 27,000-23,700 B.C. (Stone and Borns 1986; Muller and Calkin 1993; Cadwell and Muller 2004; Ridge 2003, 2004).

Along the Atlantic coast, the Ronkonkoma moraine marks the maximum ice position on eastern Long Island, but it is truncated by the Harbor Hill moraine, which extends westward as the terminal moraine in New Jersey (Stone, Stanford, and Witte 2002); thus, the Ronkonkoma margin is older than the maximum position in New Jersey. In central New Jersey, the maximum Late Wisconsin ice position is indicated by the Madison moraine, while the Budd Lake, Townsbury, Mountain Lake, and Foul Riff moraines mark the maximum ice margin position in western New Jersey. The maximum position is the Olean moraine in Pennsylvania. Proglacial lakes formed adjacent to these ice margins. The age range for these moraines is around 28,000 to 19,600 B.C. After reaching these maximum positions, the Laurentide ice front began to recede, sometimes with small readvances. The Harbor Hill moraine is truncated by a system of recessional moraines (Sands Pont, Fishers Island, Charlestown, Buzzards Bay, Ogdensburg-Culvers Cap, Bloomfield). A younger set of recessional moraines (Wolfs Rock, Sandwich, Sussex) marks the ice position around 18,200 B.C. (Ridge 2004).

In eastern Pennsylvania, in the vicinity of Stroudsburg, mastodon fossils from Marshall’s Creek (Leap’s Bog) have radiocarbon ages of 12,300-12,200 B.C. (McNett 1985). South of Marshall’s Creek, the first (lowest) terrace in the Delaware River valley consists of outwash sediments from the meltwaters from the Laurentide ice margin. The Shawnee-Minisink site is situated along the west side of the Delaware River just north of the moraine formed at the time of the maximum advance of the Late Wisconsin ice (McNett 1985). The locality contains a postglacial depositional sequence. Artifacts were recovered from the second terrace (above the first terrace) in loess overlying outwash gravels, and stratified below clay and sand deposits. The three oldest ages from the loess range 11,000-10,400 B.C. (McNett 1985).

In north-central New Jersey, the Mud Pond and Cherry Ridge moraines mark recessional ice-margins positions. Nearby, at Highland Lake (near Vernon) mastodon remains have been radiocarbondated to 10,900 B.C. (Jepson 1964). In the Hudson Valley of southeastern New York, there are several radiocarbondated Pleistocene faunal localities (Guilday 1968; Fisher and Reilly 1969; Funk, Fisher, and Reilly 1970). West of the Wallkill River, at the Dewey Parr locality, remains of stag-moose, dated to about 11,400 B.C., were found along the interface between peat and marl deposits. East of the Wallkill River, mastodon remains, also found along a peat-marl contact, at the Arborio locality, have been dated to 9700 B.C. (Fisher and Reilly 1969). The best-known Late Wisconsin locality in this area is Dutchess Quarry Cave, where caribou bones dated to 13,000 B.C. were found in deposits that also contained artifacts (Guilday 1968; Funk, Fisher, and Reilly 1970).

Mammoth and mastodon fossils have been recovered in the offshore continental shelf region south of Long Island and off the coast of New Jersey (Richards 1959; Whitmore et al. 1967). Besides mammoth and mastodon, fossils of horse (Equus), tapir (Tapirus), walrus (Odobenus rosmarus), helmeted musk-ox (Symbos) and stag-moose have been found in this region of the continental shelf, sometimes at depths estimated at 160 meters below present sea level.

The southern limit of the Laurentide ice sheet in the New England region along coastal Massachusetts is associated with the Buzzards Bay, Cape Cod Bay and South Channel lobes (Uchupi et al. 1996; Mickelson and Colgan 2004). These lobes extended to Martha’s Vineyard, Nantucket, and onto the continental shelf by about 22,000 B.C. Prior to 19,600 B.C. these lobes had begun to recede, leaving an exposed coastal outwash plain and a glacial lake in the area north of Martha’s Vineyard and Nantucket and south of Cape Cod. In northeastern Massachusetts, crustal depression by glacial loading led to a highstand (highest relative sea level) that was 33 meters higher than present between...
about 15,700 and 15,000 B.C., as reflected by glaciomarine sediments and shoreline features. (This highstand may have reached 130 meters above present sea level in central Maine.) By about 12,000 B.C. the region was deglaciated; crustal rebound led to a drop in relative sea level. The lowstand appears to have occurred later in New Hampshire and Maine; in the Gulf of Maine the lowstand at 55 meters below present sea level dates to around 11,000-10,500 B.C. (Barnhardt et al. 1995). About 12,000 B.C. relative sea levels were around 55 to 43 meters lower than sea level today. Sea level was 55 meters below present levels at 9600 B.C. and had risen to 30 meters below by 8100 B.C. (Uchupi et al. 1996).

The Bull Brook site, in northeastern Massachusetts near Ipswich, west of the Gulf of Maine, is on a kame terrace associated with cross-bedded sands and gravels (Byers 1954). The site contains caribou remains and fluted points. The oldest radiocarbon age for the locality is 8600 B.C. on charcoal, although the relationship between the charcoal and the fluted points is unclear (Byers 1959). Relative sea levels were about 55 meters below present sea levels at the time of the human presence at Bull Brook, if the artifacts date to about 11,000-10,500 B.C. (Pelletier and Robinson 2005). A mastodon was recovered from Ipswich, Massachusetts, below the late Wisconsin high sea-level strandline, and mammoth remains have been recovered along coastal Maine (Odale, Whitmore, and Grimes 1987; Caldwell 1992; Hoyle 1998; Hoyle et al. 2004).

Varves and radiocarbon dates have been used to determine the deglaciation chronology in New England and adjacent areas (Ridge 2004). By about 12,000-11,800 B.C. the position of the Laurentide glacier extended from the Carthage-Harrisonville ice margin and Star Lake moraine east of Ontario in northern New York State to the Bridgeport-Middlesex-Littleton-Bethlehem readvance margins to the east in the Champlain Valley and northern Vermont and New Hampshire. For example, the Middlesex advance buried wood dated to 11,900 B.C. (F.D. Larsen

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**Fig. 6.** Paleogeographic reconstruction about 11,000 B.C., indicating landscape settings potentially associated with Gainey and Bull Brook artifact assemblages (approximately contemporaneous or variants of Clovis artifacts), with glacial lobes, ice caps, moraines, and lakes. The Ulverton-Tingwick moraine south of the St. Lawrence River may date to about 11,100 B.C., which would indicate a position south of what is depicted here and would imply a younger age for Lake Vermont (Dyke, Moore, and Robertson 2003).
2001; Richard and Occhietti 2005). In eastern Maine, the ice position may have been at the Pineo Ridge moraine (Ridge 2004). By 11,400 B.C. (11,600-11,300 B.C.) the ice position apparently was near the United States–Canada border, possibly at the Covey Hill–Enosburg Falls margin or the Frontier moraine (Richard and Occhietti 2005). If this is correct, it would imply a younger deglaciation chronology than depicted in figure 5. In the Connecticut River valley a low-water stage of Lake Hitchcock may have been in existence until after 10,500 B.C. and thus could be contemporaneous with human presence in the region (Ridge 2003).

The deglaciation of New England left a landscape that was inhabited by Rancholabrean mammals. For instance, the Ivory Pond mastodon, found in a bog in the Berkshire Mountains of western Massachusetts, dates to around 11,400 B.C. (Moeller 1984). Deglaciation of southwestern New Hampshire exposed the landscape around the Whipple site situated along the Ashuelot River (Curran 1984). It is on a terrace or deltaic deposit. Hardwood charcoal from the locality provided an age range of 11,000 to 10,000 B.C. (C.V. Haynes et al. 1984). The Whipple site contains caribou remains. The Vail site in northwest Maine, adjacent to New Hampshire and southern Quebec, is situated in a knob and kettle deglaciated landscape, next to an abandoned channel of the Magalloway River (Gramly and Rutledge 1981; Gramly 1982). Radiocarbon ages on charcoal from the locality range from 11,100 to 10,200 B.C. In northern New Hampshire, the Littleton-Bethlehem moraines may indicate the ice position around 11,800 B.C. (Thompson, Fowler, and Dorion 1999; W.B. Thompson et al. 2002). By about 11,400 B.C. the ice position had melted back at least to the position of the Frontier moraine.

Moraines were formed in the interval after 11,500 to 10,400 B.C. during deglaciation of the Appalachian region of Quebec, south of the present-day Saint Lawrence River and adjacent to Vermont, New Hampshire, and Maine (Richard and Occhietti 2005). Northwestward recession of the Laurentide ice from the Frontier moraine is marked by the Ditchfield moraine, the Cherry River–East Angus moraine, the Mount Ham moraine, and—by about 11,100 B.C.—the Ulverton-Tingwick moraine (Parent and Occhietti 1999; Occhietti et al. 2001, 2004). Lake Vermont is associated with the Mount Ham and Ulverton-Tingwick moraines. The Saint Raphael moraine marks the position of the ice along the northeastern Appalachian piedmont just before marine waters extended into the region (fig. 10).

By 10,800-10,500 B.C. the ice margin (St. Maurice lobe) was present at the Saint Narcisse moraine north of the present location of the Saint Lawrence River (Cummings and Occhietti 2001; Richard and Occhietti 2005). The Saint Narcisse and Saint Edouard moraines have been attributed to climatic fluctuations associated with the Younger Dryas (LaSalle and Shilts 1993). This deglaciation chronology, based on radiocarbon dating of terrestrial plants is younger (later) than the ice retreat model of Dyke, Moore, and Robertson (2003), depicted in figures 5 and 6, which relied on dates from marine shells. Although the Laurentide ice was north of the present Saint Lawrence River, separate local glaciers were present in northern Maine and adjacent areas of Quebec (including the Gaspe Peninsula), and New Brunswick, Nova Scotia, and Newfoundland.

Many proglacial lakes were formed along the edge of retreating Laurentide ice margins (LaRocque, Dubois, and Leblon 2003). During initial deglaciation the most prominent of these lakes were in the Hudson and Connecticut river valleys. The glacial maximum position in southern New York is marked by the Ronkonkoma moraine on Long Island (Cadwell, Muller, and Fleisher 2003). As the Laurentide ice margin began to recede from this margin, Lake Albany formed in the Hudson valley. Lake Albany may have an age range of about 18,200-11,700 B.C. (Dineen and Hanson 1992). An estimate for the Hidden Valley moraine is based on wood dated to 12,300 B.C. and peat overlying mastodon remains dated to 12,300 B.C. (Dineen and Hanson 1992). Musk-ox from glacial lake sediments date to 11,200 B.C. Lake Hitchcock formed as ice melted from the Connecticut Valley with the melting of the Connecticut Valley lobe (Dickson and Colgan 2004). The lake probably existed about 16,800-11,500 B.C. Radiocarbon dates on plant fragments indicate the beginning of glacio-eustatic postglacial rebound around 14,800 B.C. (J.R. Stone and G.M. Ashley 1995).

Recession from the Valley Heads moraine in central New York led to the development of proglacial lakes (Cadwell, Muller, and Fleisher 2003; Cadwell and Muller 2004). When the ice melted from westernmost New York, waters flowed from Lake Warren, which was in the Lake Erie basin. As the ice receded from the Erie-Ontario Lowlands, glacial Lake Iroquois was formed (Pair and Rodrigues 1993). It expanded in size as the ice receded into the Saint Lawrence Lowland. As the ice receded northward into Quebec, the Saint Lawrence River drainage was connected to the sea. A rise in the sea level led to the flooding of the Saint Lawrence valley by the Champlain Sea. The marine waters extended westward to the eastern side of the Ontario basin (fig. 7).

During the Sherbrooke phase, glacial Lake Memphremagog formed south of the Mount Ham moraines and drained into glacial Lake Vermont (Parent and Occhietti 1999). As the glacial front receded from western New York State, water was able to drain from Lake Warren (Muller and Prest 1985; Cadwell and Muller 2004). Glacial retreat from the Ontario basin led to the expansion of glacial Lake Iroquois into the Saint Lawrence Lowland. Melting of the front to the Saint Raphael moraine led to the coalescence of glacial Lakes Iroquois, Vermont, and Memphremagog into proglacial Lake Candona (Parent and Occhietti 1999). When the ice margin melted back from the Saint Raphael moraine, the Saint Lawrence Lowland was connected to the sea (Occhietti et al. 2001). Eustatic sea level rise led to the expansion of marine waters of the Champlain Sea around 11,100 B.C. into the Ontario basin (Richard and Occhietti
2005). The Champlain Sea marine deposits are interstratified with and overlie the Saint Nicolas phase drift (LaSalle and Shilts 1993; Cummings and Occhietti 2001). Based on fossils of Arctic fox (Alopex lagopus) dated to 10,800 B.C. from Lafleche Cave in southern Quebec, tundralike landscapes may have existed in this region during the time of the Champlain Sea (Harington 2003).

Maine was deglaciated between 15,800 and 9600 B.C. (Newman, Genes, and Brewer 1985; W.B. Thompson 2001; Borns et al. 2004). As the ice-front retreated, there was a marine transgression by waters of the Atlantic that flooded the previously glaciated landscapes. The initial ice-margin retreat appears to have occurred about 15,400-14,000 B.C. (fig. 2) with glaciers in contact with the sea (Kaplan 1999). The Pond Ridge moraine marks the ice-margin position at perhaps 14,300 B.C. Around 14,000 B.C., relative sea levels may have been 70 meters higher than today (Barnhardt, Belknap, and Kelly 1997). The Androscoggin moraine in western Maine and eastern New Hampshire marks the ice front at around 13,600-12,700 B.C. (W.B. Thompson 2001; W.B. Thompson et al. 2002). Woolly mammoth remains found redeposited in shallow marine deposits date to about 12,300 B.C. in southwest Maine (Borns et al. 2004; Hoyle et al. 2004). By 11,000 B.C. the relative sea levels in southeast Maine may have been 35 meters below present levels. A lowstand of 55 meters below the present occurred soon after, perhaps around 10,800 B.C. By 8400 B.C. sea levels had begun to rise to a depth of about 20 meters below the present (Barnhardt, Belknap, and Kelly 1997).

The Younger Dryas period, 11,000-9600 B.C. (Borns et al. 2004), is reflected in environmental changes but not a glacial readvance in northern Maine; the Mars Hill moraine indicates the approximate position of the Laurentide glacial ice during that time. There may have been a readvance of the northern Maine ice cap during the Younger Dryas, based on radiocarbon dates of around 10,500 B.C. (fig. 7).

If global sea levels were at times 60-180 meters (Bloom 1983) lower during the Late Wisconsin, presently submerged regions of the Atlantic continental shelf would have been exposed and potentially available for migration...
and habitation. The Gulf of Maine, situated northeast of Massachusetts and east of Maine, and the adjacent area of the Bay of Fundy, between New Brunswick and Nova Scotia, would have been exposed during intervals of lower sea levels. On the continental shelf in Massachusetts Bay, northwest of Boston (and south of the Bull Brook site), a mastodon tooth dated to 11,000 B.C. was recovered at about 40-45 meters below the present level, and a mammoth tooth (attributed to *Mammuthus jeffersonii*) dated to 10,800 B.C. was recovered in 55-meter deep water (Odale, Whitmore, and Grimes 1987).

- **Glaciation in the Maritime Provinces** The Canadian Maritime region was completely covered by glacial ice during the late Wisconsin maximum advance. Besides the influence of Laurentide ice, there were several local sources of glaciation on New Brunswick, Nova Scotia, and Newfoundland (D.R. Grant 1989). Glaciers centered in Maritime Canada have been designated the Appalachian Glacier complex. An ice center may have existed in Prince Edward Island and northeast New Brunswick during the early part of the Late Wisconsin (Stea et al. 1998, 2001; Stea 2004). Designated as the Escuminac phase, glaciers spread northward onto the Magdalen Shelf and southward as the Chignecto glacier. This has been dated, based on radiocarbon shell ages, to 22,900 B.C. There may have been a contemporaneous glacier in the highlands of New Brunswick.

- As the Escuminac glacier center dissipated, another glacial center formed over Nova Scotia during the Scotian phase, with shell dates indicating an age of about 19,000 to 16,200 B.C. Ice from this ice center expanded into the Bay of Fundy and the Atlantic coast, forming the Scotian Shelf End Moraine (King and Fader 1986). The Chignecto phase of glaciation with several local ice centers on Prince Edward Island, Nova Scotia, and New Brunswick may date to after 14,100 B.C. (Stea 2004). After an interval of recession, there appears to have been a reactivation of glaciation in New Brunswick and Nova Scotia during the Younger Dryas. The Younger Dryas appears to be contemporaneous with the Collins Pond phase readvance, based on radiocarbon dates on wood in a paleosol buried by till of 10,900 and 10,800 B.C. (Stea 2004). A set of radiocarbon
ages on wood from localities in central and southwestern New Brunswick (Joe Lake, Splan Pond) and across Nova Scotia (Lac à Magie, Little Lake, Chase Pond) suggests the initiation of a cooling event around 10,900 B.C. that ended around 9600 B.C. (Mayle, Levesque, and Cwynar 1993).

In New Brunswick, major Late Wisconsin deglaciation began around 15,400 B.C. (Foisy and Prichonnet 1991). Deglaciation in the coastal areas may have begun around 14,600 B.C. and been completed by about 12,000 B.C., except for the possible persistence of local ice caps. Several of the moraines and ice margins found in New Brunswick have been correlated with ice positions in Maine. The Saint John moraine consists of geomorphic features and deposits indicating the position of the ice margin as a continuation of the Pineo Ridge moraine in Maine estimated to date to 14,300 B.C., based on radiocarbon measurements on shells (Rampton et al. 1984; A. Seaman 2004). There is another moraine on Campobello Island that may be a continuation of the Pineo Ridge moraine dated to 13,600 B.C. In the Saint John region, the Sheldon Point moraine has been dated 13,200-11,600 B.C. (Broster 2005). An active ice cap in the Caledonian Highlands may have blocked the Saint John River, leading to a glacial lake during the Madawaska phase of glaciation, possibly around 12,000 B.C. (Rampton et al. 1984; Stea and Mott 1998). There is evidence of a glacial advance during the Younger Dryas near Todd Mountain (Lamothe 1992). The residual ice cap over the Caledonian Highlands may have led to the formation of glacial Lake Acadia in the Saint John River valley around 11,000 B.C. (A. Seaman 2004; Dickinson 2005).

Changes in the paleogeography and sea levels in the vicinity of Nova Scotia and south of New Brunswick are associated with the regional deglaciation framework. As the Laurentide ice receded from the Gulf of Maine, between 19,600 and 16,400 B.C., the ice center on Nova Scotia was isolated. This led to the glacial phase associated with the Scotian Shelf End Moraine complex (the Scotian Ice Divide). Melting of the Scotian ice mass led to several distinct ice caps centered in the Nova Scotia highlands and mountains. The coastal regions that had been depressed by glacial ice rebounded, and relative sea levels in the Bay of Fundy rose to about 30 meters above present sea level.

![Figure 9: Paleogeographic reconstruction about 6900 B.C. (early Holocene). The Laurentide ice margin is fronted by Lake Ojibway (Dyke, Moore, and Robertson 2003).](image-url)
After about 14,800 B.C., coastal areas around the Bay of Fundy appear to have been ice-free, although there appears to have been a readvance reflected by the Gilbert Lake moraine associated with the Chignecto phase of glaciation. Local ice centers and glacial lakes persisted 12,800-11,000 B.C. After 11,000 B.C. Nova Scotia was ice-free, except for small remnant glaciers in the highlands. These glaciers were reactivated around 10,800 B.C. This reactivation led to the blocking of drainages and the formation of glacial lakes (Stea and Mott 1998).

Between 11,600 and 11,000 B.C., the relative sea level in the Gulf of Maine had dropped to 60 meters below present level, providing a potential corridor for biotic migration (Stea et al. 1994; Barnhardt 1995). Mastodon remains have been recovered from Nova Scotia and adjacent coastal shelf areas (MacDonald 1968; Cooke, Harington, and Sollows 1993). The Debert site overlies an ablation till and possible eolian and outwash deposits in Nova Scotia. At the time humans were at the site, snowfields may have been present immediately to the north in the Cobequid Mountains and regionally there may have been active glaciation with ice caps on the Nova Scotian peninsula and possibly Cape Breton Island (Borns 1965). Radiocarbon dates on charcoal from hearth features average 10,500 B.C.; the range is 11,000-9900 B.C. (G.F. MacDonald 1968), which
suggests that the locality may have been inhabited during the Younger Dryas. Periglacial conditions may have been present in Nova Scotia during the Younger Dryas (Stea and Mott 1989, 1998).

Newfoundland also had an ice cap. It may have been present during the Middle Wisconsin and appears to have persisted to about 8900 b.c. (Dyke et al. 2002). Some coastal areas may have been deglaciated by about 12,000 b.c. (fig. 4). The Ten Mile Lake moraine may mark a Younger Dryas advance of the Newfoundland Ice Cap (D.R. Grant 1992, 1994). During the Last Glacial Maximum, lowered sea levels would have exposed the coastal region of the Grand Banks southeast of Newfoundland, as well as the Gulf of Saint Lawrence between Newfoundland and New Brunswick and Nova Scotia. To the north of Newfoundland, the ice-edge margin may at times have been a region of high biological productivity (Bradley and Stanford 2004). Mastodon and mammoth fossils have been recovered from the Grand Banks off the coast of Newfoundland (Whitmore et al. 1967; G.F. MacDonald 1968).

Younger Dryas and Glaciation in Ontario, Quebec, and Labrador

In southern Quebec, the Saint Narcisse moraine appears to be the ice position between 10,800 and 10,500 b.c. during the early part of the Younger Dryas, while the Mars-Batiscan-Saint-Maurice moraine system may represent the ice-margin during the late Younger Dryas, around 10,500-9700 b.c. (fig. 10) (McDonald and Shilts 1971; La Salle and Elson 1975; Dubois and Dionne 1985; LaSalle and Chapdelaine 1990; Richard 1994; Cummings and Occhietti 2001; Occhietti et al. 2004; Richard and Occhietti 2005). Moraines correlated with the Saint Narcisse moraine indicate the position of the Laurentide ice margin elsewhere in Quebec and in Labrador (e.g., the Baie-Trinité, Bradore, and Belle Amours moraines). There are several moraines north of these Younger Dryas-age moraines that reflect the margins of the melting Laurentide ice sheet during the Holocene.

South of Hudson Bay, in western Ontario, the Marquette advance marks the position of the Laurentide ice around 9600 b.c. The Cochrane margin, associated with the presence of Lake Ojibway, dates to about 7400 b.c. (Fullerton 1994). In western Quebec another Holocene deglaciation feature is the Harricana moraine, estimated to date to about 6900 b.c. (Dubois and Dionne 1985). On the southeast side of the Hudson Bay the Sakami moraine is a prominent feature that may postdate the draining of Lake Ojibway and the flooding of the region by marine waters of the Tyrrell Sea around 6900 b.c. (based on radiocarbon measurements on shells).

As ice receded from the Baie-Trinité moraine in eastern Quebec (correlated with the Saint Narcisse moraine to the west), a possible stillstand (state of equilibrium between advancing and melting) led to the formation of North Shore moraine (correlated with the Aguanus-Kenamu moraine) perhaps around 9100-8900 b.c. (Dubois and Dionne 1985; Occhietti et al. 2004). In Labrador, this could be correlated with the Sebaskachu-Little Drunken moraines, although these margins may be slightly younger, 8100-6900 b.c. (Occhietti et al. 2004). Using radiocarbon measurements on shells, the minimum age of the marine limit and deglaciation for the southernmost Labrador coast was estimated at 15,400-11,000 b.c., while the marine limit for the northernmost Labrador coast was estimated at 8100 b.c. (Clark and Fitzhugh 1992). The age of deglaciation as estimated by the marine limit should provide an indication of the earliest possible availability of habitable landscapes. A small remnant ice cap persisted east of Hudson Bay until after 3800 b.c.

Conclusion

Geomorphic features and sedimentary deposits provide the basis for developing a geologic framework for the glaciated region of eastern North America. Stratigraphic relationships and radiocarbon measurements provide critical information on the Late Pleistocene evolution of landscapes in this region. During the Middle Wisconsin the southern Great Lakes region, New England, and the Canadian Maritimes appear to have contained nonglaciated landscapes, while Ontario, Quebec, Labrador, and Newfoundland were covered in ice. Glacial ice expanded southward from Ontario, reaching a maximum position by 22,000-21,000 b.c. or earlier south of the present-day Great Lakes basins. Lobes of the Laurentide ice sheet and several independent ice centers were present in New England and the Canadian Maritimes. The Laurentide ice sheet extended southward into northern Pennsylvania, New Jersey, and southern New York and expanded westward onto the Atlantic continental shelf.

Between 15,400 and 11,400 b.c. the ice margins had receded northward, leading to the presence of glacial lakes and deglaciated landscapes in the southern Michigan and Huron basins. The glacial margin was north of the Erie basin by about 12,800 b.c. and north of the Ontario basin by 12,000 b.c. Glacial ice persisted in the Lake Superior basin about 9400-8900 b.c. but was apparently north of the rest of the Great Lakes by about 11,000 b.c. Although the Laurentide margin seems to have been north of the Saint Lawrence River by about 10,500 b.c., separate ice centers still existed in northern New England and the Canadian Maritimes. Independent centers of ice may have persisted in Maine, New Brunswick, and Nova Scotia after 9600 b.c.

Glacial lakes were contemporary with the ice margins. These have been used to estimate the age of archeological occurrences and identify landscape contexts that may have been inhabited by late Pleistocene human groups.
Deglaciation also led to significant changes in relative sea levels along the Atlantic coast. At times, lowered sea levels would have exposed land now submerged along the continental shelf of the Atlantic Ocean. Fossils of Rancholabrean fauna—including mammoths and mastodons—have been found in these coastal areas. Rancholabrean fauna in the region can be used to constrain the timing of deglaciation and to infer the presence of landscapes that could have been inhabited by Late Pleistocene human groups. There are some indications of a human response to changing environmental conditions associated with the Younger Dryas (cf. Newby 2005). The geologic framework of eastern North America indicates the presence of deglaciated landscapes with Rancholabrean fauna prior to 9600 B.C. At that time, the Laurentide ice margin was mostly north of the Great Lakes and was north of the Saint Lawrence River, and there were local glaciers present in northern New England and the Atlantic coastal provinces of Canada. By the end of the Pleistocene in eastern North America, deglaciated landscapes extended from the Great Lakes to the Atlantic coast.