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POST-VALLEY HEADS DEGLACATION OF THE ADIRONDACK MOUNTAINS AND ADJACENT LOWLANDS

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INTRODUCTION

In the early 1900s, Herman LeRoy Fairchild (1909, 1912, 1919) developed a regional deglacial chronology for New York State in a series of New York State Museum Bulletins. Fairchild's reports were accompanied by a collection of maps that encompassed the entire state at a scale of approximately 1:1,267,200 and depicted the locations of glacier ice, proglacial lakes and major rivers at different stages of ice recession. Fairchild's work was published prior to the advent of radiocarbon dating techniques and did not have supporting numerical age control, but his synthesis was a masterful work and continues to be the most comprehensive treatment of deglacial history in New York.

The regional deglacial chronology of the Adirondack region presented here is less ambitious in scale than Fairchild's and was inspired by recent work on glacial stratigraphy in the lowlands peripheral to the Adirondack Upland (Figure 1). There is also now an improved understanding of the chronology of deglacial events in New York (Cronin, Rayburn, Guilbault, Thunell, and Franzi 2012; DeSimone et al. 2008; Franzi, Rayburn, Knuepfer, and Cronin 2007; Parent and Occhietti 1988; Rayburn, Cronin, Franzi, Knuepfer, and Willard 2011; Rayburn, Franzi, and Knuepfer 2007; Rayburn, Knuepfer, and Franzi 2005; Ridge 1997; Ridge, Brennan, and Muller 1990; Ridge, Franzi, and Muller 1991; Ridge and Franzi 1992; Stanford 2009) and New England that has been augmented by the North American Varve Chronology and radiocarbon ages (NAVC of Ridge 2016; Ridge et al. 2012); formerly the New England Varve Chronology (Antevs 1922, 1928; Ridge 2003, 2004) and paleomagnetic correlations (Ridge 2004; Ridge et al. 1990). Our intent is to summarize the current state of understanding of the style and chronology of deglaciation, identify areas where information is scarce or issues should be addressed, and encourage a new wave of scientific exploration and discovery in the Adirondack region.

***Figure 1:** Physiographic regions of northeastern New York. The red stippled area is the Saranac Intramontane Basin (SIB; Buddington 1953).*

Regional Physiography

The physiography of the Adirondack Upland and surrounding regions had a profound effect upon ice movements, patterns of ice recession, and the nature of glacial foreland and proglacial sedimentary environments during the last glacial maximum (LGM). The Adirondack Upland is a domal uplift about 250 km in diameter that is underlain predominantly by high-grade Mesoproterozoic metamorphic rocks (Figure 1) and surrounded by peripheral lowlands carved into lower Paleozoic sedimentary rocks. The High Peaks region contains 43 mountain peaks that lie above 1,219 meters (m) in elevation, including the highest summits in New York, Mt. Marcy (1,629 m) and Algonquin Peak (1,559 m). The region is characterized by rugged terrain where local relief may exceed 1,000 m (Cressey 1977). The High Peaks are surrounded by lower terrain in which many peaks, especially in the south-central highlands, exceed 600 m in elevation but generally local relief is less than 300 m. The Saranac Intramontane Basin (SIB; Buddington 1953) is an ovate, northeast–southwest trending structural basin bounded by low mountains, generally less than 800 m elevation that encompasses the headwaters of the upper Saranac, Salmon, St. Regis, and Raquette drainage basins. The region is characterized by the occurrence of many large lakes and well over a hundred smaller ponds that are developed in deposits of ice-contact stratified drift and glacial-fluvial outwash.

The Adirondack Mountains: Obstruction or Source of Glacial Ice

An important question is how the Adirondack Upland influenced the regional pattern and timing of ice recession. Specifically, we shall address when and how the Adirondacks impeded regional ice flow and focused flow into lobes in adjacent lowlands. The lowland ice lobes impounded large proglacial lakes that drained in a succession of lowering levels, often punctuated by high-magnitude breakout floods, as lower outlets were uncovered with ice recession. The formation of ice lobes and the chronology of proglacial lake succession in adjacent lowlands was a central theme of Fairchild's (1909, 1912, 1919) deglacial reconstructions. However, many of the lobes he envisioned had surface gradients that were too low to be realistic. Alternatively, the Laurentide Ice Sheet (LIS) may have overtopped parts of the Adirondacks and entered adjacent lowlands through much of the deglaciation history, especially from the southern and southwest Adirondacks into the Mohawk Valley.

Our reconstructions address how the Adirondack Upland influenced the timing and pattern of ice recession in the region. The Adirondacks are critical in this regard because of their position at the transition between two glacial regimes with very different ice sheet dynamics. To the east is the rugged terrain of New England, not unlike the Adirondacks, that must have impeded ice flow and triggered a steeper ice sheet profile. This part of the LIS may have been more stable than areas to the west in regards to its sensitivity to climate events and glaciological changes, and it was minimally influenced by calving in glacial lakes. West of the Adirondacks, and in the surrounding lowlands, is the eastern limit of a Midwestern style of ice sheet that advanced across a smoother glacial bed and was floored by clayey sediment (e.g., Eyles and Doughty 2016). The ice sheet likely had faster flow, a gentler surface slope, and at times may have formed ice streams (Briner 2007; Hess 2009; Kerr and Eyles 2007). In these areas, the LIS was very sensitive to climate events and calving in large lakes and glacial readvances covered much greater distances. One may view the Adirondack region as one in which slow upland ice anchors the system and is forced to work in tandem with streaming lowland ice.

An additional issue is whether or not local alpine glaciers occupied parts of the High Peaks following the retreat of the LIS. Early proponents of the concept (e.g., Alling 1918, 1920; Johnson 1917; Ogilvie 1902) alluded to morphological evidence, such as cirques and moraines, as evidence for small alpine glaciers in a few of the highest Adirondack headwater valleys, but Fairchild (1913, 1932) considered a late Pleistocene phase of alpine glaciation to be unlikely. Craft (1969, 1976, 1979) concluded that landforms and sediments in multiple Adirondack valleys were the products of small alpine glaciers and suggested that some of the alpine glaciers may have extended down valley for several to more than 10 kilometers (km). However, more recent work on proglacial lake successions in the AuSable and Boquet valleys (Deimer and Franzi 1988; Franzi 1992; Franzi, Barclay, Kranitz, and Gilson 2015; Franzi et al. 2007; Rayburn et al. 2007) places significant constraints on the extent of possible former Adirondack alpine glaciation. Based on all the available evidence, we consider it unlikely that independent local glaciers fed by snowfall in the High Peaks region developed during the most recent deglaciation. Rather, the headwater valleys of the Adirondack High Peaks region were ice-free or occupied by remnant blocks of continental ice or ice-marginal lakes as the LIS wasted and receded from the region.

PALEOGEOGRAPHIC RECONSTRUCTIONS

Reconstructing the paleogeography of the Adirondack and adjoining regions depends upon stratigraphic analysis of glacial, lacustrine, and marine deposits and upon the spatial distributions of ice marginal and proglacial lake deposits and landforms. Deglacial reconstructions are often hampered by complex or ambiguous stratigraphic relationships, poor preservation or lack of exposure of physical evidence, and outdated or unavailable stratigraphic documentation over large areas. Consequently, paleogeographic reconstructions must extend from areas where the stratigraphy is well documented into adjacent areas, which often have little or no stratigraphic control.

The paucity of detailed contemporary glacial-stratigraphic information is particularly acute throughout most of the Adirondack Uplands, although the works of Craft (1969, 1976, 1979), Gurrieri and Musiker (1990), Muller, Sirkin, and Craft (1993) are notable exceptions. The paleogeographic reconstructions presented here are derived from detailed field studies of glacial foreland deposits and landforms in the western Mohawk Valley (e.g., Muller, Franzi, and Ridge 1986; Ridge 1997; Ridge and Franzi 1992; Ridge et al. 1991), the Ontario and St.

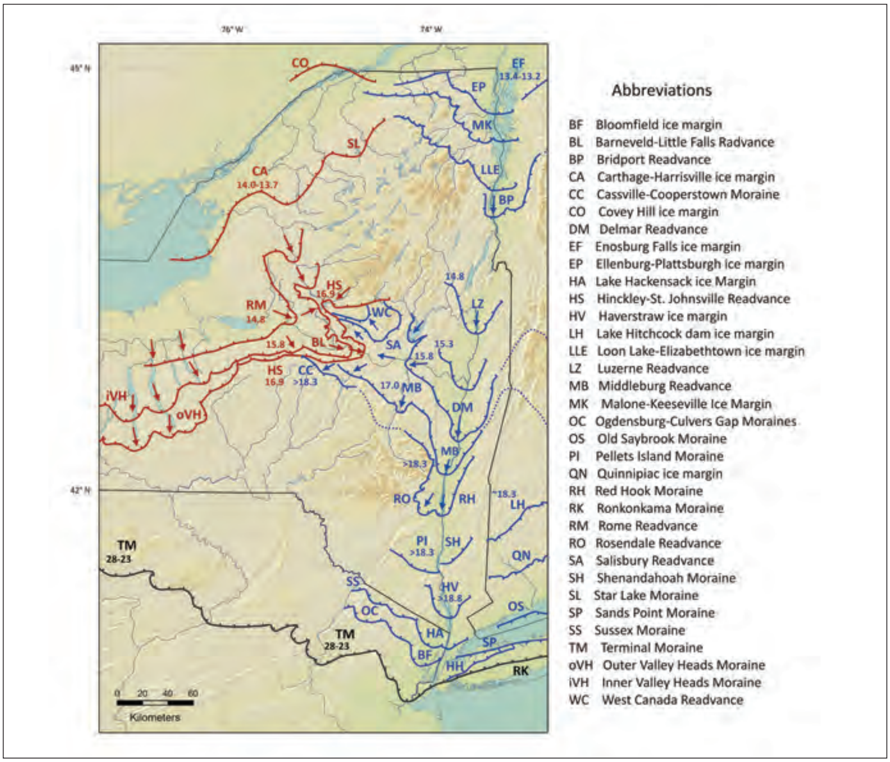
Lawrence lowlands (e.g., Muller and Prest 1985; Occhietti, Parent, Shilts, Dionne, Govare, and Harmard 2001; Pair, Karrow, and Clark 1988; Pair and Rodrigues 1993; Richard and Occhietti 2005), the Champlain Lowland (Chapman, 1937; Cronin et al. 2012; Denny 1974; Franzi et al. 2007; Parent and Occhietti 1988; Rayburn et al. 2011, 2007, 2005) and the upper Hudson drainage basin (Connally and Sirkin 1973, 1971; DeSimone et al. 2008; DeSimone and LaFleur 2008; Dineen and Hanson 1992; Stanford 2009).

Extrapolation of ice margins from these areas into the Adirondack Uplands was guided by the use of glacier-profile models (Benn and Hulton 2010; Schilling and Hollin 1981) along well-constrained ice flow lines and the projection of ice-surface equipotential lines into regions with less stratigraphic control. Franzi (2002, unpublished) and (Franzi et al. 2015) used this technique to correlate ice margin positions in the AuSable, Boquet, and Saranac valleys of the northeastern Adirondack Upland. In most instances, these techniques produced realistic correlations but were most effective when used in combination with field evidence to correlate well developed ice-marginal deposits and landforms in adjacent valleys.

Shorelines for the regional proglacial lakes Albany, Coveville, Fort Ann, and Iroquois, as well as the upper marine limit of the Champlain Sea were recreated by fitting a trend surface to the surface elevations of shoreline deposits (compiled by Rayburn 2004 and DeSimone 2016, personal communication) and intersecting the lake planes with a 50 m digital elevation model (DEM) for New York and adjacent parts of New England and southeastern Canada. Smaller Adirondack proglacial lake shorelines were approximated by projecting a first-order trend surface from their presumed lake outlet elevation at a northward gradient of 0.75 m/km (Denny 1974; Franzi 1992; Franzi et al. 2015; Rayburn et al. 2005).

Numerical ages for the ice front positions across central and eastern New York that appear on Figure 2 were transferred to New York from New England as follows. A precise chronology of deglaciation was developed in New England based on a calibrated 5659-year varve chronology, varve counts, and radiocarbon ages tied to the varves (Ridge et al. 2012), as well as cosmogenic ^{10}Be ages of glacial boulders (Balco, Briner, Finkel, Rayburn, Ridge, and Schaefer 2009). Correlations between ice margins and events in New England and New York, and thus the transfer of numerical ages to New York, are based on the correlation of paleomagnetic records of remanent declination in fine-grained, laminated glaciolacustrine deposits in both areas (Ridge 2003, 2004; Ridge et al. 1990). It should be emphasized that this correlation has a greater uncertainty than just the uncertainty of radiocarbon ages or their calibration. Paleomagnetic declination values have an uncertainty of up to +4° and declination records have a time lag of up to 2-3 centuries between the two areas dependent on the rate of westward drift of dipole and non-dipole components of the geomagnetic field through time. In other words, like today, the two areas rarely have exactly the same geomagnetic declination at the same time, but are generally within 4-5°. Ages for ice margin positions in both New England and New York are constantly revised as new radiocarbon ages, improved radiocarbon calibration, improved calibration of the varve chronology in New England, and new paleomagnetic data become available (cf. Ridge 2016, for the most recent update). All radiocarbon dates in this manuscript are calibrated using Calib 7.0.4 and the IntCal13 data set (Reimer et al. 2013; Stuiver and Reimer 1993), and they are presented at 2σ uncertainty. Calibrated ages alleviate radiocarbon age variability due to temporal variations in ^{14}C production rates, reservoir effects, variable isotopic fractionation, and contamination. The reader is referred to Reimer et al. (2013) or Stuiver and Reimer (1993) for further explanation of radiocarbon calibration methods.

***Figure 2:** Deglacial chronology for New York showing the locations of principal moraines and other ice marginal deposits and landforms (modified after Ridge 2003, 2016). Numbers indicate ages in calibrated (U-TH) kyr. B.P. Arrows indicate ice-front positions that are the limits of glacial readvances.*



REGIONAL SYNTHESIS

Early Deglaciation from the Last Glacial Maximum (LGM)

The LIS reached its last glacial maximum (LGM) position in southern New York between 28-23 cal. kyr B.P. (Ridge 2003, 2004, 2016) (Figure 2). The Adirondack region probably influenced subglacial ice flow, but it is likely that ice was actively flowing over the highest peaks (Ogilvie 1902; Taylor 1897) at the LGM, although, as on high peaks in New England, it may not have significantly eroded high elevation land surfaces (Bierman, Davis, Corbett, Lifton, and Finkel 2015). Deglacial drawdown of the LIS into major lowlands caused thinning of ice over upland regions and lobation of the ice front. Proglacial Lake Albany (LaFleur 1968; Woodworth 1905a) fronted the receding ice margin in the Hudson Lowland during the initial stages of deglaciation and expanded northward with ice recession. The general northward recession was punctuated by sometimes extensive but short-lived readvances in the Hudson and Mohawk lowlands. The southwestern Adirondacks probably first emerged from the ice about 18-19 cal. kyr B.P. and divergence into separate ice streams began at this time (Ridge et al. 1991). Ice flow into the Mohawk Valley eventually diverged into eastern (Mohawk) and western (Oneida) ice lobes as ice thinning and recession continued. Proglacial lakes at the ice margins and between the opposing Mohawk and Oneida ice lobes eventually gave way to the first period of free eastward drainage and fluvial conditions (Figure 3) as the ice margin evacuated the eastern end of the Mohawk Valley. This fluvial interval is represented by an erosional unconformity and fluvial gravel in western Mohawk Valley stratigraphic sequences (Muller et al. 1986; Ridge 1997; Ridge et al. 1991) that correlate with the Erie Interstadial (Mörner and Dreimanis 1973) or Erie Phase (Karrow, Dreimanis, and Barnett 2000) in eastern Great Lakes stratigraphic nomenclature. The fluvial gravel and unconformity were later buried by deposits of glacial readvances and lakes during the Valley Heads glaciation as discussed below.

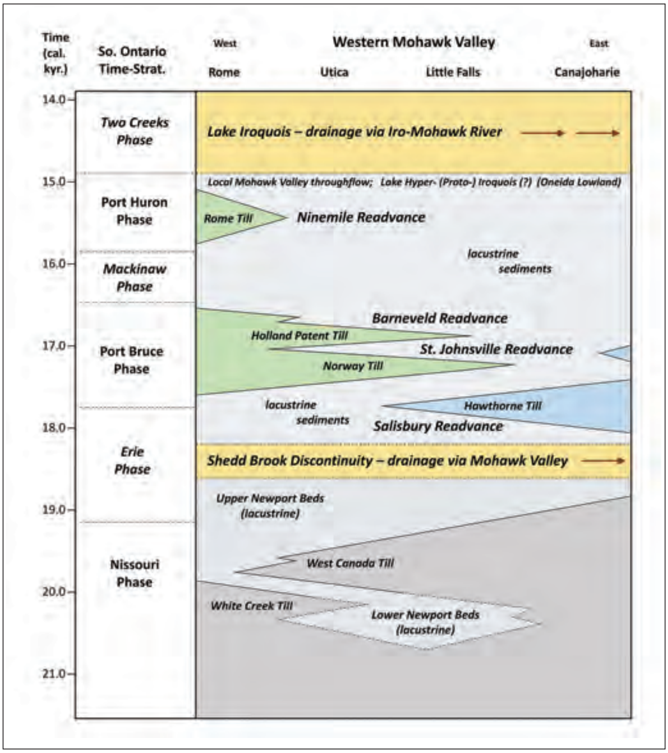


Figure 3: *Time-distance diagram showing the lithologic relationships of late Wisconsinan stratigraphic units in the western Mohawk Valley (after Muller et al. 1986).*

Valley Heads Readvance and the St. Johnsville-Canajoharie Ice Margin

In central New York, glacial ice from the Ontario Lowland pushed southward into the Finger Lakes region and reached its maximum extent in central New York at the outer Valley Heads Moraine. The Valley Heads Readvance marked the end of the Erie Interstadial and eastward fluvial drainage through the Mohawk Valley ceased. Drainage from the glacier and runoff from the unglaciated uplands at the Valley Heads maximum was diverted southward to the Susquehanna drainage basin (Figure 4). The Valley Heads moraines that formed along the Appalachian escarpment from the Finger Lakes region eastward to about Oneida consist of thick heads of outwash and ice-contact stratified drift in the valleys. These deposits form the present-day drainage divide between the Susquehanna and St. Lawrence drainage basins (Cadwell and Muller 2004).

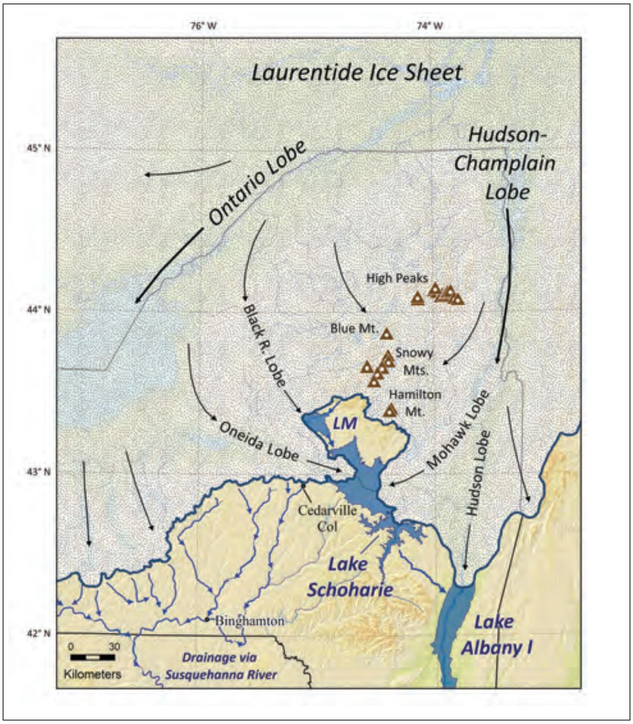


Figure 4: *Paleogeography of glacial ice lobes and proglacial lakes in northeastern New York at the maximum extent of the Valley Heads Readvance in central New York. The abbreviation LM refers to proglacial Lake Miller in the upper West Canada Creek Valley. The brown triangles indicate the possible locations of nunataks as determined by glacier-profile modeling.*

The Valley Heads readvance in the Finger Lakes region generally correlates with asynchronous readvances of the Mohawk and Oneida lobes in the western Mohawk Valley (Muller et al. 1986; Ridge 2003, 2004; Ridge, Brennan, and Muller 1990; Ridge and Franzi 1992). The Mohawk lobe advanced westward as the Salisbury Readvance (SA on Figure 2), fed by ice streaming through Sacandaga trough in the southeastern Adirondacks. Proglacial lakes were re-established in the western Mohawk Valley, and these lakes deepened and widened progressively as lower eastern outlets were blocked by the westward advancing Mohawk Lobe. The highest proglacial lake level, Lake Cedarville, drained southward through Cedarville col to the Unadilla River in the Susquehanna drainage basin (Ridge 1985; Ridge and Franzi 1992; Ridge et al. 1991). The Mohawk lobe receded from its Salisbury Readvance maximum as ice from the Oneida lobe continued its eastward advance into the Mohawk Valley as the Hinckley-St. Johnsville Readvance (Figure 4; SH on Figure 2; Ridge and Franzi 1992). Early recession of the Mohawk lobe may have been triggered by increased calving caused by rising lake water in the western Mohawk Valley, while the Mohawk Lobe advanced westward into a widening valley. Advancing Oneida lobe ice experienced the same lake level rise but instead was advancing eastward into a narrowing valley that diminished calving potential and appears to have stabilized the advance of the lobe. Oneida ice overrode varved lacustrine deposits that were deposited over older till deposits of Mohawk provenance (Hawthorne till of Figure 3), indicating that a short period of lacustrine sedimentation intervened between the Mohawk and Oneida advances (Ridge 1985; Ridge et al. 1990; Ridge and Franzi 1992). Proglacial Lake Cedarville may have fronted the advancing Mohawk and Oneida lobes during the Salisbury and Hinckley-St. Johnsville advances in the Mohawk Valley, but this lake lasted only until the Mohawk lobe receded far enough east to open drainage to Lake Schoharie. The Hinckley-St. Johnsville (SH on Figure 2; Oneida Lobe) and Canajoharie (Mohawk Lobe) ice margins shown on Figure 4 depict the deglacial paleogeography of northern New York at the culmination of the Valley Heads Readvance. Proglacial lakes Miller and Schoharie (Figure 4) were impounded along the margins of the Oneida and Mohawk lobes. Outflow from Lake Schoharie was directed southeastward through Catskill Creek to proglacial Lake Albany, which in turn drained southward through the Hell's Gate threshold in New York City (Stanford 2009).

Nunataks began to emerge as the ice thinned over the Adirondack High Peaks and the south-central highland regions. The suture between glacial ice sourced in the Ontario lobe and that sourced in the Hudson-Champlain lobe probably occurs in the highland formed by Snowy Mountain, Blue Mountain, and the High Peaks (Figure 4). The ice margin retreated briefly from its Valley Heads maximum before a short readvance to the Barneveld-Little Falls ice margin in the western Mohawk valley (BL on Figure 2; Ridge and Franzi 1992). This readvance was fronted by a lower level of Lake Schoharie (Delanson outlet; Lake Gravesville of Ridge and Franzi [1992]) in the Mohawk Valley. Lower Lake Schoharie drained to the Lake Amsterdam level, and eventually Lake Albany inundated the lower Mohawk Valley as ice recession uncovered the eastern end of the Mohawk Valley.

Ninemile Ice Margin

The Ninemile ice margin depicts the deglacial paleogeography at the terminus of the Ninemile Readvance (Ridge and Franzi 1992) near Rome in the western Mohawk Valley (Figure 5). Deposits and landforms associated with the Stanwix Readvance (Fullerton 1971, 1980) may have been formed by the more extensive Ninemile Readvance (Ridge and

Franzi 1992). Regional ice margin correlations suggested that the Ninemile Readvance may have been associated with the Luzerne Readvance in the upper Hudson Valley (Figure 2; Ridge 2003, 2004, 2016). Our placement of the Hudson–Champlain ice margin north of the Luzerne Readvance position, however, is consistent with the asynchroneity of glacial readvances observed in the Mohawk Valley (Ridge 1985; Ridge et al. 1990; Ridge and Franzi 1992). Furthermore, DeSimone et al. (2008) and DeSimone and LaFleur (2008) questioned the physical evidence for the Luzerne Readvance and suggested that the ice marginal deposits and landforms described by (Connally and Sirkin 1971, 1973) represent a short-lived recessional moraine. The upper Hudson Valley ice margin depicted in Figure 5 lies approximately 10 km north of Fort Ann, near the point at which proglacial Lake Albany (ABII) drained to the Quaker Springs level (DeSimone et al. 2008).

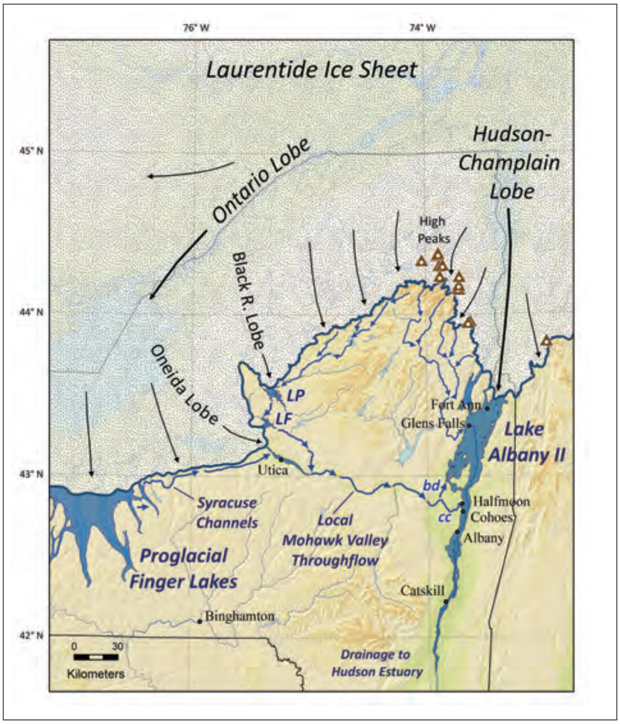


Figure 5: *Paleogeography of glacial ice lobes and proglacial lakes in northeastern New York at the maximum extent of the Ninemile Readvance in the western Mohawk Valley. The abbreviations refer to proglacial lake Frenchville (LF) in the upper Mohawk Valley, proglacial Lake Port Leyden (Fairchild 1912) in the Black River Valley, and the Crescent (cc) and Ballston-Drummond (bd) distributary channels of the Iro-Mohawk River near Cohoes. The brown triangles indicate the possible locations of nunataks as determined by glacier-profile modeling.*

Ridge and Franzi (1992) suggested that Lake Amsterdam occupied the western Mohawk Valley at the culmination of the Ninemile Readvance. Mohawk Valley lakes may also have been responsible for an early high phase of Lake Iroquois (Lake proto- or hyper-Iroquos; Domack, Leventer, Kopp, Lucas, Patacca and Scholz 2016; Domack, Scholz, Owen, Lothrop and Winsor 2016; Fairchild 1909; Fullerton 1971, 1980) in the eastern Oneida Lowland following recession of the Oneida Lobe from its Ninemile terminus. Base level controls for Mohawk Valley lakes at this time are problematic because the ice margin in the upper Hudson Valley was located well north of the confluence of the Mohawk and Hudson rivers, and the Hudson or Mohawk ice lobes could not have served as ice dams. The most likely controls for Lake proto- or hyper-Iroquois are bedrock constrictions in the Mohawk Valley at Moss Island in Little Falls or "The Noses" at Canajoharie. However, projection of the proto- or hyper-Iroquois lake terraces identified by Domack et al. (2016a, 2016b) fall close to the elevation of the Little Falls threshold but well below that of the Canajoharie threshold. These data are consistent with the Little Falls threshold as the outlet for Lake proto- or hyper-Iroquois but requires incision of the Canajoharie threshold prior to the creation of the lake.

Late lake phases in the Mohawk Valley and Oneida Lowland probably predate the eastward drainage of proglacial lake outflow from the Great Lakes basins through the Mohawk Valley and may correlate with a brief period of westward ice marginal drainage in the Erie and western Ontario basins. The additional outflow from the proglacial Great Lakes established the Iro-Mohawk River in the Mohawk Valley (Figures (Figures 3 and 6), which may have incised the bedrock threshold at Little Falls and caused the drop of lake level from Lake proto- or hyper-Iroquois to Lake Iroquois (main phase) in the Oneida and Ontario lowlands. This outflow may also have facilitated the drop in lake level from Lake Albany II to Lake Quaker Springs in the Hudson Lowland. Wall (1995) estimated the maximum discharge of the Iro-Mohawk River exceeded 42,500 m^3/s, which is more than ten times the maximum discharge of the modern Mohawk River over the past 90 years of record (DeSimone et al. 2008; Wall 2010). Ongoing work in the eastern Oneida Lowland (Domack et al. 2016a, 2016b) may shed light on the later stages of ice recession in the Mohawk Valley and Oneida Lowland.

Lake Albany drained through a succession of short-lived lower lake phases (lakes Albany II, Quaker Springs, and Coveville [DeSimone et al. 2008]) as the Hudson–Champlain Lobe receded into the upper Hudson Valley. Proglacial lake level fell to the Lake Coveville level (Coveville Stage of Lake Vermont) when the ice margin receded to a position marked by the Street Road ice-contact delta north of Ticonderoga (DeSimone et al. 2008; Stanford 2009). The outlet location for Lake Coveville has not yet been established, but recent work in the upper Hudson Valley suggests that a sediment dam near the village of Halfmoon is a likely candidate (Figure 6; DeSimone et al. 2008). Eastward fluvial drainage from the Mohawk Valley initially entered proglacial Lake Albany II in two distributary channels; the Ballston and Crescent channels (Figure 5). The Ballston channel expanded northward and split into two separate channels, the Saratoga Lake and Round Lake channels, when proglacial Lake Albany II fell to Lake Quaker Springs and then to Lake Coveville level.

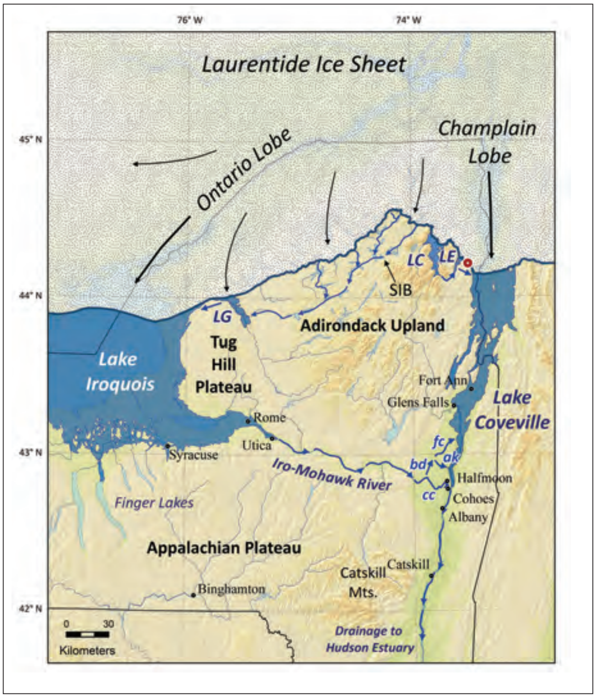


Figure 6: *Paleogeography of glacial ice lobes and proglacial lakes in northeastern New York at the Carthage–Loon Lake–Elizabethtown ice margin. The abbreviation SIB refers to the Saranac Intramontane Basin (Buddington 1953). PProglacial lakes Chapel (LC) and Elizabethtown (LE) formed in front of the ice margin in the Keene and Boquet valleys, respectively (Franzi et al. 2016). Proglacial Lake Glenfield (LG) was impounded in the Black River Valley (Fairchild 1912). Ballston-Drummond (dc), Fish Creek (fc), and Anthony Kill (ak) distributaries of the Iro-Mohawk River drain into Lake Coveville in the Hudson and Champlain lowlands. The red circle indicates the location of the Elizabethtown musk-ox fossil in proglacial Lake Hoisington (Franzi et al. 2016).*

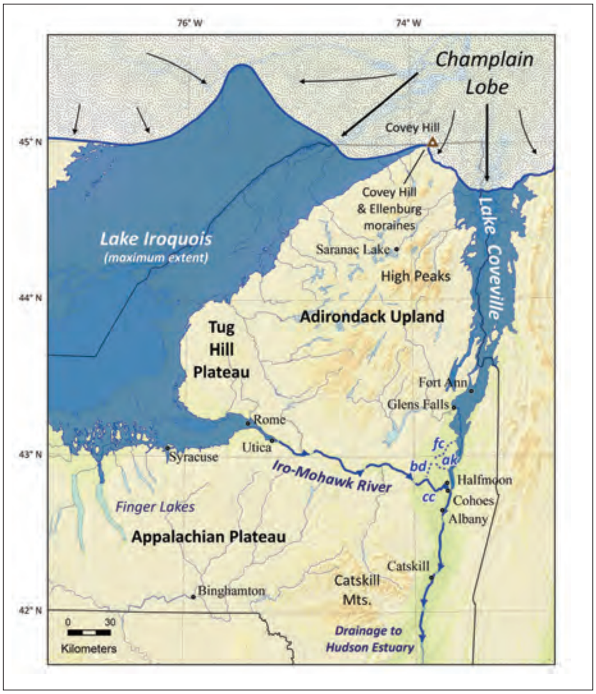


Figure 7: *Paleogeography of glacial ice lobes and proglacial lakes in northeastern New York at the maximum extents of proglacial lakes Iroquois in the Ontario and St. Lawrence lowlands and Coveville in the Champlain Lowland. The Crescent (cc) distributary channel of the Iro-Mohawk River remains active but the Ballston-Drummond (dc), Fish Creek (fc), and Anthony Kill (ak) are abandoned at this time.*

The three-channel distributary system (Stoller 1918) reached its full development in Lake Coveville (DeSimone et al. 2008). Flow in the Crescent channel occurred under steeper hydraulic gradients and progressively captured a greater proportion of fluvial throughflow from the Mohawk Valley. Eventually, all of the Iro-Mohawk River discharge was diverted to the Crescent channel (Figure 6). Fluvial drainage was established in the lower Hudson Valley south of the Halfmoon threshold and flowed southward to the Hudson estuary near New York City (Stanford 2009). The potholes at Cohoes were either exhumed or created after the onset of fluvial conditions in the lower Mohawk Valley. Some of the potholes described by James Hall (1871) were remarkably narrow (~1 m) and deep (>7 m) and may have formed by cavitation under extraordinarily high-magnitude flow conditions (Wall 2010). Cohoes Falls retreated rapidly from the vicinity of the present Hudson-Mohawk confluence and Iro-Mohawk River flow eroded the gorge below the falls. One of the potholes would later become the resting place for the Cohoes mastodon, but that fossil could not have been preserved as long as the Iro-Mohawk River remained active. The mastodon fossil yielded an age of 12.82-13.10 cal. kyr. B.P. (pooled mean NYSM VP101; Feranec and Kozlowski 2016), which post-dates the Iroquois breakout at Covey Hill, the drainage of Lake Iroquois to the Lake Frontenac level, and the abandonment of the Iro-Mohawk River (Figures 6, 7, and 8).

Proglacial lakes formed in the north-draining Finger Lakes valleys as ice receded from the Valley Heads Moraine (Figure 5; Cadwell and Muller 2004; Mullins and Hinchey 1989; Mullins et al. 1996) in central New York. Initially, the lakes drained south across the moraine, but as ice recession continued, the lakes expanded, lower outlets were uncovered, and later outflow, supplemented by outflow from proglacial lakes in the Great Lakes region, was diverted eastward along the ice front. The eastward ice-marginal drainage carved an extensive system of ice marginal channels across interfluves along the northern flank of the Appalachian Plateau. The channel system is best preserved between Syracuse and Oneida where there is evidence for at least two episodes of channel cutting by ice marginal, supraglacial, or subglacial drainage (Sissons 1960).

Most of the south-central Adirondack highlands and the southern High Peaks regions were probably ice-free and several High Peaks nunataks remained at this time (Figure 5). The ice margin generally followed the southwest-northeast trend of the low mountains that form the southern flank of the Saranac Intramontane Basin west of the High Peaks region. Meltwater was directed southwestward to proglacial lakes in the Black River Valley and ultimately to the western Mohawk Valley. Meltwater east of the High Peaks region flowed southward via the Hudson and Schroon rivers to the upper Hudson Lowland near Glens Falls (Figure 5).

Carthage–Loon Lake–Elizabethtown Ice Margin

Lake Coveville expanded northward as the ice margin receded into the Champlain Lowland and small proglacial lakes formed in north-draining tributaries in the northeastern Adirondack Mountains. The Carthage–Loon Lake–Elizabethtown ice margin (CLLE; Figure 6) is marked by ice-contact deltaic deposits and ice-marginal channels in the AuSable, Boquet, and Black valleys, but it is not associated with any known readvances. The ice margin is significant, because its age is constrained by a radiocarbon age on a musk-ox vertebrae of 11.28 ± 0.11 14C kyr. BP (AA-4935; Cadwell and Pair 1991), which corresponds to a calibrated age between 13.03–13.45 cal. kyr. B.P. when corrected for estimated $\delta^{13}C$ (Rayburn et al. 2007, 2011). The musk-ox bone was discovered by a gravel pit operator in prodeltaic facies in a small ice-contact delta near Elizabethtown.

The CLLE ice margin correlates with the Loon Lake stand of Denny (Denny 1974) in the upper Saranac River basin and the Elizabethtown ice margin (Franzi et al. 2015; Rayburn et al. 2007) in the AuSable, Boquet and Champlain valleys (Figure 6). Proglacial lake outflow in the AuSable, Boquet, and Black valleys was eastward to Lake Coveville in the Champlain Valley. The ice margin generally follows the northern flank of the SIB westward to the northern end of the Black River Valley. Large masses of stagnant ice were left in the SIB and thick deposits of ice-contact stratified drift and outwash were deposited by meltwater streams that drained westward from the upper Saranac River basin to the upper St. Regis River basin (Denny 1974). The meltwater outflow eventually entered proglacial lakes in the lower Black River Valley and ultimately into proglacial Lake Iroquois, which occupied the Ontario Lowland (Figure 6). Lake Iroquois received inflow from proglacial lakes in the eastern Great Lakes basins (Muller and Prest 1985) and discharged eastward across the Little Falls threshold via the Iro-Mohawk River. It is probable that all of the Iro-Mohawk River discharge flowed through the southernmost, or Crescent, distributary channel at Cohoes (Figure 6).

Ellenburg–Plattsburgh Ice Margin

Lakes Iroquois and Coveville expanded northward with further ice recession in the St. Lawrence and Champlain lowlands, respectively. The Ellenburg–Plattsburgh ice margin (Franzi et al. 2015) corresponds to Denny's ice front position 8 (Denny 1974) and depicts proglacial Lake Iroquois in the Ontario and St. Lawrence valleys and Lake Coveville in the Champlain Valley near their maximum extents (Figure 7). The ice margin follows the trend of a small east-west trending recessional moraine north of Clinton Mills and the Ellenburg Moraine, a large generally north-south trending moraine that crosses the upper North Branch of the Chazy River Valley near Ellenburg Depot (Denny 1974; Franzi et al. 2007). A small proglacial lake impounded in front of the Ellenburg Moraine in the upper North Branch valley probably drained southeastward along the ice front (Franzi et al. 2007, 2015). The outflow may be responsible for ice marginal channels and ice-contact stratified deposits just north of Plattsburgh (Franzi et al. 2015). These deposits and landforms immediately predate the breakout of Lake Iroquois into the Champlain Lowland.

Lake Iroquois Breakout and Lakes Frontenac and Fort Ann

Ice recession from the Ellenburg-Plattsburgh ice margin uncovered the Covey Hill threshold, and Lake Iroquois drained catastrophically across the St. Lawrence–Champlain divide into Lake Coveville (Denny 1974; Franzi et al. 2007; Franzi, Rayburn, Yansa, and Knuepfer 2002; Rayburn et al. 2005). The breakout released 570 ± 85 km^3 of water storage from the Ontario and St. Lawrence lowlands at an estimated discharge of 83,000-92,000 m^3/s during the waning stages of the flood (Rayburn 2004; Rayburn et al. 2005). Proglacial lake level in the Ontario and St. Lawrence lowlands dropped by about 14 m, and the Lake Iroquois outlet near Rome was abandoned (Figures 7 and 8). The Iroquois breakout ended eastward Iro-Mohawk River drainage in the Mohawk Lowland and represents the earliest opportunity for the preservation of the Cohoes mastodon fossil in the pothole near Cohoes Falls (DeSimone et al. 2008; Wall 2010). Falling water level in the Ontario and St. Lawrence lowlands stabilized at the Covey Hill threshold and Lake Frontenac became established with outflow through The Gulf at Covey Hill (Muller and Prest 1985; Pair and Rodrigues 1993). Sustained outflow of approximately 56,000 m^3/s (Rayburn et al. 2005) from Lake Frontenac carved The Gulf, a narrow, 1.5 km-long gorge cut deeply into the Potsdam Sandstone south of Covey Hill (Denny 1974; Muller and Prest 1985; Pair and Rodrigues 1993).

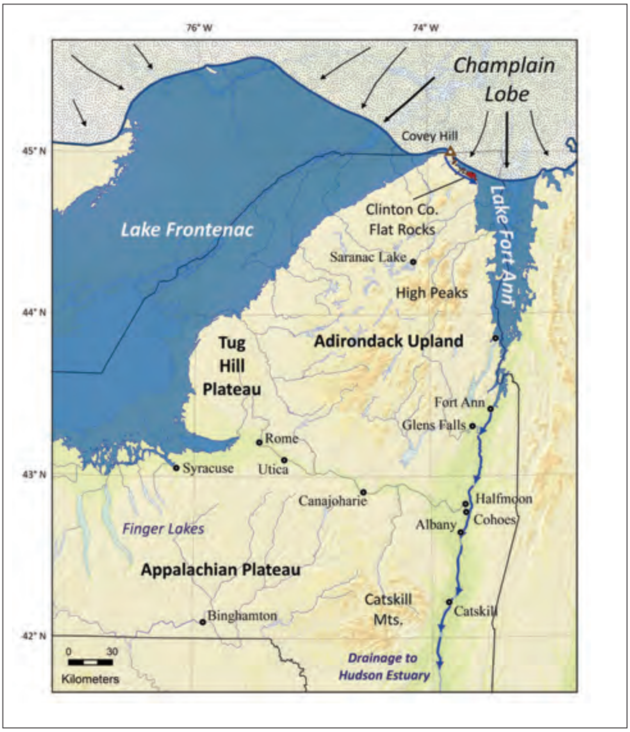


Figure 8: *Paleogeography of glacial ice lobes and proglacial lakes in northeastern New York showing proglacial Lake Frontenac in the St. Lawrence Lowland and proglacial Lake Coveville in the Champlain Lowland. The spatial extents of the Clinton County Flat Rocks was derived from Denny (1974).*

The flood waters were directed southeastward along the ice front in the Champlain Lowland, stripping the surficial cover from large areas and carving channels and plunge pools in the Potsdam Sandstone. The intense scour created a discontinuous 30 km long belt of exposed sandstone surfaces that are known locally as the Clinton County Flat Rocks (Figure 8; Denny 1974; Franzi et al. 2007; Rayburn et al. 2005; Woodworth 1905a, 1905b). The physical environment of the Flat Rocks today is characterized by extreme deficiencies in nutrients and soil moisture, and some locations host globally rare sandstone-pavement jack pine barrens (Franzi and Adams 1993). The ice-marginal flood waters deposited the coarse ice-contact boulder deposits that comprise Cobblestone Hill where they entered Lake Coveville at the southeastern margin of Altona Flat Rock (Woodworth 1905b; Denny 1974; Franzi et al. 2002; Rayburn et al. 2005). Franzi et al. (2002) traced Coveville shorelines northward to Cobblestone Hill and recognized that the deposits occur in two terraces that correspond roughly to Coveville and Fort Ann proglacial lake levels in the Champlain Lowland. They proposed that transmission of the Iroquois breakout floodwave through the Champlain and upper Hudson lowlands caused the failure of the Coveville threshold and caused Lake Coveville to drain to the Fort Ann level (Figure 8; Franzi et al. 2002; Rayburn et al. 2005). Falling water levels in the Champlain Lowland encountered higher, temporary outlet channels near the Fort Ann threshold, creating a succession of ephemeral Lake Fort Ann strandline features in the lowland (Rayburn 2004; Rayburn et al. 2005). The highest of these features define the "Upper Fort Ann" water surface but Lake Fort Ann only became stable at the "Lower Fort Ann" level as continued incision uncovered the bedrock threshold near Fort Ann (Chapman 1937; Rayburn 2004; Rayburn et al. 2005). The drainage of Lake Coveville to the upper Lake Fort Ann level released an additional 130 ± 20 km^3 from water storage in the Champlain Lowland for a total storage loss of approximately 700 km^3 for the combined Iroquois–Coveville breakouts. The lake outflow was directed southward into the upper Hudson Lowland, where erosion exhumed the pre-glacial Fort Ann Branch and Battenkill-Hudson channels used by the Hudson River today.

Drainage of Lake Frontenac

Lake Frontenac drained as the ice front receded from the north flank of Covey Hill, and lake levels in the Ontario–St. Lawrence and Champlain lowlands merged at the Fort Ann level (Figure 9). The Iroquois and Frontenac drainage events are recorded by erosional unconformities and coarse flood deposits in cores and outcrops in the northern Champlain Lowland. Franzi et al. (2007) estimated the duration of Lake Frontenac to be about 50 years based upon estimates from compositional changes in varved clay deposits at Whallonsburg in the Champlain Lowland. Unlike the earlier Lake Iroquois breakout, the drainage of Lake Frontenac left little physical evidence of a catastrophic discharge event. That evidence may be buried by younger deposits or the drainage may have occurred as a broad flow over, under, or through the ice front (Rayburn et al. 2005; Franzi et al. 2007). The drainage of Lake Frontenac released about 2,500 ± 375 km^3 (Rayburn 2004; Rayburn et al. 2005) of storage in the Ontario and St. Lawrence lowlands that was directed southward into the upper Hudson lowland. Lake Fort Ann existed for about 170 years, as evidenced by post-Lake Coveville varved clays from sediment cores collected from the Champlain Lowland (Rayburn et al. 2011).

Figure 9: *Paleogeography of glacial ice lobes and proglacial lakes in northeastern New York following the drainage of Lake Frontenac and the expansion of Lake Fort Ann into the St. Lawrence Lowland. The spatial extents of the Clinton County Flat Rocks was derived from Denny (1974). The brown triangle marks the location of Covey Hill.*

Champlain Sea

Ice-margin recession in the lower St. Lawrence Lowland near Warrick, Quebec opened a connection to the Atlantic Ocean causing Lake Fort Ann to drain, releasing approximately 1,500 km^3 of proglacial lake water to the Gulf of St. Lawrence (Rayburn et al. 2005), and marine water inundated the isostatically depressed St. Lawrence and Champlain lowlands (Figure 10). The marine episode, known as the Champlain Sea, is marked by abundant marine microfauna, invertebrates, fish, and mammals in stratigraphic sections in the Champlain and St. Lawrence lowlands (Cronin 1988; Cronin et al. 2012; Feranec, Franzi, and Kozlowski 2014; Harrington 1988; Hunt and Rathburn 1988; Steadman, Kirchgasser, and Pelky, 1991). Notable marine mammal fossils from Champlain Sea deposits in New York include the Norfolk whale (Steadman et al. 1991) from the St. Lawrence Lowland near Potsdam and two seal fossils from Plattsburgh (Feranec et al. 2014) in the Champlain Lowland. The marine incursion began between 13.12 and 12.85 cal. kyr. B.P. (Cronin et al. 2008, 2012), but the transition was not straightforward. Micropaleontological and isotopic evidence from cores of proglacial lake and marine deposits in the St. Lawrence and Champlain Valleys (Rayburn et al. 2007, 2011; Cronin et al. 2008, 2012) indicate that freshwater conditions returned shortly after the initial marine incursion before reverting to the main marine phase of the Champlain Sea. Cronin et al. (2012) attributed the freshening event to influx of large quantities of freshwater from proglacial Lake Agassiz in west-central Canada. Rayburn et al. (2011) estimated the duration of the freshening interval to be about 120 years. The freshwater interval that was followed by a mixed "transitional phase," containing both freshwater ostracodes and marine forams following the freshening event, likely marked punctuated shorter pulses of high-discharge freshwater influx during the return to full marine conditions. Marine conditions persisted until differential isostatic rebound raised the northern portion of the basin above sea level and Lake Champlain became established in the Champlain Lowland. The Champlain Sea-Lake Champlain transition began around 9.4 cal. kyr. B.P. and was completed by about 8.6 cal. kyr. B.P. (Belrose 2015).

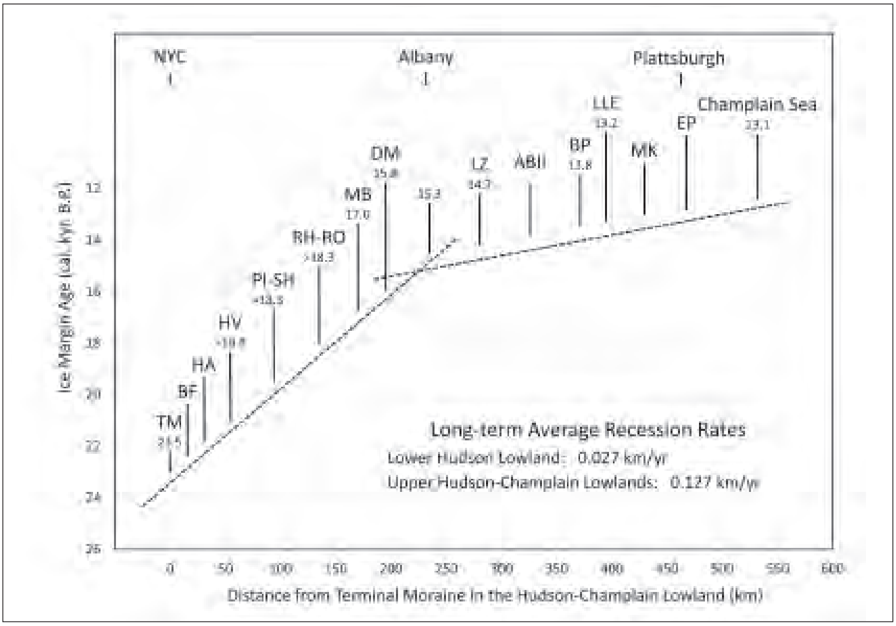
Figure 10: *Paleogeography of glacial ice lobes and proglacial lakes in northeastern New York at the maximum extent of the Champlain Sea.*

CONCLUDING REMARKS

Rate of Ice Recession

Ice recession from the terminal moraine in New York began approximately 23 cal. kyr. B.P. in the lower Hudson Valley (Figures 2 and 11), and by 16-17 cal. kyr. B.P., the ice front reached the Albany region and Mohawk Valley. The average retreat rate in the lower Hudson Valley was approximately 0.03 km/yr, but this long-term average rate includes several possible short-term readvances. Long-term average post-Valley Heads retreat rates are nearly four times greater (about 0.14 km/yr; Figure 11) in the upper Hudson and Champlain lowlands. Franzi et al. (2007) reported short-term recession rates in the range of 0.19 to 0.44 km/yr in the Champlain Lowland. Recession rates of this magnitude are typical of calving ice margins that are associated with unstable buoyant ice (Benn, Warren, and Mottram 2007; Joughin, Smith, Shean, and Floricioiu 2014). We consider it likely that rapid post-Valley Heads recession rates in the Ontario-St. Lawrence and Champlain lowlands were a function of late-glacial climatic amelioration, changing patterns of ice flow and calving dynamics. The rate of ice retreat in the deep lake basins adjacent to the Adirondack Upland contrast sharply with that in the lower Hudson and Mohawk valleys, where ice recession was interrupted by short-term readvances.

Figure 11: *Comparison of ice recession rates in the lower Hudson and upper Hudson-Champlain lowlands. Abbreviations are keyed to Figure 2. The ice margin position at the time of the Champlain Sea incursion is approximated by the ice front position depicted in Figure 9, which most likely produces and under estimate of recession rate. The location of the Lake Albany II (ABII) ice margin is shown in Figure 5.*



Paleoclimate Implications

The weakening of Atlantic Meridional Overturning Circulation (AMOC) caused by decreasing North Atlantic salinity is often invoked to explain global climate change. Broecker et al. (1989) suggested that the diversion of Lake Agassiz outflow from the Gulf of Mexico to the North Atlantic via the Great Lakes and St. Lawrence Lowland may have triggered the late Pleistocene Younger Dyras climate reversal. This hypothesis has been the subject of considerable debate (cf. Cronin et al. 2012). Rayburn et al. (2005, 2007) noted the correspondence in age between steady-state outflow and breakout flood events from large proglacial lakes in the Ontario, St. Lawrence, and Champlain lowlands and the record of late glacial climate changes from the Greenland GISP2 (Taylor et al. 1993) ice core and deep-sea sediment cores (Hughen, Southon, Lehman, and Overpeck 2000; Figure 12). Furthermore, immediately following the initial establishment of the Champlain Sea, a strong freshwater discharge flowing through the St. Lawrence Lowland caused a return to freshwater conditions in the Champlain Sea basin for at least a century. The freshwater phase was followed by shorter pulses of high freshwater influx during the transition back to full-marine conditions (Cronin et al. 2012). Rayburn et al. (2011) dated the transition between 12.74-13.07 cal. kyr. B.P., which is supported by a varve-count estimate that the transition occurred at least 330 years following the establishment of Covey Hill ice margin. The radiocarbon age/varve correlation agrees well with the Elizabethtown ice margin age and places the cessation of large meltwater discharges at the inception of the Younger Dryas (Figure 12). The timing, magnitude and duration of late glacial freshwater discharges from the St. Lawrence and Champlain Valley region may have been sufficient to affect AMOC and trigger the Younger Dryas interval (Figure 12).

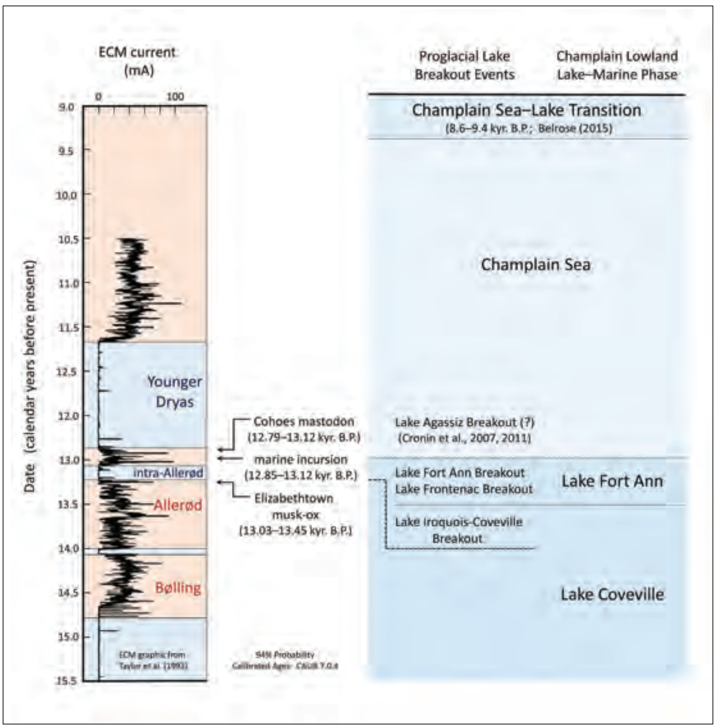


Figure 12: *Timing of proglacial lake phases and breakout events in the Champlain Lowland and the late Pleistocene-early Holocene climate variability, as recorded by the yearly average electrical conductivity in the GISP2 ice core at Summit, Greenland (Taylor et al. 1993).*

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