LATE QUATERNARY HISTORY OF NORTHEASTERN NEW YORK AND ADJACENT PARTS OF VERMONT AND QUEBEC

70TH ANNUAL REUNION

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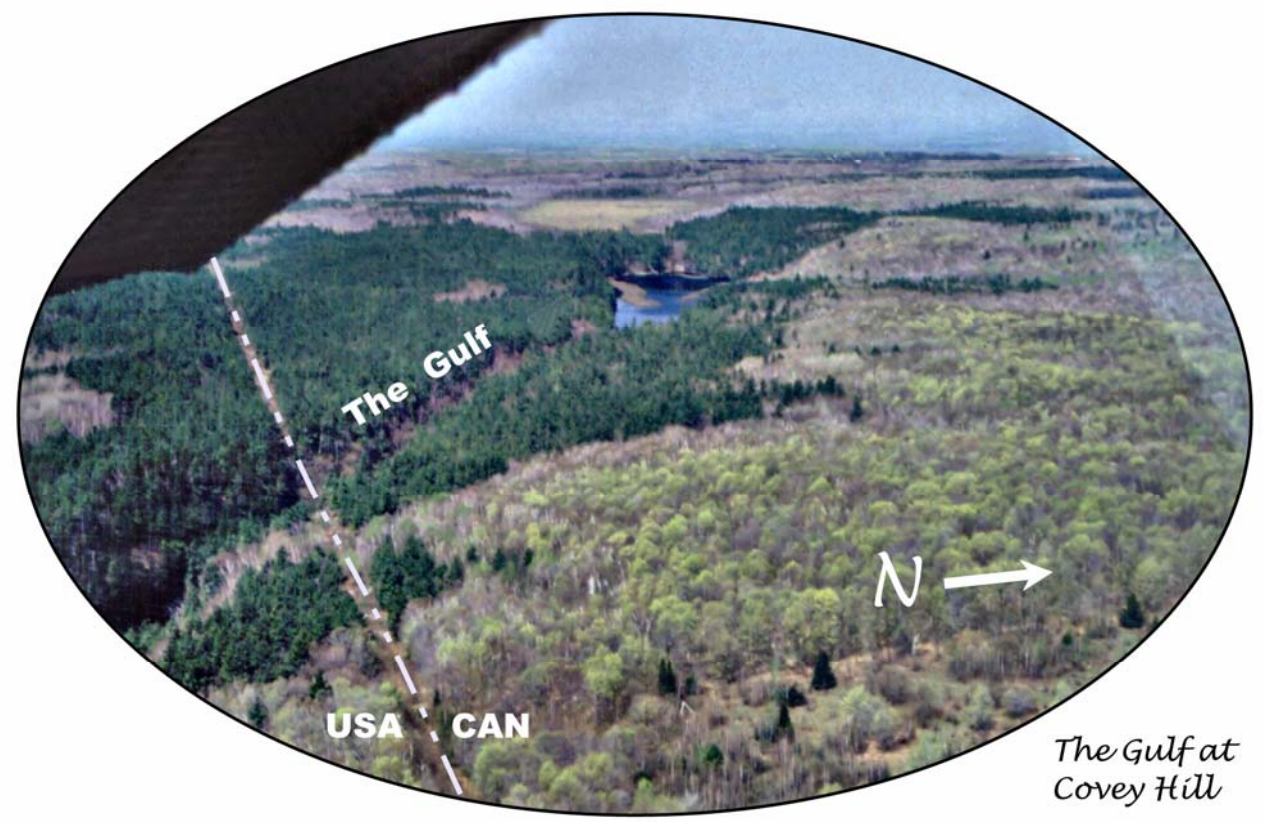

The Gulf at Covey Hill

Hosted by

State University of New York at Plattsburgh

1-3 June, 2007

David A. Franzi, John A. Rayburn, Peter L.K. Knuepfer and Thomas M. Cronin

WE DEDICATE THIS FIELD TRIP TO

CHARLES S. DENNY

Photo courtesy of Wayne Newell

FOR HIS MANY CONTRIBUTIONS TO OUR UNDERSTANDING OF THE GLACIAL HISTORY OF THE CHAMPLAIN AND ST. LAWRENCE LOWLANDS

LATE QUATERNARY HISTORY OF NORTHEASTERN NEW YORK AND ADJACENT PARTS OF VERMONT AND QUEBEC

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70th Reunion
Northeastern Friends of the Pleistocene

SUNY Plattsburgh
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LATE QUATERNARY HISTORY OF NORTHEASTERN NEW YORK AND ADJACENT PARTS OF VERMONT AND QUEBEC

Introduction

The geographic location and physiography of northern New York (Figs. 1 and 2) played an important role in late Wisconsin glacial events in northeastern North America. Ice flow at the last glacial maximum (ca. 28,000–24,000 calendar years B.P.; Ridge, 2003) was concentrated in terrestrial ice streams in major lowlands (e.g. Dreimanis and Goldthwait, 1973; Hughes et al., 1985; Occhietti et al., 2001). Ice in the St. Lawrence–Ontario–Erie lowlands flowed southwestward along the front of the Appalachian Plateau in western New York into central Ohio. Ice in the Hudson–Champlain lowlands flowed southward to the Atlantic Coastal Plain where it merged with ice from New England to form the terminal moraine on Long Island. The lowlands also served as proglacial lake basins and drainage routes for meltwater during deglaciation. Outflows from large proglacial lakes in the Great Lakes and Agassiz basins were directed at different times through the Mohawk–Hudson, Champlain–Hudson, and St. Lawrence lowlands. The first two drainage routes sent outflow to the mid-Atlantic region while the latter directed outflow to the Gulf of St. Lawrence in maritime Canada. Broecker et al. (1989) suggested that changes in routing of proglacial lake outflow such as these may have affected North Atlantic thermohaline circulation, and consequently, late glacial climate. The glacial-stratigraphic record in New York, therefore, may be the "Rosetta Stone" that links glacial chronologies in New England and the Midwest (e.g. Karrow et al., 2000; Muller and Calkin, 1993; Ridge, 2003) and provides important new insights on late Wisconsin climate.

This trip examines morphologic and stratigraphic evidence for ice-sheet retreat from the Adirondack Upland, the creation and evolution of deep proglacial lakes and the subsequent marine incursion of the Champlain Sea in the St. Lawrence and Champlain lowlands. The general chronology of late glacial events, especially those associated with proglacial water bodies in the region, has been recognized since early studies by Emmons (1842), Woodworth (1905a, 1905b), Fairchild (1912, 1919) and Taylor (1924). Chapman (1937) conducted the first extensive study of freshwater and marine water bodies in the Champlain Lowland. He recognized two stages of a proglacial lake which he called Lake Vermont–the Coveville (upper) and Fort Ann (lower). The relationship between Lake Vermont and coeval lakes in the upper Hudson Lowland remains unresolved despite considerable study (e.g. Connally and Sirkin, 1969, 1971, 1973; Wagner, 1972; Parrott and Stone, 1972; DeSimone and LaFleur, 1985, 1986; Stanford and Harper, 1991; Wall and LaFleur, 1995; Connally and Cadwell, 2002). Rayburn et al. (2005) review the literature on this subject and suggest that the Coveville Stage of Lake Vermont was contemporaneous with the latest stage of Lake Albany and that both water levels were controlled by a threshold in the Hudson Lowland. Lake Albany may have been impounded behind the Terminal Moraine at the Hudson Narrows with its outlet across a threshold at Hell Gate (Stanford and Harper, 1991). Franzi et al. (2002) and

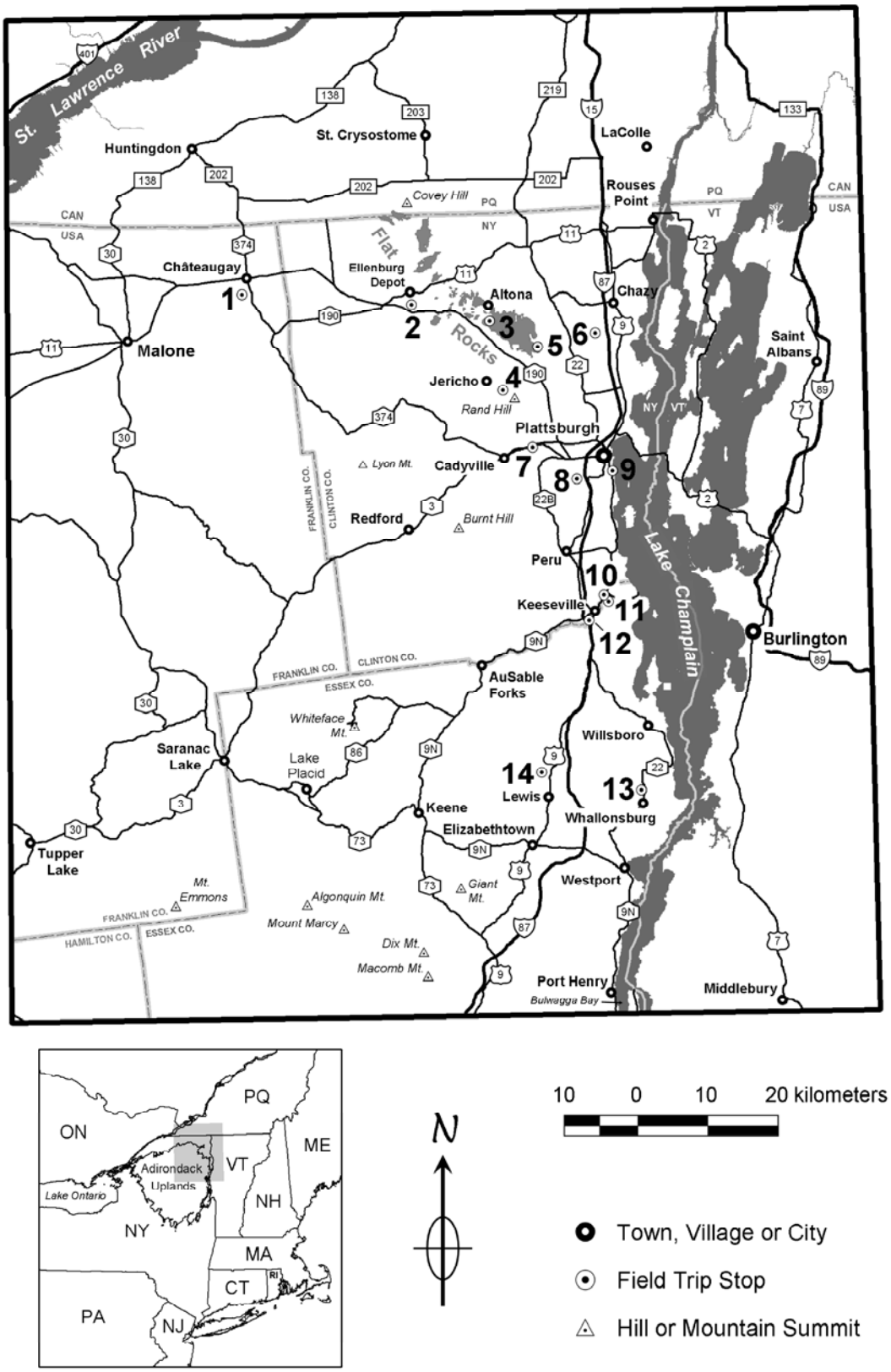

Figure 1. Map of the study area that shows the locations of field trip stops.

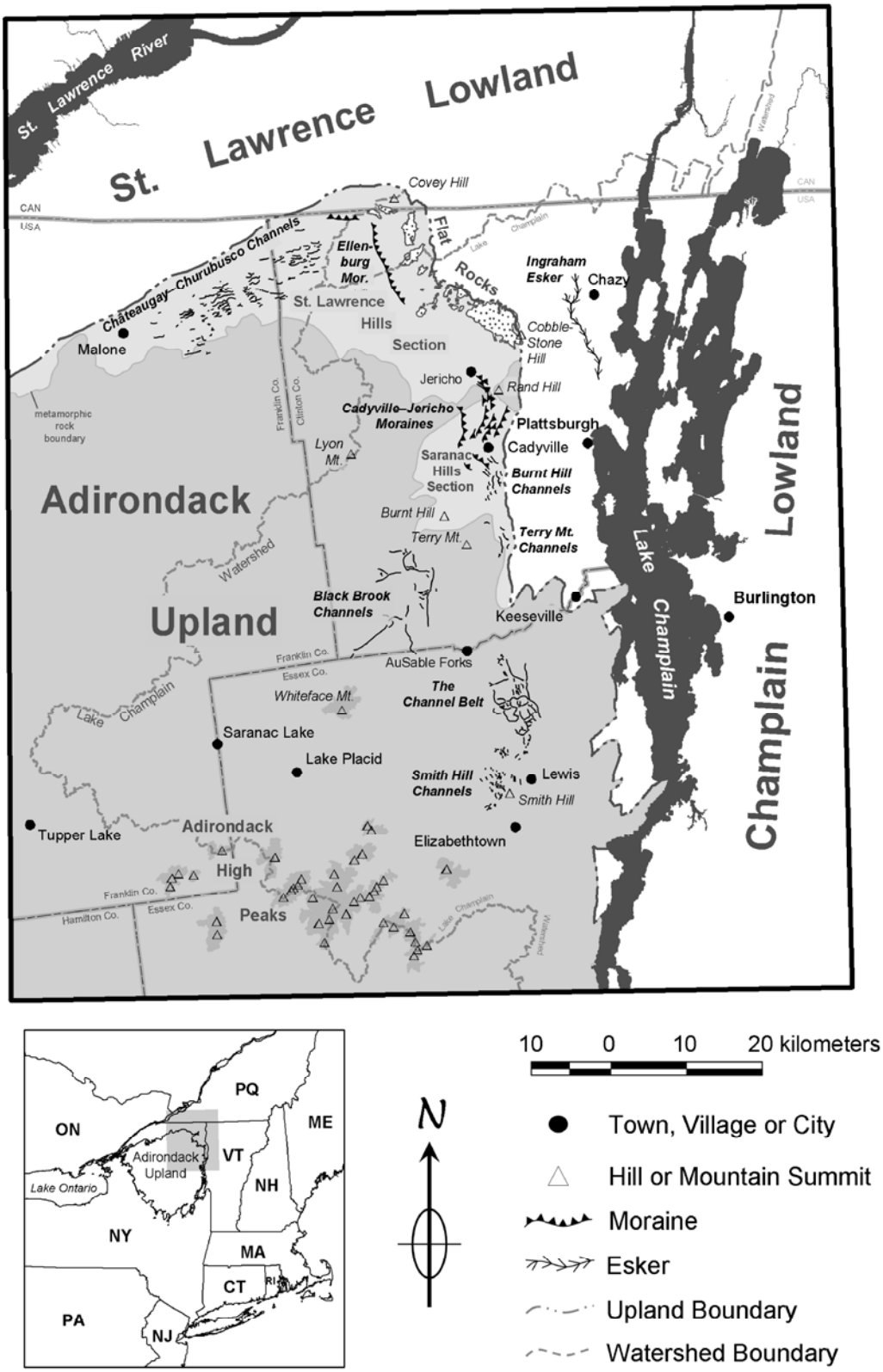

Figure 2. Physiographic map of the study area that shows the locations of physiographic subdivisions and landforms discussed in the text.

Rayburn et al. (2005) traced Coveville strandline deposits and landforms as far north as Cobblestone Hill at Altona Flat Rock and suggested that at this point during ice recession the lake dropped to the Fort Ann level. The outlet threshold for the Fort Ann Stage is located at a topographic constriction near Fort Ann, New York (Chapman, 1937; Rayburn et al., 2005). Outflow across the threshold incised the Fort Ann outlet over time and resulted in a multitude of strandline features that range in elevation between well-defined upper and lower limits in the Champlain Lowland (Franzi et al., 2002; Rayburn, 2004; Rayburn et al., 2005, 2007b).

Simultaneous ice recession in the Ontario–St. Lawrence lowlands allowed proglacial Lake Iroquois to expand along the northern flank of the Adirondack Upland (Coleman, 1937; MacClintock and Stewart, 1965; Denny, 1974; Clark and Karrow, 1984; Muller and Prest, 1985; Pair et al., 1988; Pair and Rodrigues, 1993). Lake Iroquois drained eastward through the Mohawk Lowland until lower outlets near Covey Hill, PQ were exhumed and outflow from later proglacial lakes in the St. Lawrence Lowland was rerouted across the St. Lawrence–Champlain drainage divide to Lake Vermont in the Champlain Lowland. Franzi et al. (2002), Rayburn (2004) and Rayburn et al. (2005, 2007b) suggested that the discharge breached the Coveville Stage dam in the Hudson Lowland and subsequently caused the drainage of Lake Albany and the lowering of Lake Vermont to the Fort Ann Stage in the Champlain Lowland. Lake Fort Ann eventually became confluent with the proglacial lakes in the Ontario–St. Lawrence lowlands after ice retreated from the northern slope of Covey Hill (MacClintock and Stewart, 1965; Clark and Karrow, 1984; Muller and Prest, 1985; Pair et al., 1988; Pair and Rodrigues, 1993). The confluence did not involve a change in the outlet location; thus we retain the name Fort Ann for all post-Coveville proglacial lake stages in the St. Lawrence and Champlain lowlands. The St. Lawrence Lowland portion of Lake Fort Ann has been variously referred to as Lake Belleville, Lake Trenton, Lake St. Lawrence and Lake Candona (e.g. Clark and Karrow, 1984; Pair et al., 1988; Pair and Rodrigues, 1993; Parent and Occhietti, 1988, 1999; Occhietti et al., 2001). The reader is referred to Pair and Rodrigues (1993) for a discussion of these names. Northward ice recession from the lower St. Lawrence Lowland near Quebec allowed the proglacial lakes to drain and marine water invaded the isostatically depressed St. Lawrence and Champlain lowlands.

Our recent work in the region focuses upon the extent, chronology and paleoclimatic implications of proglacial water bodies in the northern Champlain Lowland (Rayburn, 2004; Rayburn et al., 2005, 2007b). We shall present morphologic, stratigraphic and paleontologic evidence for glacial events in the region and new radiocarbon dates that refine the deglacial chronology. Some of the descriptions and discussions presented in this paper are derived from earlier field trips (Diemer and Franzi, 1988; Franzi and Cronin, 1988; Franzi and Adams, 1993, 1999; Franzi, et al., 1994, 2002; Landing et al., in press) and are updated where appropriate.

Physiography and Geological Setting

The St. Lawrence and Champlain lowlands form a broad, contiguous lowland region that is underlain by Cambrian and Ordovician sedimentary rocks (Fig. 2). The central

portions of the lowlands are underlain by relatively thick glacial, lacustrine and marine deposits and are characterized by low to moderate local relief (generally less than 100 meters). Local relief along the northern margin of the St. Lawrence Lowland and northwestern margin of the Champlain Lowland, in the St. Lawrence Hills subdivision of Cressey (1977), ranges up to a few hundred meters. The St. Lawrence Hills subdivision depicted in Fig. 2 is approximately bounded by the 200-meter contour line on the north and east and the outcrop boundary of Adirondack metamorphic rocks to the south. This area is underlain by lower Paleozoic sedimentary rocks, primarily the Cambrian Potsdam Sandstone, and includes the area around Covey Hill, PQ and the "Flat Rocks" in Clinton County, NY. We propose the name Saranac Hills for an area with similar relief and geology east of Plattsburgh.

The Adirondack Upland is a dome-shaped upland region primarily underlain by high-grade Proterozoic metamorphic rocks. Summit elevations throughout most of the northeastern Adirondack Upland range from 600 and 1000 meters but local relief is generally less than 700 meters. The highest summit elevations occur in the Adirondack High Peaks where many summits are greater than 1200 meters and local relief commonly exceeds 1000 meters.

Drainage patterns within the study area are influenced by regional geology. The principal streams in the region, including the Salmon, Chateaugay, Chazy, Saranac, AuSable and Boquet rivers, represent the northeastern portion of a radial drainage pattern developed in the Adirondack Upland. The St. Lawrence River and Lake Champlain are part of the tangential master stream network that developed in the lowlands surrounding the Adirondack Upland (Ruedemann, 1931; Morisawa, 1985) (Fig. 3 and 4).

Glacial Deposits and Landforms

Ice-Flow Indicators

The direction of late Wisconsin ice movement in the Champlain Lowland and northeastern Adirondack Upland region is inferred from striated bedrock exposures, till-pebble fabrics, roches moutonnées, drumlins, moraines, and compositional trends in tills (Ogilvie, 1902; Alling, 1916, 1918, 1919, 1920; Miller, 1926; Kemp and Alling, 1925; MacClintock and Stewart, 1965; Denny, 1974; Craft, 1976; Gurrieri and Musiker, 1990). Two predominant directions of flow are indicated in the published literature: a southerly flow that presumably relates to Late Wisconsin overriding of the Adirondack Upland by the Laurentide Ice Sheet and a late-glacial flow pattern that was strongly controlled by local physiography. In most instances striation orientation reflects the last ice movement in the region. Kemp and Alling (1925) and Craft (1976) suggested that local alpine glaciers might have contributed to late-glacial ice flow in parts of the Adirondack Upland.

Figure 3. Map of northern New York showing the radial and tangential drainage pattern that developed in and around the Adirondack Upland.

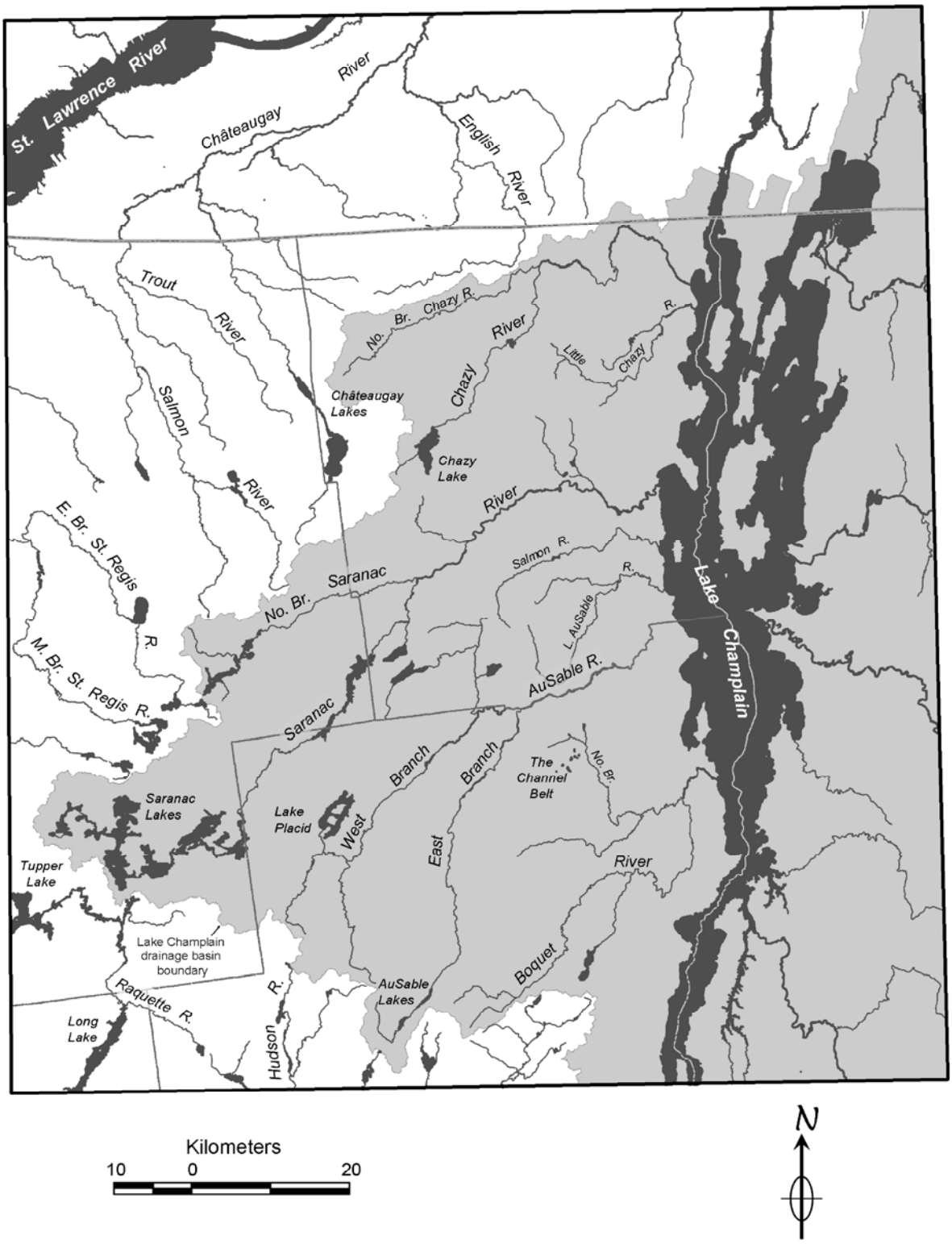

Figure 4. Principal drainage systems in northeastern New York.

Meltwater Channels

The morphology and continuity of meltwater channels and channel systems in the region reflect the magnitude and duration of meltwater discharge, the location of meltwater flow relative to the glacier margin, and the composition and structure of the substratum into which the channels are cut.

Small- to medium-size channels that are sub-parallel to the contours of hillslopes often occur as anastomosing channel systems that are cut into surficial deposits or, less commonly, bedrock. Small bedrock channels often originate at drainage divides and may have served as outlets for short-lived proglacial lakes. Individual channels generally range from 0.1–1.0 km long, 10–150 m wide, and 1–20 m deep, but the channel systems often occur in belts 0.5–2.0 km wide and several kilometers in lateral extent. Good examples of these channel systems occur near Chateaugay (MacClintock and Stewart, 1965; Denny, 1974; Pair and Rodrigues, 1993) (Field Trip Stop 1), along the northwestern margin of the Champlain Lowland between Jericho and Cadyville (Field Trip Stop 4), between Cadyville and Peru (Denny, 1974), and north and west of Smith Hill near Lewis (Field Trip Stop 14) (Franzi, 1992) (Fig. 2). The distal ends of many channels or channel systems open onto fluvial or deltaic sandplains that are graded to bedrock thresholds or proglacial impoundments.

Larger and more extensive channels and channel systems cut into bedrock that originate at cols on drainage divides were probably cut by meltwater outflow from proglacial lakes (Alling, 1916, 1918, 1919, 1920; Kemp and Alling, 1925; Miller, 1926; Denny, 1974; Diemer and Franzi, 1988). The outflow channels, presently abandoned or occupied by underfit streams, may attain depths greater than 30 meters and can often be traced more than 2 km. Bedrock thresholds at the outflow channel heads provided base-level control for glaciolacustrine and glaciofluvial sedimentation in the source basin. The elevations of outlet thresholds on drainage divides in Adirondack valleys generally decrease down valley, a distribution that is consistent with the systematic recession of active, valley-bound continental ice lobes (Diemer and Franzi, 1988).

The most extensive outflow channel system is found in the "Channel Belt" (Kemp and Alling, 1925) between Ausable Forks and Lewis (Fig. 2). Individual channels may contain deep, circular to ovate plunge basins that are presently occupied by small ponds or swampy depressions. The Channel Belt network heads at the South Gulf and The Gulf (not to be confused with The Gulf at Covey Hill) outflow channels, on the divide between the Ausable and Boquet drainage basins and was probably cut by the combined erosional effect of outflow and ice-marginal meltwater drainage. Outflow from these channels was responsible for deposition of the ice-contact stratified deposits at Oak Hill (Field Trip Stop 14). Other well-developed bedrock channel systems occur in Wilmington Notch (Diemer and Franzi, 1988), the upper Black Brook watershed between Redford and AuSable Forks (Miller, 1926; Denny, 1974), and south of The Gulf near Covey Hill (MacClintock and Stewart, 1965; Denny, 1974) (Fig. 2).

Rock Pavements

Rock pavements are large areas of exposed bedrock that are commonly associated with meltwater channels and channel systems. The largest rock pavements in the region

are the sandstone pavements known locally as "Flat Rocks" in Clinton County (Fig. 2). The Flat Rocks comprise a discontinuous, 5-kilometer-wide belt of sandstone pavements that extend approximately 30 km southeastward into the Champlain Valley from Covey Hill, PQ. The pavements were created by the erosional effects of catastrophic floods from the drainage of glacial Lake Iroquois and younger post-Iroquois proglacial lakes in the St. Lawrence Lowland (Woodworth, 1905a, 1905b; Chapman, 1937; Denny, 1974; Franzi and Adams, 1993, 1999; Franzi et al., 2002). Outflow from the breakout of proglacial lakes in the St. Lawrence Lowland was initially directed southeastward along the ice margin where it crossed the English, North Branch and Great Chazy watersheds before eventually emptying into Lake Vermont. The sandstone pavements generally occur on the drainage divides between watersheds where flood scour was greatest and the exposed surfaces were not subsequently covered (Denny, 1974). Altona Flat Rock (Field Trip Stops 3 and 5) is the largest sandstone pavement in the region and lies at the southeastern end of the "Flat Rock" belt where flood water from the Iroquois breakout entered Lake Vermont (Denny, 1974; Franzi and Adams, 1993, 1999) (Fig. 5). Smaller sandstone pavements occur south of Cadyville on the east flank of Burnt Hill where they are associated with outflow channels from proglacial lakes in the Saranac Valley (Denny, 1974) (Fig. 2).

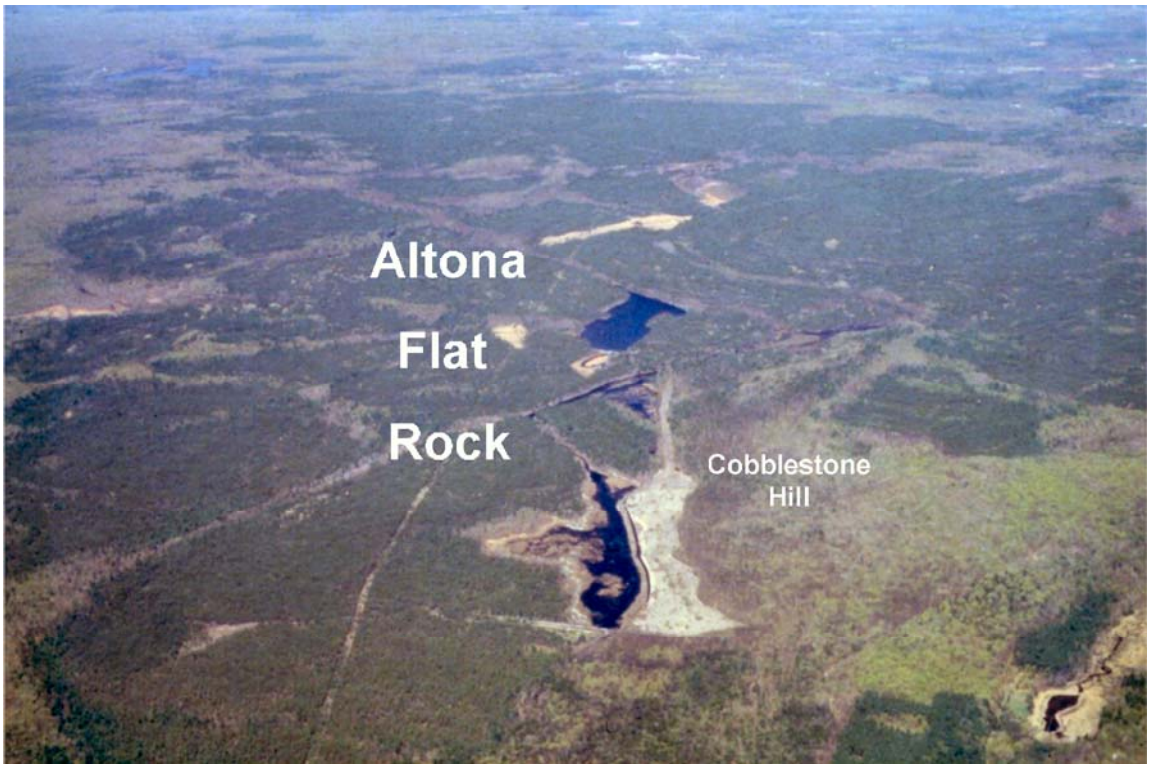

Figure 5. Aerial photograph of Altona Flat Rock and Cobblestone Hill with a view to the northwest. The margin of the sandstone pavement is delineated by the extent of the jack pine barren (dark forest cover) that covers its surface. The remnant of William H. Miner's failed hydroelectric dam is seen in the foreground (Franzi and Adams, 1993, 1999).

Moraines

Denny (1967, 1970, 1974) mapped and described several recessional moraine deposits in the northeastern Champlain Lowland. The largest and most extensive moraines in the region are located in the Saranac Valley near Cadyville and in the Great Chazy River Valley near Ellenburg Depot (Fig. 2). The Ellenburg Moraine (Field Trip Stop 2) consists of a single north-trending ridge that ranges between 300 and 500 meters wide and rises 25–30 m above the surrounding terrain (Denny, 1974). The moraine is composed of interbedded sand, gravel and diamicton that are commonly deformed and offset by normal faults, primarily on its eastern flank. Diamicton interbeds are generally massive to crudely stratified and range from a few decimeters to a few meters in thickness (Denny, 1974; Franzi et al., 1993). The moraine rises to the north where it intersects a low-relief, east-trending recessional moraine north of Clinton Mills (Denny, 1974). A small segment of recessional moraine south of Miner Lake is probably contemporaneous with the Ellenburg Moraine (ice margin 8 of Denny, 1974).

The Cadyville Moraine (the DeKalb Moraine of Taylor, 1924)) consists of a north-trending belt of linear till ridges and knolls and kame sand and gravel bodies that span the Saranac Valley (Fig. 2). The ridges are typically composed of pebbly, sandy till with interbedded sand and gravel (Denny, 1974). Local relief between ridge crest and adjacent swales ranges from a few meters to approximately 20 m and ridge crests are commonly spaced 60–260 m apart. The length of individual ridges typically ranges from a few hundred meters to about 0.5 km (Denny, 1974). The swales and channels that Denny described in northern part of the moraine extend northward to Jericho (Field Trip Stop 4). The southern portion of the Cadyville Moraine terminates against the northeastern flank of Burnt Hill (Fig. 2).

Diamictons

Massive, overconsolidated diamictons, interpreted to be subglacial lodgement or basal meltout till, typically form the basal glacial unit in the Champlain Lowland. Till occurs primarily as a discontinuous (1–3 m thick) veneer over bedrock on hillslopes and upland areas. The underlying bedrock is commonly polished and its surface may contain evidence of ice flow such as striations, friction cracks and crescentric gouges. The color, texture and composition of till deposits in the region are variable and reflect local provenance (Denny, 1974; Craft, 1976; Franzi and Cronin, 1988). Good examples of the wide range in till texture and composition over relatively short distances can be observed at Stops 7, 8 and 9 of this field trip.

Nonglacial diamictons consist primarily of intercalated diamicton and stratified deposits. The diamictons occur as lenticular to planar beds that range from a few centimeters to a few meters thick. Individual diamicton beds are massive to crudely graded and are laterally continuous for a few decimeters to tens of meters. Stratified interbeds range from thin, discontinuous sand, silt and clay laminae to massive, planar bedded and cross-stratified sand and gravel beds more than a meter thick. Bedded diamictons are commonly associated with ice-marginal deposits, such as the Cadyville-Jericho and Ellenburg moraines (Field Trip Stops 2 and 4) (Denny, 1974; Franzi and Adams, 1993; Franzi et al., 1994) and ice-proximal or subglacial lacustrine environments, as are exposed in the Keeseville area at Ausable Chasm (Fig. 6), Mud Brook (Field Trip

Stop 11) and Keeseville Industrial Park (Field Trip Stop 12). A greater relative proportion, thickness, and continuity of diamicton to stratified beds are generally associated with ice-proximal or valley-side environments (Diemer and Franzi, 1988; Franzi et al., 2002).

Figure 6. Bedded diamicton facies overlying Potsdam Sandstone at Ausable Chasm.

Coarse-grained Stratified Deposits

Stratified sand and gravel deposits are associated with fluvial, glaciofluvial, subaqueous outwash fan, deltaic, beach, lacustrine, and marine environments. The texture and structure of these deposits are variable and depend on the nature and energy conditions at the site of deposition.

Deltas and beaches provide important evidence for reconstructing the extent of former proglacial lake and marine shorelines. Deltas commonly occur as gently sloping sandplains at the mouths of tributary valleys. The deposits generally grade upward from ripple cross-laminated to planar bedded, fine to medium sand, to planar bedded and trough cross-stratified, poorly sorted, coarse sand and gravel (Diemer and Franzi, 1988). The deltas may have been fed by meteoric streams from deglaciated upland areas, ice-marginal or proglacial meltwater streams or by outflow streams from proglacial lakes in adjacent valleys. Large lacustrine delta plains, deposited primarily by meteoric streams,

are commonly found where major rivers entered Lake Vermont or the Champlain Sea. Multiple delta terraces attest to the regrading of inflowing streams as proglacial lake or marine levels dropped.

The Ingraham Esker in Clinton County (Field Trip Stop 6) (Fig. 2) is a 25-km-long, roughly north-south trending, sinuous ridge composed primarily of stratified sand and gravel (Woodworth 1905a, 1905b; Fisher, 1968; Denny, 1972, 1974; Diemer, 1988). The ridge ranges from 100 to 300 meters wide and rises 3 to 10 meters above the surrounding terrain (Diemer, 1988). The esker deposits are interbedded with fine-grained lacustrine deposits, including varved clays, and discordantly overlain by fossiliferous gravel, sand, and fine grained marine deposits (Woodworth 1905a, 1905b; Denny, 1972, 1974; Diemer, 1988). Denny (1972, 1974) believed that the ridge formed as an esker in a subglacial tunnel and that its present low relief was due to reworking of the esker deposits by waves and currents in Lake Vermont and the Champlain Sea. Diemer (1988) conducted a detailed sedimentological study of the esker deposits and concluded that the ridge is composed primarily of subaqueous fan deposits. He suggested that the present relief of the esker might be more a primary consequence of subaqueous fan deposition than later resedimentation.

Glacial Sedimentology and Stratigraphy of the Champlain Lowland

Sediment exposures in upland portions of the region are generally scattered and incomplete and thus interpretations of glacial events are derived largely from morphostratigraphic analysis of glacial and lacustrine landforms (e.g. Alling (1916, 1918, 1920; Kemp and Alling, 1925; Craft, 1976; Diemer and Franzi, 1988; Franzi, 1992). Sediment exposures in the Champlain Lowland are more abundant and complete but are often in poor condition because the materials are susceptible to mass movement. Our understanding of the glacial stratigraphy in the Champlain Lowland improved greatly since 2004 when we began to collect vibracore samples from locations with thick sequences of late-glacial sediment (Fig. 7). Our multi-proxy analyses of the cores are on-going; however, preliminary results indicate that the cores contain a high-resolution record of late-glacial events. Cores from which stratigraphic information has been collected to date are Peru, Plattsburgh Air Force Base Marina (PAFB), Plattsburgh State University, Beekmantown, Salmon River, Long Pond and Bulwagga Bay (Fig. 7). The glacial stratigraphy in the Champlain Lowland (Fig. 8) is grouped into four principal stratigraphic units that are defined by sediment-facies associations and microfossil assemblages (Rayburn et al., in prep.). Diamicton (till) underlies the stratigraphic section at most locations and is overlain by proglacial lake and marine sediments deposited in glacial Lake Vermont and the Champlain Sea, respectively. Contained within the lacustrine and marine sediments are deposits associated with proglacial lake breakout events from the Great Lakes or Agassiz basins that were routed through the Champlain or St. Lawrence lowlands.

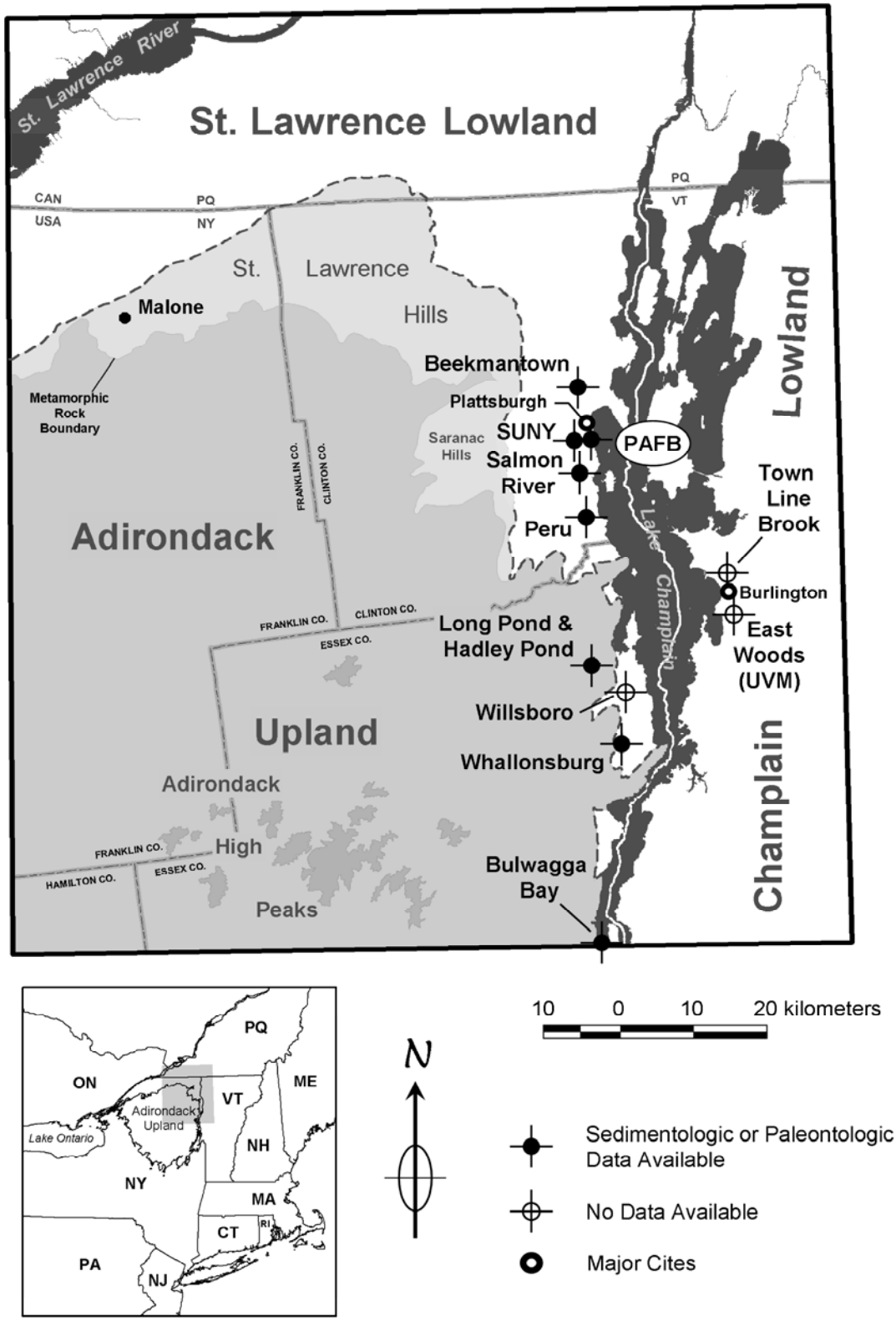

Figure 7. Map of northeastern New York showing sediment-core locations.

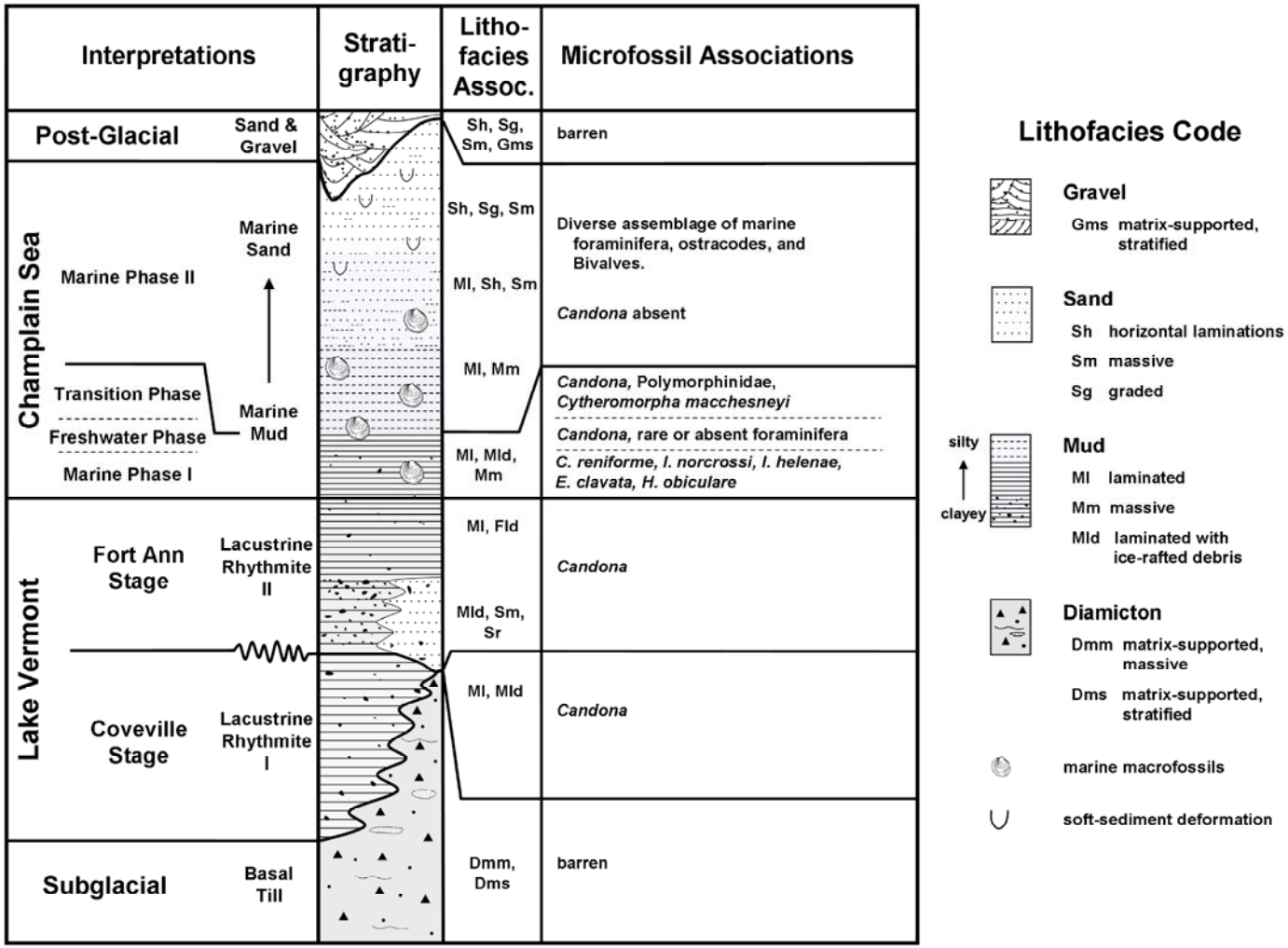

Figure 8. Generalized and preliminary stratigraphic section of late Pleistocene deposits from land-based cores and surface exposures in the Champlain Lowland of northeastern New York.

Basal Diamicton (Till)

The basal till in sediment cores from the Peru, PAFB, PSU, Beekmantown and Salmon River cores consists of 0.5–1.5 m of generally dark gray, calcareous, massive, overconsolidated, clast-rich diamicton. The contact with the overlying lacustrine clay rhythmites unit is sharp. The bedded diamicton facies in exposures near Keeseville (Field Trip Stops 11 and 12) is more than 30 meters thick at Mud Brook. These deposits probably represent ice-proximal or subglacial, interstratified sediment-flow (Evenson et al., 1977; Lawson, 1981) and rain-out (Eyles and Eyles, 1983) diamictons, subaqueous outwash (Rust and Romanelli, 1975; Rust, 1987), and pelagic lacustrine deposits.

Lacustrine Rhythmite Facies I

The lower lacustrine rhythmite facies consists of alternating, millimeter to centimeter scale, fine silt and clay laminations that contain ice-rafted sediment clasts and dropstones. Rhythmite thickness, particle size and carbonate content generally decrease upward through the section. The unit is barren of fossils with the exception of sparse occurrence of the freshwater ostracode *Candona subtriangulata*. The unit is more than 16 m thick at Whallonsburg but thins northward and is not found north of Beekmantown. The base of the lower rhythmite facies rests sharply over the underlying diamicton.

Lacustrine Rhythmite Facies II

The lower rhythmite facies transitions abruptly into the upper rhythmite facies, but the contact between the units varies from south to north in the Champlain Lowland. The contact in sediment cores from Whallonsburg, in the southern portion of the basin, is marked by a sharp change from millimeter-scale silt and clay laminations that contain less than 2% total carbonate and little ice-rafted debris (IRD) to millimeter- and centimeter-scale laminations with abundant IRD and carbonate content between 2 and nearly 10%. The IRD-rich, calcareous rhythmite facies is about 0.6 m thick at Whallonsburg and contains approximately 50 sediment couplets. The base of the upper lacustrine unit in cores from the northern Champlain Lowland is marked by a 0.1–1.5 m thick layer of medium to fine sand that rests unconformably over the lower rhythmites. The sand contains rip-up clasts from the underlying rhythmite unit and is generally massively bedded, but grades upward to ripple-cross-laminated sand near its upper contact in the Peru core. The spatial and stratigraphic distribution of sediment facies at the base of the upper lacustrine unit is consistent with an abrupt, short term (≈ 50 years if the rhythmites are annual, i.e. varves) influx of discharge from the north and the southward transport of icebergs containing calcareous debris through the Champlain Lowland.

The IRD-rich, calcareous rhythmite and massive sand facies at the base of the upper lacustrine unit are conformably overlain by 2–5 m of millimeter-scale silt and clay rhythmites that become thinner, finer-grained and less calcareous upward through the unit. The freshwater ostracode *Candona* is the only microfossil found in the unit. The upward trends in lamina thickness, texture, and carbonate content are consistent with a gradual transition from an ice-proximal to ice-distal lacustrine environment. The base of the unit in some cores from the northern Champlain Lowland consists of red-brown and

gray rhythmite couplets that are similar to those described by Godsey et al. (1999) and shown to be annual (varves). A total of 168 rhythmite couplets were counted in the Peru core.

Marine Mud Facies

The base of the marine mud facies is marked by a 0.2-m-thick zone of silt and clay laminae with occasional fine sand partings that transitions upward into massive, to thinly bedded mud. Weak laminae consisting of fine black organic material and small bits of macerated organic material occur throughout. The facies is further defined by the first occurrence of marine microfauna, most notably foraminifera. Four distinct microfaunal assemblage zones have been identified in the marine mud facies: Marine I Zone (lower), Freshwater Zone, Transition Zone and Marine II Zone (upper) (Fig. 8). The lower marine zone is characterized by a high-diversity marine fauna dominated by *Cassidulina reniforme* and including *Islandiella norcrossi*, *I. helenae*, *Elphidium clavata*, and *Haynesina orbicularis*. The freshwater zone is characterized by the return of *Candona* and the near absence of marine forams. The freshwater zone is composed of thinly laminated (0.1–1.0 cm thick) rhythmites that are similar to rhythmites in the underlying upper lacustrine unit. Approximately 120 sediment couplets occur in the freshwater zone in the Beekmantown core and a thin red-brown clay bed or lamination occurs near its upper contact in the PAFB and Beekmantown cores. The significance of this layer is not clear because sources of red sediment occur locally and red clay clasts and laminations occur in the underlying lacustrine rhythmites, but its occurrence within the freshwater zone of the marine mud facies is unique and may indicate an external source, perhaps the upper Great Lakes or Agassiz basins (Rayburn et al., in prep.). The overlying transition zone contains a mixed assemblage of *Candona*, Polymorphinidae foraminifera and the brackish-water ostracode *Cytheromorpha macchesneyi*. The transitional unit is approximately 0.9 m thick at Salmon River and consists of thinly laminated clay rhythmites. Rayburn et al. (in prep) suggest sedimentation rates were 0.22–0.8 cm yr^{-1}, if the rhythmite couplets represent annual layers. Black organic laminations mark the boundary between the transition zone and the upper marine zone.

Marine Sand Facies

The upper marine facies grades upward from laminated and thinly bedded mud to thinly bedded medium–fine sand with occasional mud interbeds and laminations. Individual sand beds are 0.01–m thick and are commonly normally graded, horizontally laminated or massive. Flame structures, sediment diapirs and other soft-sediment deformation structures are found occasionally in outcrop. Fossil abundance decreases upward through the section.

Sand and Gravel Facies

The marine sand facies is unconformably overlain by 0–2 m of cross-bedded, horizontally bedded and massive, pebbly coarse sand, sandy gravel and coarse–medium sand. Beds range from several centimeters to more than a meter thick and extend laterally for several meters parallel to the paleocurrent. The base of most beds, especially gravel beds, is marked by erosional troughs cut into underlying units. This facies is best exposed in a gravel pit on the Salmon River in Plattsburgh where the gravel surface is

terraced, but it is missing at the PAFB marina, Salmon River and Beekmantown core sites.

Glacial History

East-Central Adirondack Region

The Ausable and Boquet watersheds are north-draining basins located in the east-central Adirondack Upland and west-central Champlain Lowland (Fig.4). Deglacial drawdown of ice into valley ice streams blocked north-directed runoff and impounded deep proglacial lakes (Diemer and Franzi, 1988). Alling (1916, 1918, 1919, 1920; Kemp and Alling, 1925) developed the first proglacial lake chronology for the Ausable and Boquet valleys. Diemer et al. (1984), Diemer and Franzi (1988) and Franzi (1992) recognized that deglaciation occurred by generally synchronous recession of active ice lobes from the Champlain Lowland and that many fluvial and lacustrine deposits originally thought to be related to regional lakes were graded to local, ice-marginal impoundments in tributary valleys. Many of the lake stages originally proposed by Alling (1916, 1918, 1919, 1920) and Kemp and Alling (1925) have been abandoned or renamed to reflect the location of their presumed outlets (Franzi, 1992). Northward expansion of proglacial lakes in the Ausable and Boquet valleys was punctuated by catastrophic lake-drainage events. Upland proglacial lakes in the lower portions of Ausable and Boquet watersheds were eventually succeeded by Lake Vermont and later by the Champlain Sea as ice receded northward in the Champlain Lowland.

A portion of a musk-ox (*Ovibos moschatus*) vertebra was found in 1986 in prodeltaic rhythmites exposed in a gravel pit near Elizabethtown (Fig. 9) (Rayburn et al., 2007b). The rhythmites were deposited in Lake Hoisington, a small proglacial lake in the Black River Valley that was controlled by an outlet approximately 10 meters above the level of the Coveville Stage of Lake Vermont (Fig. 10). Coarse deltaic sediments were deposited over the rhythmites by outflow associated with the breakout of Lake Underwood in the Boquet Valley and its subsequent drainage to the Lake Elizabethtown level. The bone yielded ages of 11,280 ± 110 ^{14}C years B.P. (AA-4935, purified collagen fraction) and 10,750 ± 800 ^{14}C years B.P. (humin fraction), which with appropriate δ^{13}C correction is equivalent to 13,438–13,020 calendar years B.P (Rayburn et al., 2007b). The ice margin at this time was fronted by the Coveville Stage of Lake Vermont in the mid-Champlain Lowland and correlates with proglacial lakes Elizabethtown and Chapel in the Boquet and Ausable valleys, respectively (Fig. 7).

Recession of the ice margin farther north in the Keene Valley resulted in the successive lowering of lake levels in the Keene Valley, probably by sudden breakout, as lower outlets on the Ausable–Boquet drainage divide were uncovered (Diemer and Franzi, 1988; Franzi, 1992). The Channel Belt (Kemp and Alling, 1925) (Fig. 2) became active during this period and outflow was directed into the upper North Branch Boquet River. The ice-marginal channels and ice-contact stratified drift deposits at the Oak Hill wollastonite quarry (Field Trip Stop 14) are probably associated with outflow through the South Gulf channel (Franzi, 1992).

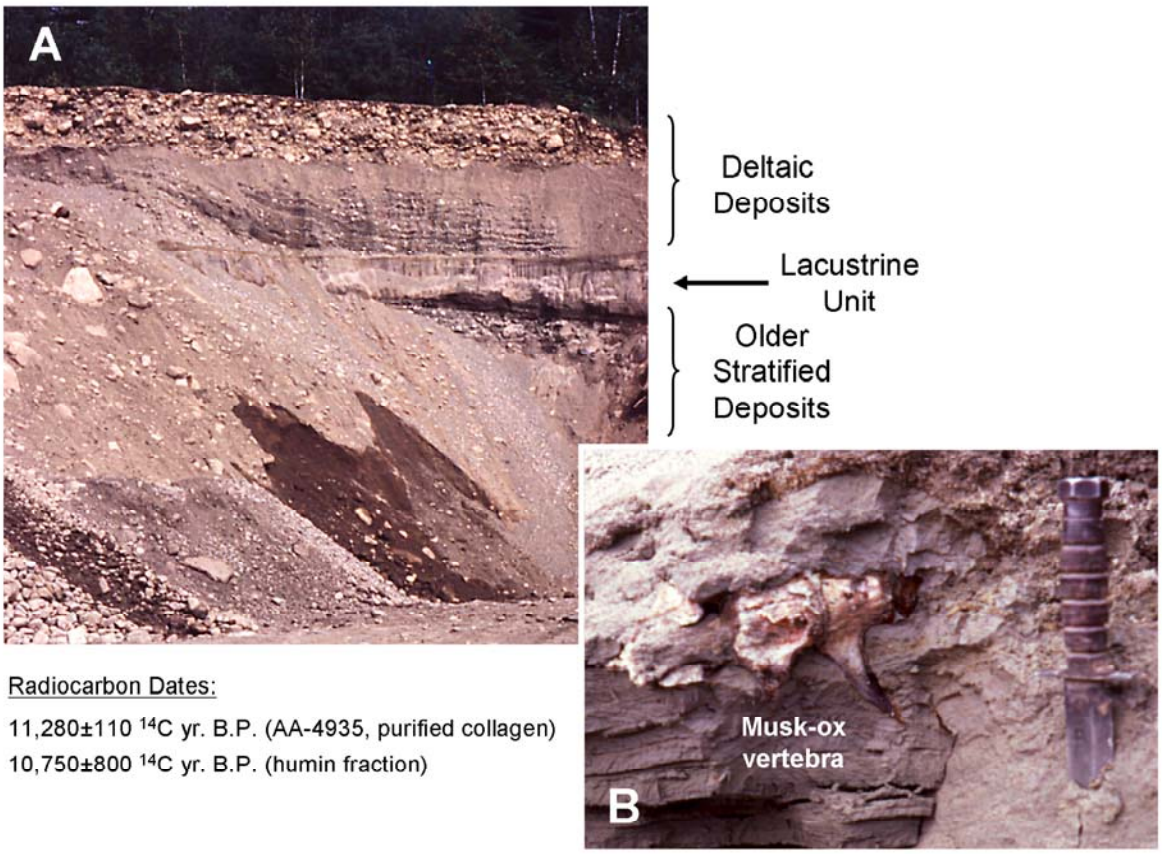

Figure 9. A. Elizabethtown gravel pit where a musk-ox (*Ovibos moschatus*) bone was found. B. Musk-ox vertebra buried in proglacial lake rhythmites (knife handle is approximately 10 cm long).

Northeastern Adirondack Region

Principal drainage systems in the northeastern Adirondack region include the Saranac and Chazy watersheds, which drain eastward to Lake Champlain and the Chateaugay, Trout, and Salmon watersheds, which drain northward to the St. Lawrence River (Fig. 4). Meteoric drainage from freshly deglaciated headwater regions was directed toward the receding ice front, which impounded local proglacial lakes. Proglacial lake levels and drainage routes changed as lower outlets were exhumed with further ice recession. For example, early drainage in the upper Saranac Valley (e.g. ice margin 1 of Denny, 1974) was directed westward across the Adirondack Upland but shifted to the southeast as the ice front ice retreated to the mid-Saranac Valley (ice margins 2 and 3b of Denny, 1974). Sand plains above 444 meters elevation in the Saranac Valley near Redford represent deltaic deposits graded to outlets and bedrock channel systems in the upper Salmon and Ausable watersheds. Outflow from these lakes was directed along the ice front southward through the Black Brook channel (Fig. 2) to proglacial lakes in the Ausable Valley (Diemer and Franzi, 1988; Franzi, 1992).

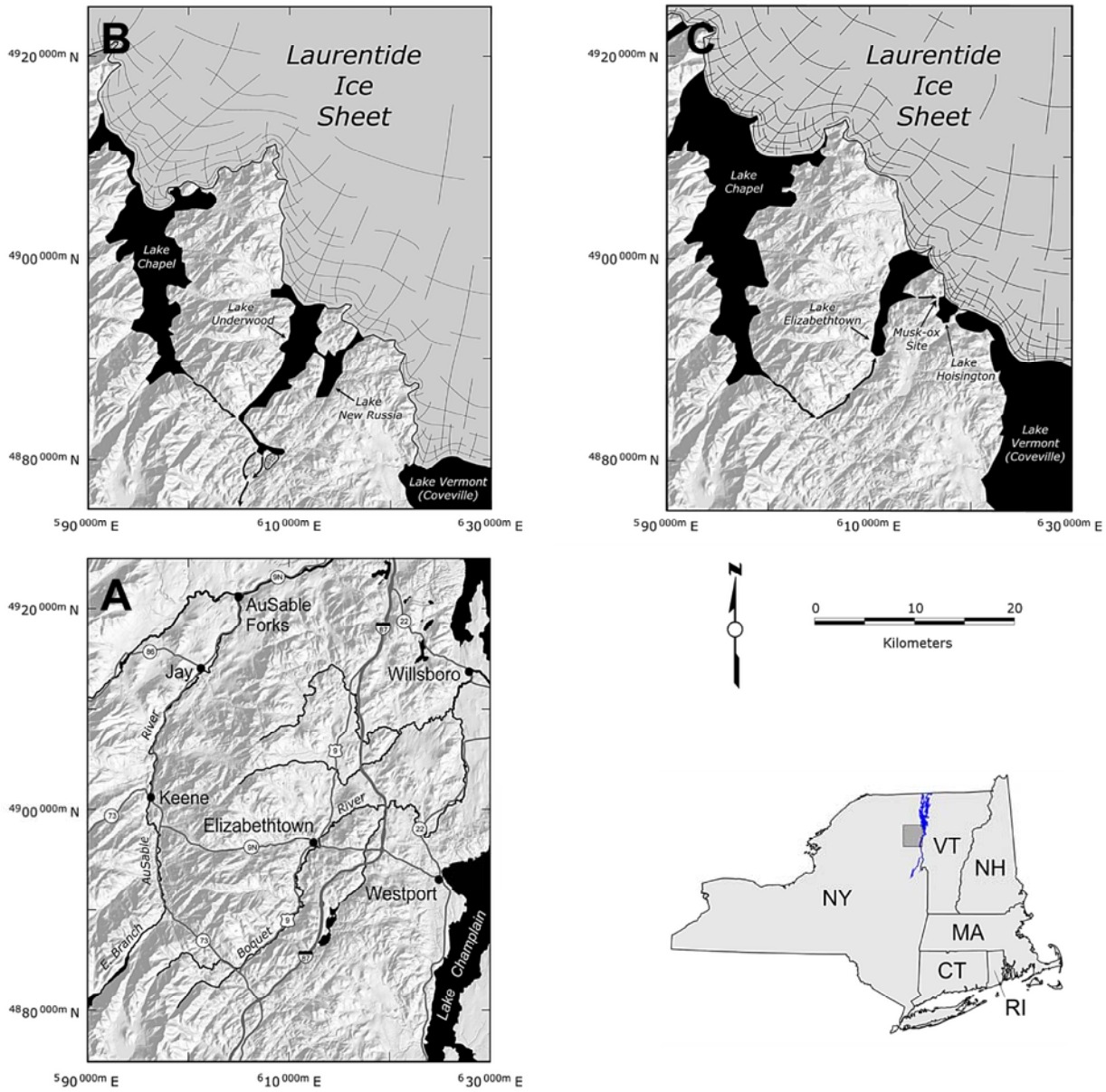

Figure 10. Shaded relief map of the Ausable and Bouquet valleys in the Elizabethtown–Westport area (from Rayburn et al., 2007b). The map was constructed from USGS 1:25,000 digital elevation models. The coordinates are Universal Transverse Mercator (UTM), Zone 18N, North American Datum 1927 (NAD27). A. Map showing the locations of principal towns, roads, rivers and water bodies. B. Reconstruction of the paleogeography and ice margin immediately preceding the break out of proglacial Lake Underwood. C. Paleogeography and ice margin following the breakout of Lake Underwood. The Underwood breakout was responsible for the deltaic deposits that prograded over the lacustrine sediment containing the musk-ox bone.

Cadyville Moraine

Ice recession in the Saranac Valley halted long enough to build the Cadyville Moraine east of Plattsburgh (ice margin 4b of Denny, 1974) (Fig. 2). The moraine deposits and associated ice-marginal meltwater channels occur in an arcuate, concave-eastward belt 1.5 to 4.0 km wide that stretches from Jericho (Field Trip Stop 4) at its northern end to the southern Saranac River drainage divide south of Cadyville. The Saranac River occupies a narrow postglacial channel that cuts through the moraine on the valley floor. The moraine lies on top of a prominent bedrock step at the head of Kent Falls Gorge, where the Saranac River descends more than 100 meters into the Champlain Lowland (Denny, 1974). The rock may have acted as a buttress that helped stabilize the ice front at this location when the moraine formed. Ice-marginal channels cut into moraine deposits on the northern margin of the Saranac ice lobe directed drainage southwestward toward the Saranac Valley. Small sand plains at the southern end of the channel system may represent deltas built into proglacial lakes in the Saranac Valley. Outflow from these lakes scoured rock pavements and cut ice-marginal channels on the east slope of Burnt Hill. Hanging channel heads at elevations between 218 and 335 m on the drainage divide south of Cadyville probably represent semi-stable lake outlet thresholds. Outflow from Saranac Valley lakes discharged southward into proglacial lakes in the Salmon River Valley, which in turn discharged around the ice margin on the east flank of Terry Mountain and into the Coveville Stage of Lake Vermont in the Champlain Lowland west of Peru. Large deltaic sand plains in the Saranac Valley east of Redford were deposited by the Saranac River into the western end of the proglacial lakes controlled by the outlets south of Cadyville.

Denny (1974) correlated the Cadyville Moraine with ice-contact stratified deposits and ice-marginal channel systems in the headwaters of the North Branch Chazy River south of Ellenburg Center and in the Chateaugay Valley near Brainardsville (ice margin 4a of Denny, 1974). The ice front blocked north-draining river valleys such as the Chateaugay, Trout, and Salmon and impounded proglacial lakes that drained initially southward and ultimately westward to the Ontario basin during the early stages of ice retreat. Ice recession from the northern flank of the Adirondack Upland may have been interrupted by at least one readvance. MacClintock and Stewart (1965) used till stratigraphy in St. Lawrence Seaway exposures and the spatial distribution of till properties to propose late-glacial (Malone and Fort Covington) readvances. Clark and Karrow (1983) later demonstrated that till facies changes were controlled by underlying bedrock composition and that only one surface till unit exists. Denny (1974) believed that the ice front may have receded northward to near the international border before it readvanced to the southern limit of the Chateaugay Channel system (ice margin 6 of Denny, 1974). He described a complex sequence of initial eastward, then westward and resumed eastward meltwater drainage across the St. Lawrence-Champlain drainage divide associated with the readvance and subsequent ice-front recession. The early eastward drainage phase was invoked to explain the occurrence of thin till and small rock pavements on the east side of the divide near Ellenburg (Denny, 1974). The reader is referred to Denny (1974) for a detailed account of these drainage changes. We note, however, that Denny's evidence for early eastward drainage is largely circumstantial and given the lack of evidence for a readvance on the west side of the divide (Clark and

Karrow, 1983), a single uniform recession from the northern Adirondack region may be more plausible.

Chateaugay-Churubusco Channels

Ice-marginal and proglacial lake drainage were rerouted westward to Lake Iroquois in the Ontario and upper St. Lawrence lowlands as the ice front receded from the northern margin of the Adirondacks Upland and into the low- to moderate-relief St. Lawrence Hills (Cressey, 1977) south of Malone and Chateaugay (Fig. 2). The drainage changes occurred progressively from southwest to northeast along the Adirondack front and were accompanied by the northeastward expansion of Lake Iroquois as the ice withdrew northward. The Chateaugay Channels (MacClintock and Stewart, 1965; Denny, 1974; Clark and Karrow, 1984; Cadwell and Pair, 1991) (Field Trip Stop 1) are a well-developed system of long, narrow, west-sloping ice-marginal channels that are cut deeply into glacial deposits in a narrow 5-km-wide band south and southwest of Chateaugay. The largest channels range from about 7 m to more than 20 m deep and 90 m to more than 120 m wide, and their bottoms are commonly covered with a boulder lag (MacClintock and Stewart, 1965). The highest channels originate at elevations just above 400 m but none extend below the shoreline of Lake Iroquois (MacClintock and Stewart, 1965; Denny, 1974; Rayburn, 2004), which implies that Lake Iroquois provided base-level control. The channel system is transected at nearly right angles by modern drainage channels (i.e. Trout, Salmon and Chateaugay rivers) that are cut deeply through the underlying glacial deposits and Potsdam Sandstone.

The channel system extends eastward to the St. Lawrence-Champlain drainage divide south of Churubusco. Cols along the divide at elevations of 407 m, 404 m and 337 m may have served as outlet thresholds for west-draining proglacial lakes in the headwaters of the Chazy River on the east side of the divide (Denny, 1974). Lake outflow from the highest of these thresholds may have augmented westward-directed meltwater runoff from the ice and meteoric runoff from northern Adirondack valleys when the Chateaugay Channels southwest of Chateaugay were cut. Channels in the Churubusco area were cut after the ice front receded farther northward.

The Chateaugay–Churubusco Channels were abandoned when a lower threshold was exhumed along the St. Lawrence-Champlain drainage divide and proglacial lake outflow in the St. Lawrence Lowland was directed eastward to the Champlain Lowland. Clark and Karrow (1984) suggested that a col north of Clinton Mills (elevation 326 ± 2 m) served as a short-lived threshold for regional proglacial lake Level I in the St. Lawrence Lowland, which was confluent with Lake Iroquois in the Ontario basin. Pair et al. (1988) and Rayburn (2004) found that Lake Iroquois strandlines fall below the Clinton Mills col and concluded that the threshold in the western Mohawk Valley controlled Lake Iroquois until the threshold at Covey Hill gap (elevation 308 ± 4 m; MacClintock and Terasmae, 1960) was opened and Lake Iroquois drained to the Lake Frontenac level.

Ellenburg Moraine

The Ellenburg Moraine is a narrow, 13 km long, generally north-south trending ridge of ice-contact stratified drift and till that was deposited in the upper North Branch Chazy River and English River valleys (Denny, 1974) (Field Trip Stop 2). The moraine is more

than 0.5 km wide and rises more than 30 meters above the valley bottom where it is incised by the North Branch near Ellenburg Depot. The ridge becomes subdued and eventually untraceable as it rises into higher terrain to the north and south. The moraine was deposited into a small proglacial lake that may have drained westward across the Clinton Mills col and into Lake Iroquois in the St. Lawrence Lowland. Denny correlated the Ellenburg Moraine with a small, east-west trending recessional moraine north of the Clinton Mills col and a recessional moraine segment in the Great Chazy River Valley south of Miner Lake (ice margin 8 of Denny, 1974). He found no evidence for eastward drainage of the North Branch-English proglacial lake. The ridge crest in the valleys, however, lies between 290 and 299 meters in elevation, well below the threshold elevation at Clinton Mills col. Thus, either the moraine was deposited in deep water, as suggested by Denny (1974), or proglacial lakes along the ice front drained eastward via small channels on the northern and eastern slope of Rand Hill in the Champlain Lowland. In the latter case, outflow from these lakes may have contributed to incision of the ice-marginal channels and the deposition of ice-contact stratified deposits near Beekmantown in the Champlain Lowland. In either case, the Ellenburg Moraine and associated deposits immediately predate the opening of the Covey Hill threshold and the breakout of Lake Iroquois.

Proglacial Lakes in the St. Lawrence and Champlain Lowlands

Lakes Iroquois and Vermont (Coveville Stage)

Rayburn (2004) and Rayburn et al. (2005) reevaluated strandline data for regional proglacial lakes and the Champlain Sea in the St. Lawrence and Champlain lowlands (Fig. 11). Isobases generally trend east-west in the region and are inclined southward with a linear gradient in the range of 0.7 m/km to 1.0 m/km. The highest regional strandline in the St. Lawrence Lowland, Lake Iroquois, can be traced northward to the Chateaugay-Churubusco channels (MacClintock and Stewart, 1965; Denny, 1974; Pair et al., 1988; Cadwell and Pair, 1991) and projects southward to a threshold in the western Mohawk Lowland near Rome. In the Champlain Lowland, the Coveville Stage of Lake Vermont expanded northward with ice recession. Franzi et al. (2002) traced Coveville Stage strandline deposits along the western flank of the lowland northward to Cobblestone Hill on the southeastern margin of Altona Flat Rock (Fig. 2). The Coveville strandline projects above the Champlain–Hudson drainage divide, which indicates that base-level control for this lake stage was located south of the Champlain Lowland.

Lake Iroquois Breakout

Ice recession from the Covey Hill threshold allowed Lake Iroquois to break out across the St. Lawrence–Champlain divide to Lake Vermont (Coveville Stage) in the Champlain Lowland. The breakout lowered water level in the St. Lawrence Lowland by 14–15 meters and released an estimated 570 ± 85 km^3 from storage in Lake Iroquois (Rayburn, 2004; Rayburn et al., 2005). The flood water was directed southeastward along the ice margin (ice margins 9–11 of Denny, 1974) where it eroded previously deposited surficial material and created the sandstone pavements that comprise the Flat Rocks in Clinton County (Woodworth 1905a; Denny, 1974; Franzi et al., 2002) (Fig. 12).

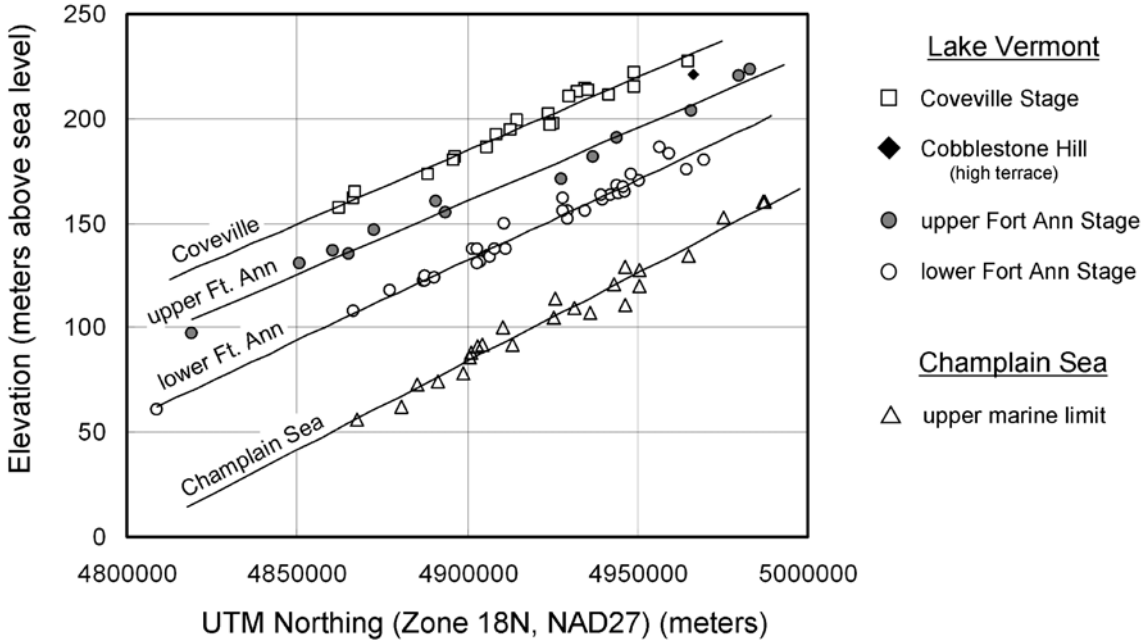

Figure 11. Isostatic rebound profiles for Lake Vermont and the Champlain Sea in the Champlain Lowland (after Rayburn et al., 2005).

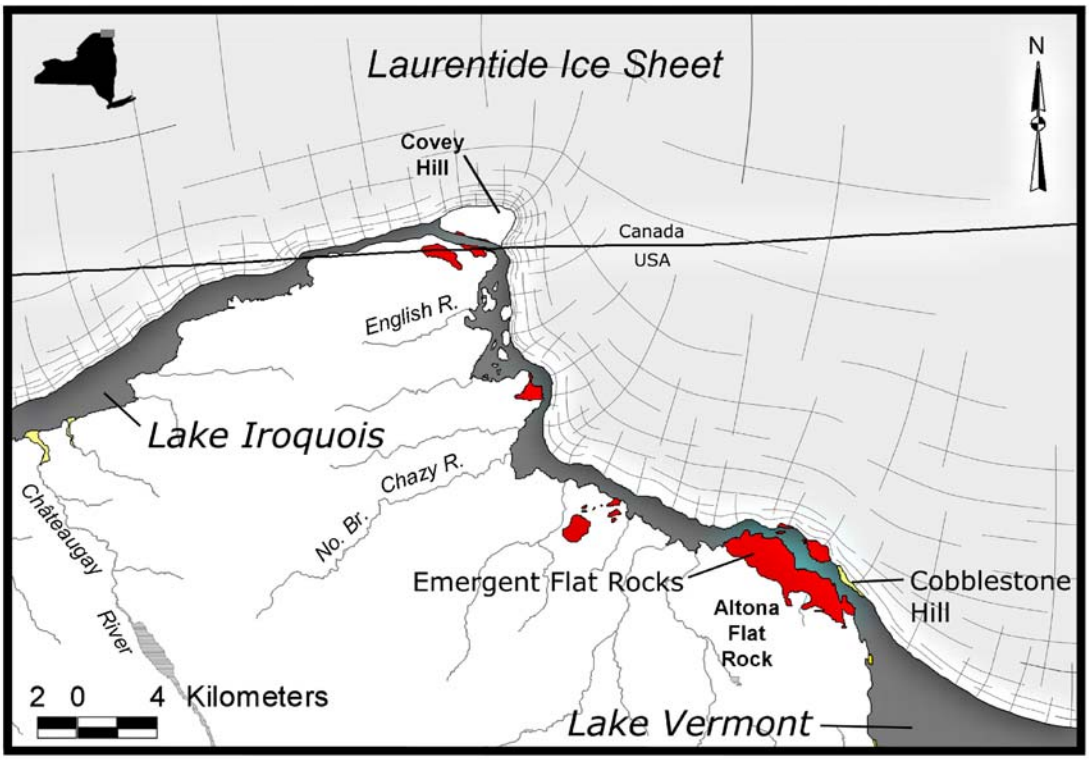

Figure 12. Paleogeographic map showing the Cobblestone Hill ice margin at the time of the Lake Iroquois breakout (from Rayburn et al., 2005).

Large, coarse-boulder deposits were laid at the mouth of the flood channel where it emptied into Lake Vermont (Fig 13). The largest of these, Cobblestone Hill (Field Trip Stop 5), is an elongate ridge more than 15 meters high, 500 m wide, and 2.5 km long that is composed of large angular boulders, almost exclusively Potsdam Sandstone, that range from 0.5 m to more than 3 m in diameter (Franzi et al., 2002). The boulders were deposited against the ice margin as evidenced by large kettle holes on the north flank of the ridge. The general southeastward decrease in boulder size and the occurrence of imbricate boulders that dip generally northwest provide additional evidence of a fluvial origin for the deposit. Rayburn (2004) and Rayburn et al. (2005) estimated that flow velocity must have been in the range of 8.1–8.9 m sec^{-1} to transport the largest (2.5–3.0 m-diameter) boulders on the surface of Cobblestone Hill. They further concluded that the Lake Iroquois breakout flood discharge was in the range of 83,000–92,000 m^3 sec^{-1}, based upon estimates of channel dimensions and velocity at the Cobblestone Hill ice margin. If this discharge is close to the average for the event, the duration of the Lake Iroquois breakout would have been approximately 2.5 months (Rayburn et al., 2005).

Figure 13. Large boulders on Cobblestone Hill near the point where the Iroquois breakout flood entered Lake Vermont on the southeastern margin of Altona Flat Rock.

Franzi et al. (2002) and Rayburn et al. (2005) identified two distinct depositional surfaces on Cobblestone Hill. The highest surface lies at an elevation between the projected strandlines of the Coveville and highest Fort Ann stages of Lake Vermont and the lower near the projected level of the highest Fort Ann strandline (Fig. 9). These data indicate that the level of Lake Vermont dropped while the Cobblestone Hill boulders were deposited against the ice margin. Franzi et al. (2002) and Rayburn et al. (2005) suggested that the of influx of floodwater from the Lake Iroquois breakout overwhelmed the dam that impounded Lake Coveville, drained Lake Albany in the Hudson Lowland and lowered Lake Vermont in the Champlain Lowland to the Upper Fort Ann level. Lake Vermont dropped nearly between the Coveville and Upper Fort Ann levels, releasing 130 ± 20 km^3 from lake storage in the Champlain Lowland. The combined lake breakouts in the St. Lawrence and Champlain lowlands sent approximately 700 ± 105 km^3 of proglacial lake water to the North Atlantic via the Hudson Lowland.

Lake Frontenac and "The Gulf"

After the initial breakout event, proglacial lake level in the St. Lawrence Lowland stabilized at the Frontenac level, which drained across the Covey Hill threshold (elevation 308 m) (Prest, 1970; Denny, 1974; Pair et al., 1988; Rayburn et al., 2005). Outflow from Lake Frontenac cut The Gulf, a narrow, 1.5-km-long gorge cut deeply (max. depth is approximately 45 m) into the underlying Potsdam Sandstone. The uppermost reach of the outflow channel is a broad, low-relief bedrock trough, approximately 1.7 km long and 500–800 m wide that is presently occupied by a large bog. The channel descends abruptly to the east, more than 30 meters, into an ovate depression that is open to The Gulf. The depression marks the location of an abandoned waterfall and plunge pool at the time Lake Frontenac outflow ended and the headward erosion of The Gulf ceased (Figs. 14 and 15). Lake Frontenac drained when the ice front receded from the north flank of Covey Hill and the proglacial lake in the Ontario–St. Lawrence lowlands merged with the Upper Fort Ann level of Lake Vermont. The drop from Lake Frontenac to Lake Vermont lowered lake level in the St. Lawrence Lowland by about 76 meters and released an estimated 2500 ± 375 km^3 from lake storage (Rayburn et al., 2005). Although the volume of water released by the drainage of Lake Frontenac is more than four times greater than the Lake Iroquois breakout, there is little geomorphic evidence in the St. Lawrence and Champlain lowlands for a catastrophic breakout. That evidence may be hidden beneath younger lacustrine and marine sediments or the drainage of Lake Frontenac may have occurred as a wide scour under, or through the thinning ice front (Rayburn et al., 2005). Outflow from the drainage of Lake Frontenac also may have facilitated incision of the threshold at Fort Ann, thus causing the drainage and stabilization of Lake Vermont at the Lower Fort Ann level.

Lake Frontenac outflow represented the cumulative drainage from much of the upper Great Lakes basin. Rayburn et al. (2005) estimated the sustained steady-state outflow from the combined Champlain–Great Lakes basins through the Fort Ann channel to be about 56,000 m^3 sec^{-1}, which is close to the average of 31,000–59,000 m^3 sec^{-1} estimated by Wall and LaFleur (1995) and Wall (1996) for Lake Iroquois outflow through the Mohawk Lowland. These estimates are probably similar to the average discharge magnitude responsible for the erosion of The Gulf at Covey Hill.

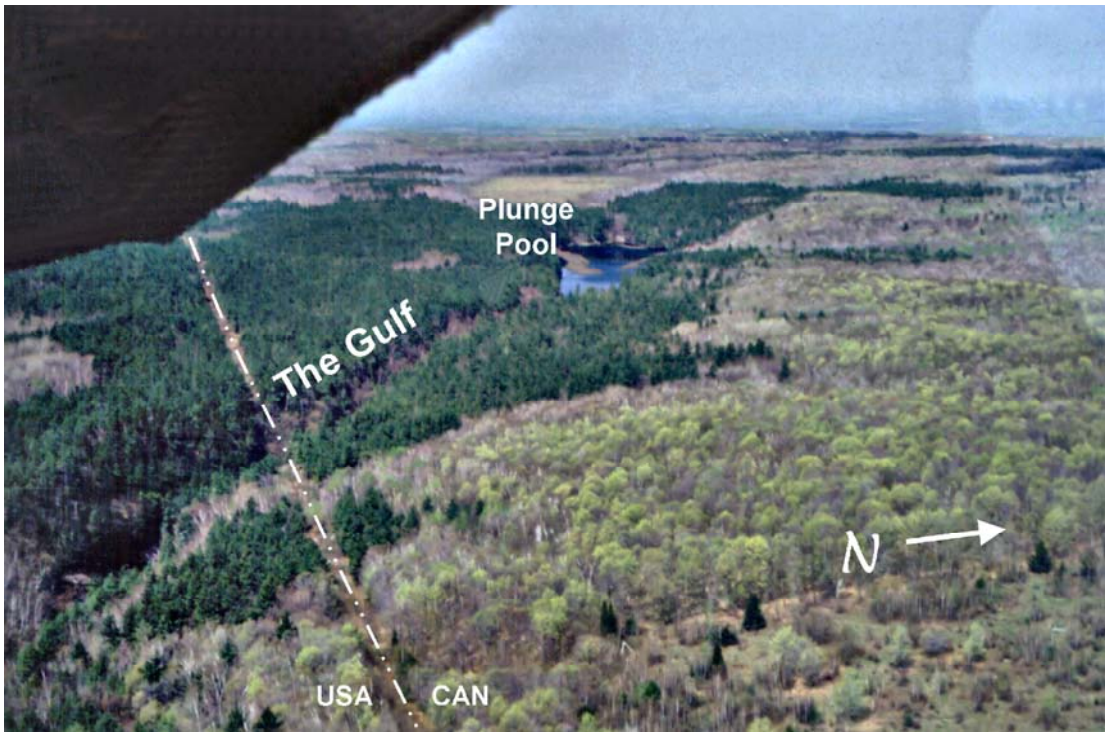

Figure 14. Aerial photograph of The Gulf at Covey Hill.

Figure 15. The plunge pool at the head of The Gulf.

The opening of the Covey Hill outlet and subsequent confluence of lakes in the St. Lawrence and Champlain lowlands temporally constrain the duration of Lake Frontenac. We propose that Lake Frontenac existed for approximately 50 years, based upon indirect evidence from the varve stratigraphy in cores collected at the Whallonsburg landslide (Fig. 16). We have completed preliminary analyses of these cores, which show changes in carbonate content that may be related to the provenance of the glaciolacustrine sediment. We interpret the steady decrease in carbonate upward through medium to thinly laminated, debris-poor varves from the 1000–830 cm depth interval as the result of ice recession from the southern Champlain Lowland and a gradual change from glacial and iceberg-rafted sources of calcareous sediment to low-carbonate sediment derived largely from local meteoric runoff from the Adirondack or Green mountains. The varves in this interval probably represent the Coveville Stage of Lake Vermont and runoff at this time is derived entirely from meteoric and ice-melt runoff within the Champlain drainage basin.

We believe that the increases in the amount of coarse debris, varve thickness and carbonate content observed in the interval 830–760 cm (Fig. 16) represents the influx of calcareous ice-rafted debris (IRD) from drainage of lakes Iroquois and Frontenac in the St. Lawrence Lowland. The sudden release of water across the St. Lawrence–Champlain divide would have accelerated calving of the ice margin and more icebergs would have moved southward in the Champlain Lowland with the additional outflow from the St. Lawrence and Great Lakes basins. The initial increase probably represents the Lake Iroquois breakout, which occurred when the ice margin stood at the Cobblestone Hill ice margin and proglacial lake level in the Champlain Lowland dropped to the Fort Ann level. The relatively thick, debris-rich, calcareous varve interval spans 48 varve years, so it is likely that the Lake Frontenac drainage is also represented. Carbonate content remains relatively high but less variable in the medium to thinly laminated, debris-poor, calcareous varves in the interval from 760 to 500 cm (Fig. 16). This interval may represent post-lake-drainage sedimentation when proglacial lake outflow from the Great Lakes was routed through the Champlain Lowland. If these estimates are correct, erosion of The Gulf occurred over a 50-yr period at an average rate of about 300 m yr^{-1} and Lake Frontenac had the same duration.

The drainage of lakes Iroquois and Frontenac is recorded by 0.3–1.5 m-thick, medium to coarse sand layers in varve sequences in sediment cores from the northern Champlain Lowland near Plattsburgh (Fig. 7). The sand layers commonly contain clay rip-up clasts and are generally massive but may grade into ripple cross-laminated sand near their upper contacts. The sedimentary sequences recovered in these cores do not allow us to differentiate individual breakout events. It is possible that the second breakout, Lake Frontenac, eroded evidence of the earlier Lake Iroquois flood.

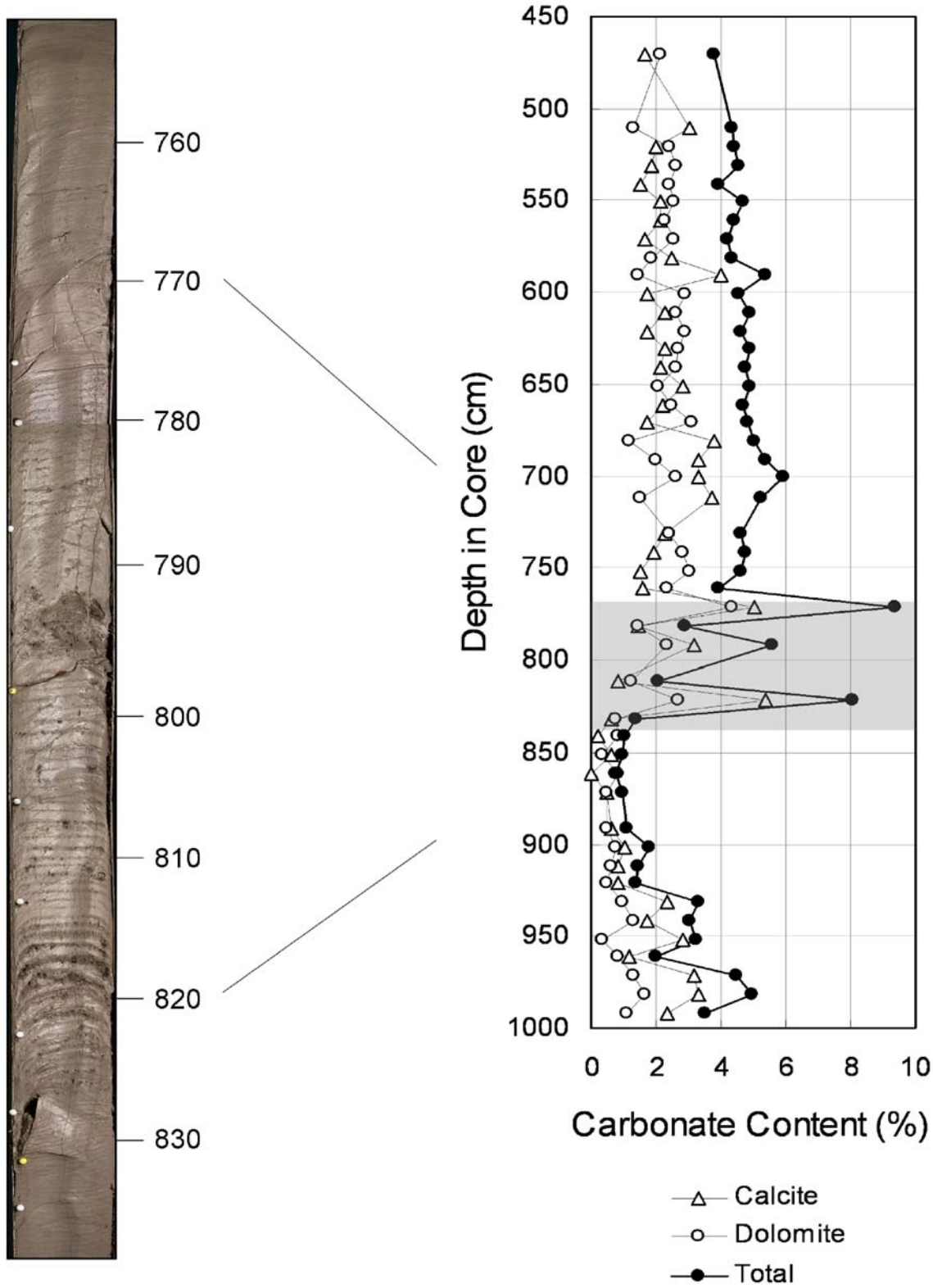

Figure 16. Section of the Whallonsburg core that depicts changes in rhythmite thickness, texture and carbonate content associated with the Lake Iroquois breakout.

The Champlain Sea

Withdrawal of the ice margin from the north flank of Covey Hill resulted in the drainage of Lake Frontenac and the merging of proglacial lakes in the St. Lawrence and Champlain lowlands at the Lower Fort Ann level. This lake persisted, with outflow to the mid-Atlantic region via the Hudson Lowland, for approximately 150–200 years (Franzi et al., 2002; Rayburn, 2004; Rayburn et al., 2005) until ice recession opened the mouth of the lower St. Lawrence Lowland, and as a result Lake Vermont drained more than 40 m to the Champlain Sea level. The drainage of Lake Vermont sent approximately 1500 ± 225 km^3 to the North Atlantic via the Gulf of St. Lawrence (Rayburn, 2004; Rayburn et al. 2005). The marine incursion probably began about 13,100–12,900 calendar years BP (Richard and Occhietti, 2005; Rayburn, et al., 2007b, in prep.)

A high-resolution record of the history of the Champlain Sea is being developed from analysis of samples collected at centimeter-scale intervals through the upper lacustrine rhythmite and marine mud facies (Fig. 8) in cores from the northern Champlain Lowland. Thinly laminated rhythmites, similar to those in the upper lacustrine unit, occur throughout the lower marine mud facies. An absolute temporal record of early Champlain Sea environments can be reconstructed from these rhythmites, if the couplets represent annual deposition. The interpretations presented below will become clearer as we examine the sediment record in other cores and the sediment record is further calibrated to the emerging radiocarbon chronology.

The initial marine incursion is represented by the disappearance of the freshwater ostracode *Candona* and appearance of diverse marine microfauna (Fig. 8). Marine conditions, however, were short-lived. Microfaunal evidence (Fig. 17) indicates that freshwater conditions were temporarily reestablished before a later and more persistent marine phase (Marine Phase II, Fig. 8) became established in the basin. Early marine rhythmites at PAFB contain 18 sediment couplets (Rayburn et al., in prep.), which indicates that near full-marine conditions may have existed for nearly two decades before the return of freshwater conditions. Rayburn et al. (in prep) estimate the durations of the Champlain Sea freshwater and transitional phases range as 100–140 yr and 260-270 yr, respectively. Thus, more than 380 years may have passed from the initial drainage of Lake Vermont before near full-marine conditions persisted in the basin.

The freshening of the Champlain Sea after the early marine phase requires a large-volume, long-duration source of fresh water. Possible sources include additional freshwater influx from proglacial lakes and ice melting in the upper Great Lakes basin, readvance of the ice margin in the lower St. Lawrence Lowland that reestablished a proglacial lake in the upper St. Lawrence and Champlain lowlands, and large freshwater discharges from large lakes in western or northern Canada. The upper Great Lakes are an unlikely source for large volumes of additional freshwater because most of the Great Lakes drainage was already routed through the Champlain Lowland (Fort Ann Stage of Lake Vermont) by the time of the initial marine incursion. The readvance scenario is also unlikely since recent studies in southeastern Quebec have found no evidence for a readvance (e.g. Parent and Occhietti, 1988, 1999; Occhietti et al., 2001) and there is no evidence for rising water level in the Champlain Lowland, which would have

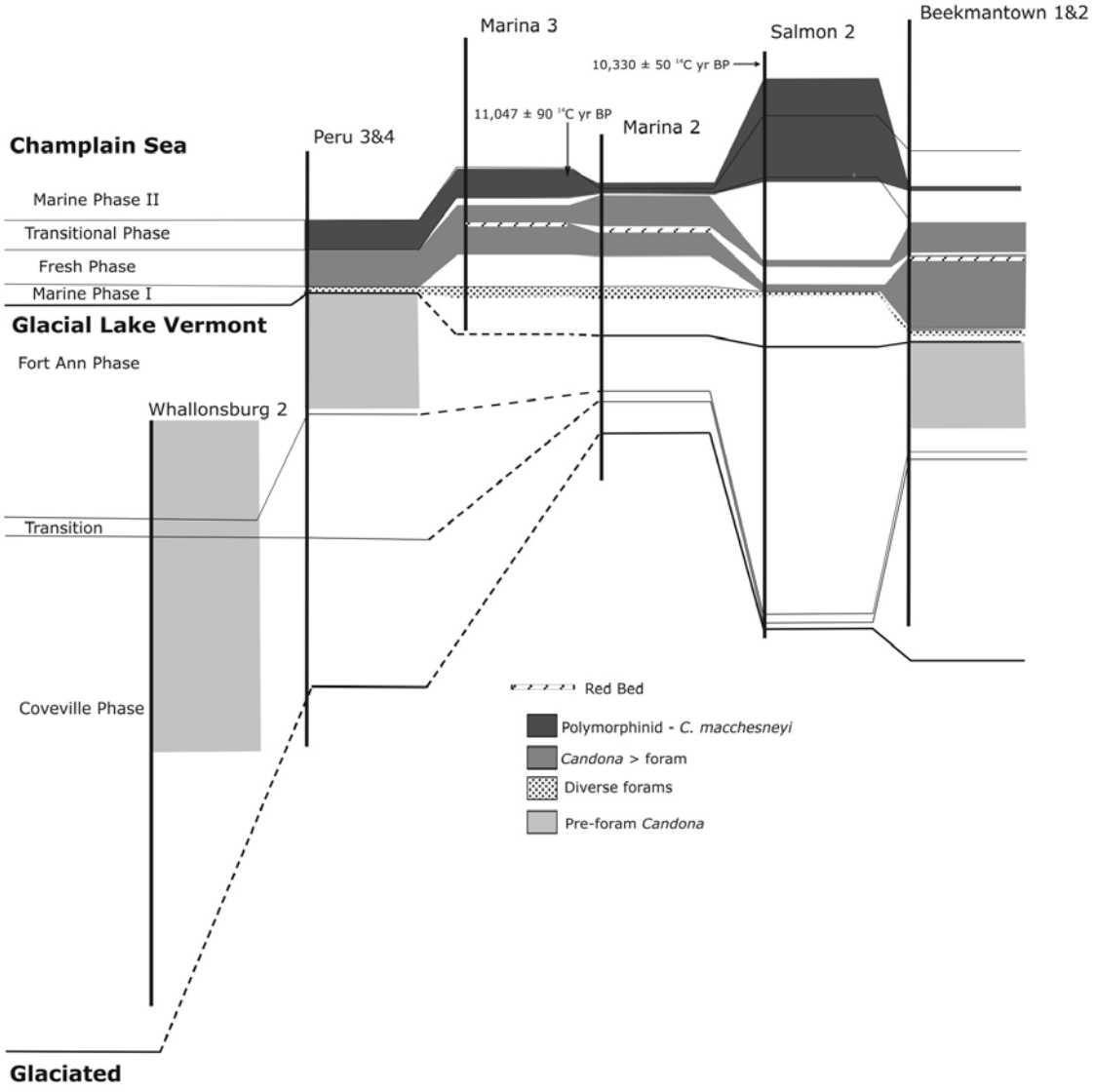

Figure 17. Microfaunal records from the PAFB Marina, Salmon River, Peru, Beekmantown and Whallonsburg cores.

accompanied the reestablishment of a proglacial lake. We consider increased influx from Lake Agassiz to be the most likely source of fresh water during the freshwater and transitional phases of the Champlain Sea (Fig. 8). Discharge of Lake Agassiz through the Great Lakes–St. Lawrence–Champlain lowlands occurred at about the time of the marine incursion (12,900 years BP; Teller 1988) and probably had sufficient magnitude and duration (Teller, 1988; Leverington et al., 2002; Teller and Leverington, 2004) to freshen the Champlain Sea for more than a century.

Post-glacial uplift caused gradual regression and shoaling of the Champlain Sea. The regressive sequence is recorded by the transition from fossiliferous, laminated to thinly

bedded marine mud to sparsely fossiliferous (or barren), thinly bedded medium–fine sand and silt in surface exposures and cores in the northern Champlain Lowland. The upper marine sand facies is unconformably overlain by coarse-grained, trough cross-bedded and horizontally bedded gravel in emergent parts of the Champlain Lowland. These deposits commonly occur in stream terraces and were probably deposited when modern streams regraded to the Lake Champlain base level. The subbottom stratigraphic record from Bulwagga Bay in Lake Champlain (Fig. 7) contains 0.5 m of foraminifera-bearing mud that is overlain by about 4.7 m of lake mud. A date of 7,340 ± 110 ^{14}C years BP (OS-5456, -26.53‰ δ^{13}C) (approximately 7,968–8,368 calendar years BP, CALIB 5.0; Stuiver et al., 2005) from a hemlock cone provides a maximum age for the formation of modern Lake Champlain.

Rates, Dates and Reservoir Effects

Radiocarbon Reservoir Effect

The majority of published dates for late-glacial events in the Champlain Lowland are radiocarbon ages from organic matter or shells from Champlain Sea deposits, primarily because of the increased abundance of datable material following the marine incursion. Champlain Sea dates range from about 13,000 to 9,700 ^{14}C years B.P. (Rodrigues, 1988). An age of about 11,500 ^{14}C years B.P. has been generally accepted for the onset of the Champlain Sea based on a large number of marine invertebrate shell radiocarbon dates (Rodrigues, 1988, 1992; Parent and Occhietti, 1988, 1999), although some non-shell dates suggest it might be younger (Ridge, 2003; Richard and Occhietti, 2005; Rayburn et al., 2007b). Rodrigues (1988) observed stratigraphic inversion of radiocarbon dates on shells at several sites where multiple dates were obtained. He suggested that freshwater and littoral marine species acquired anomalously old carbon from ancient carbonate dissolved in continental meltwater runoff. Marine species are also subject to a reservoir effect because carbon in marine waters is often significantly older than atmospheric carbon. A reservoir correction of about 410 ^{14}C years is generally applied in published Champlain Sea shell dates (e.g. Parent and Occhietti, 1999). However, recent studies of modern marine mollusks indicate that the reservoir effect may exceed 1000 years in Canadian coastal waters (Dyke et al., 2003). Rodrigues (1988) used radiocarbon ages from stenohaline (30-34 ‰ salinity) fossil associations to provide the best age estimates, and placed the marine incursion in the western Champlain Sea basin at 11,400 to 11,000 ^{14}C years B.P.

A study of marine and terrestrial fossils found together (Fig. 18) at several locations in the Champlain Lowland suggests that a simple correction for Champlain Sea marine-reservoir effects is not possible (Rayburn et al., 2006a). Rayburn et al. (2006a) estimated the radiocarbon reservoir effect on shells from three of the most common bivalves in Champlain Sea sediments to determine whether or not reservoir effect was correlated with species or water depth at the time of deposition (Fig. 19). Three *Hiatella arctica* samples from the Salmon-2 core, deposited in an estimated 60 m water depth, have an apparent 500-750 ^{14}C year reservoir effect. Calibration of these ages using the Marine04 (Stuiver et al., 2005) calibration curve accounts for most of the age difference still results

Figure 18. Salmon-2 854-859 cm. Organic layer includes moss, seeds of *caryophylls*, *saxifrages*, blueberries/cranberries, willow buds, & *Dryas* fruits (Norton Miller, New York State Museum, pers. comm., 2006). The shells are *Hiatella arctica*.

in a 200-450 calibrated year reservoir effect. Samples of *Portlandia* and *Macoma*, deposited at an estimated depth of 65-75 m, exhibit significantly larger age reservoir discrepancies. The reservoir effect for these samples is about 1200-2000 ^{14}C years and Marine04 calibration does not improve the effect significantly. The estimated reservoir effect for a combined *Hiatella–Macoma* sample (Wagner, 1972) falls between the species end members (Fig. 19), which implies that reservoir effect is species dependent. Richard and Occhietti (2005) determined that the reservoir effect for a *Macoma* sample from Lake Hertel, PQ, originally deposited in shallow marine water, was about 880 years, which suggests that depth is not a significant factor. An unexpected observation is that the *Macoma* and *Portlandia* samples all appear to have increasing reservoir effects with decreasing age. The larger marine-reservoir effects estimated for *Macoma* and *Portlandia* samples in the Champlain Lowland (Rayburn et al., 2006a) support the observations of Dyke et al. (2003) who reported elevated ^{14}C ages for deposit feeders such as these. The observed increase in reservoir effect with younger age and higher reservoir effect in *Macoma* and *Portlandia* may a major contributor to the age inversions described by Rodrigues (1988).

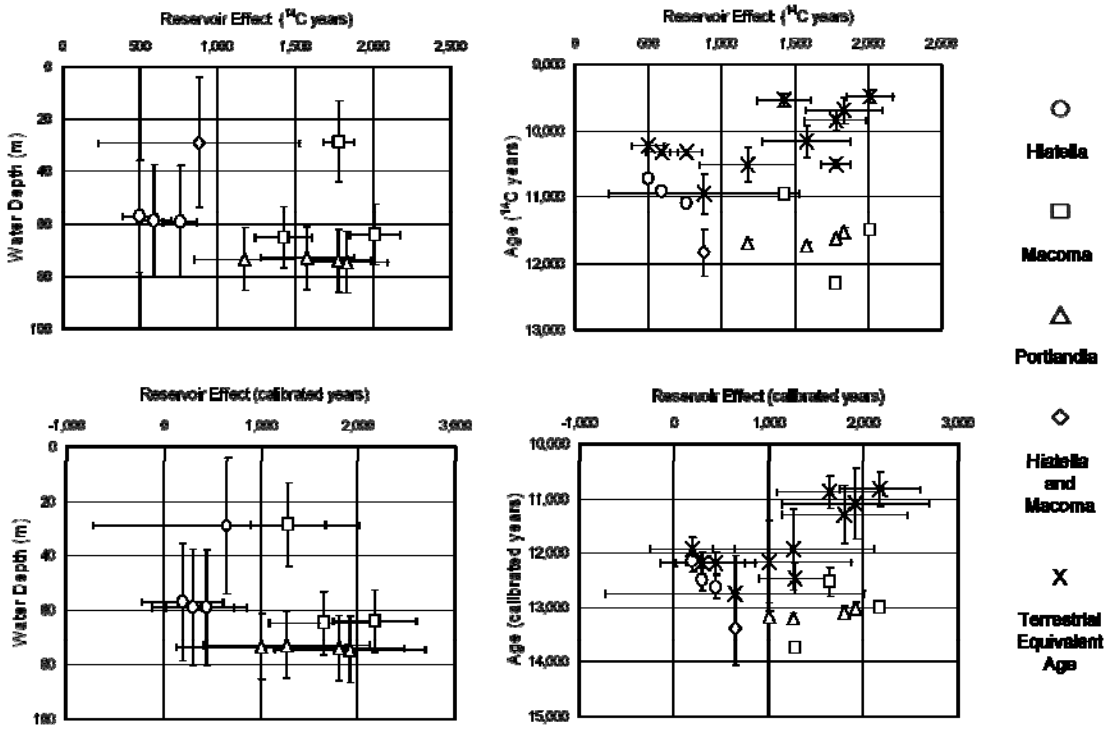

Figure 19. Radiocarbon and calibrated ages of marine shells and equivalent-age terrestrial material. All ages were calibrated using CALIB 5.0.1 (Stuiver et al., 2005) with a 2σ standard deviation. The terrestrial material was calibrated relative to the IntCal04 curve, and the shells were calibrated relative to the Marine04 curve. Water depth was estimated using paleo-elevation derived from isostatic rebound measurements (Rayburn et al., 2005) and core depth.

Despite the fact that *Hiatella* samples from the Salmon River core have a significantly smaller marine-reservoir effect than any of the other samples, there does not appear to be a single species, location or depth-dependent reservoir correction that can be applied to shells from the Champlain Sea. The 410 ^{14}C year reservoir correction (ΔR) assigned to Champlain Sea shells when calibrated with Marine04 would produce adequate correction for the ages of *Hiatella* samples from our sediment cores, but it would not produce significant improvement in age estimates for shells of other species. The often large and unpredictable errors associated with radiocarbon dating of marine shells in the Champlain Sea indicate that shell dates should not be used to measure the timing of events in the basin. The reservoir variability demonstrated among species and locations also suggests that shell ages are a poor choice for estimating sedimentation rates over the time scales in question.

Radiocarbon dates for terrestrial organic material in the Champlain Lowland, although fewer in number, are free of marine-reservoir effects and yield significantly younger ages for late-glacial events in the region. Rayburn et al. (2007b) reported two dates on terrestrial organic materials that predate the Champlain Sea in the Champlain Lowland. A musk-ox vertebra from near Elizabethtown (Figs. 9 and 10) yielded an age

of 11,280 ± 110 ^{14}C years B.P. (AA-4935) (Cadwell and Pair, 1991). The ^{14}C age is corrected to 11,362 ± 115 ^{14}C years B.P. if an estimated -20 ± 2 ‰ δ^{13}C correction is applied, which is equivalent to 13,438–13,020 calibrated years B.P. (Rayburn et al., 2007b). A piece of terrestrial wood from Long Pond in Willsboro, New York (Fig. 7) produced an age of 10,901 ± 76 ^{14}C years B.P. (Wk-10957, δ^{13}C -25.6 ± 0.2 ‰), which is equivalent to 12,995–12,793 calibrated years B.P. The wood at Long Pond occurs near the base of local lake sediments that overlie Lake Vermont (Coveville Stage) rhythmites. Long Pond was established when Lake Vermont dropped from the Coveville to upper Fort Ann level (Fig. 11) as a result of the Lake Iroquois breakout. A similar set of circumstances occurred in Boyd Pond in the northwestern Adirondack Uplands. Anderson (1988) reported a conventional radiocarbon age of 11,200 ± 190 ^{14}C years B.P. (GSC-3429) on bulk organic material recovered immediately above glaciolacustrine (Lake Iroquois) clays. The dated organic material represents the isolation of Boyd Pond from Lake Iroquois, which occurred when Lake Iroquois dropped to the Lake Frontenac level. The drainage of the glacial lakes Iroquois and Vermont occurred before the deposition of dated material in both Long Pond and Boyd Pond, making these dates minimum ages for the Lake Iroquois breakout. We suggest that the younger Long Pond age of 10,901 ± 76 ^{14}C years B.P. on wood, rather than the 11,200 ± 190 ^{14}C years B.P. age on lacustrine organic sediment in Boyd Pond, is a better minimum-age estimate because of the greater certainty of the dated material.

Dineen and Hanson (1992) published two radiocarbon ages from terrestrial organic material in Lake Albany deposits. The dates are 11,700 ± 115 ^{14}C years B.P., reported only as an AMS age on a spruce needle in moss, and 11,050 ± 450 ^{14}C years B.P. (GX-1434) from a bulk sample of laminated silty sand with moss. The samples were collected at the contact between glaciolacustrine varved clay and deltaic sand near Schuylerville, which they interpreted to be the transition between Lake Albany and Lake Quaker Springs (Dineen and Hanson, 1992). The deposits from which these samples were collected record a sudden drop in the level of Lake Albany and may correlate with the events dated at Long and Boyd ponds if proglacial lake drainage in the upper Hudson Lowland was caused by the Iroquois breakout as previously discussed.

Richard and Occhietti (2005) used terrestrial plant macrofossils recovered near Mount St. Hilaire, Quebec and an estimated ice retreat rate of 250 m/yr to estimate the age of the initial marine incursion in the St. Lawrence and Champlain lowlands to be 11,100 ± 100 ^{14}C years B.P. (13,187–12,872 calibrated years B.P.). A new radiocarbon date of 11,047 ± 90 ^{14}C years B.P (WW-5383, δ^{13}C -25.7) (13,124–12,853 calibrated years B.P.) obtained from plant fragments recovered in transitional marine sediments of the PAFB Marina-3 core is in close agreement with the estimate of Richard and Occhietti (2005) for the beginning of the Champlain Sea.

Other Dating Techniques

Antevs (1922, 1928) constructed the New England Varve Chronology (NEVC) through correlation of varve-thickness records throughout New England and the Lake Albany basin in southeastern New York. Recent attempts have been made to extend the NEVC spatially and temporally through the inclusion of varve records from the Champlain Lowland (Ridge et al., 1999; Connally and Cadwell, 2002; Ridge, 2003). The

NEVC is calibrated to an absolute time chronology by several radiocarbon ages in New England (Ridge et al., 1999; Ridge, 2003). Thus, correlation of varve sequences in the Champlain Lowland to the NEVC would provide an independent means to date the deglacial events. Our preliminary attempts to correlate rhythmites (varves) from Lake Vermont to the NEVC have met with mixed results. We have not been able to correlate Coveville Stage varves from the Long Pond and Hadley pond cores in Willsboro (Fig. 7; Rayburn, 2004) or from the Keeseville Industrial Park outcrop (Field Trip Stop 12) (Rayburn, 2004; Ridge, pers. com., 2007) to the NEVC. Ridge et al. (1999) used paleomagnetic secular variations to correlate part of the Whallonsburg (Field Trip Stop 13) sequence to the NEVC at approximately 13,600–13,200 calibrated years B.P. (Ridge, 2003; pers. com.). The varve thickness record for 247 varves immediately below the Lake Iroquois breakout event varves (Fig. 16) in the Whallonsburg core may correlate to the NEVC at approximately 13,400–13,200 calibrated years B.P. Although this result is compatible with the ages obtained by Ridge (2003) for the same deposits, the estimate should be considered preliminary.

The paleomagnetic inclination and declination records from a portion of the PAFB Marina core through the lacustrine–marine transition (Rayburn et al., 2006b) are in general age agreement with the secular paleomagnetic curves developed by Ridge et al. (1999) and Ridge (2003) (Fig. 20). The radiocarbon age of 11,052 ± 90 BP in the transitional marine zone of the core correlates well with the paleomagnetic age model of Ridge et al. (1999) and Ridge (2003). Future work on the paleomagnetic record from the Whallonsburg core may produce a better correlation because some of the data used by Ridge et al. (1999) are from outcrop at the Whallonsburg site.

Rayburn et al. (2007a) reported seven ^{10}Be cosmogenic exposure age dates for the scoured bedrock at Altona Flat Rock (Field Trip Stop 3) and six boulders from Cobblestone Hill (Field Trip Stop 5). The ages are closely clustered (Fig. 21) suggesting that exposure dating is a valid technique for dating the flood event from these land-forms. By using the production rate of 4.7 atoms/year suggested by Balco and Schaefer (2006) and adjusting for isostatic rebound the ages averaged 11,900 ± 700 calibrated years B.P., which is about 1300 years younger than the suggested radiocarbon age of the event (Fig. 21).

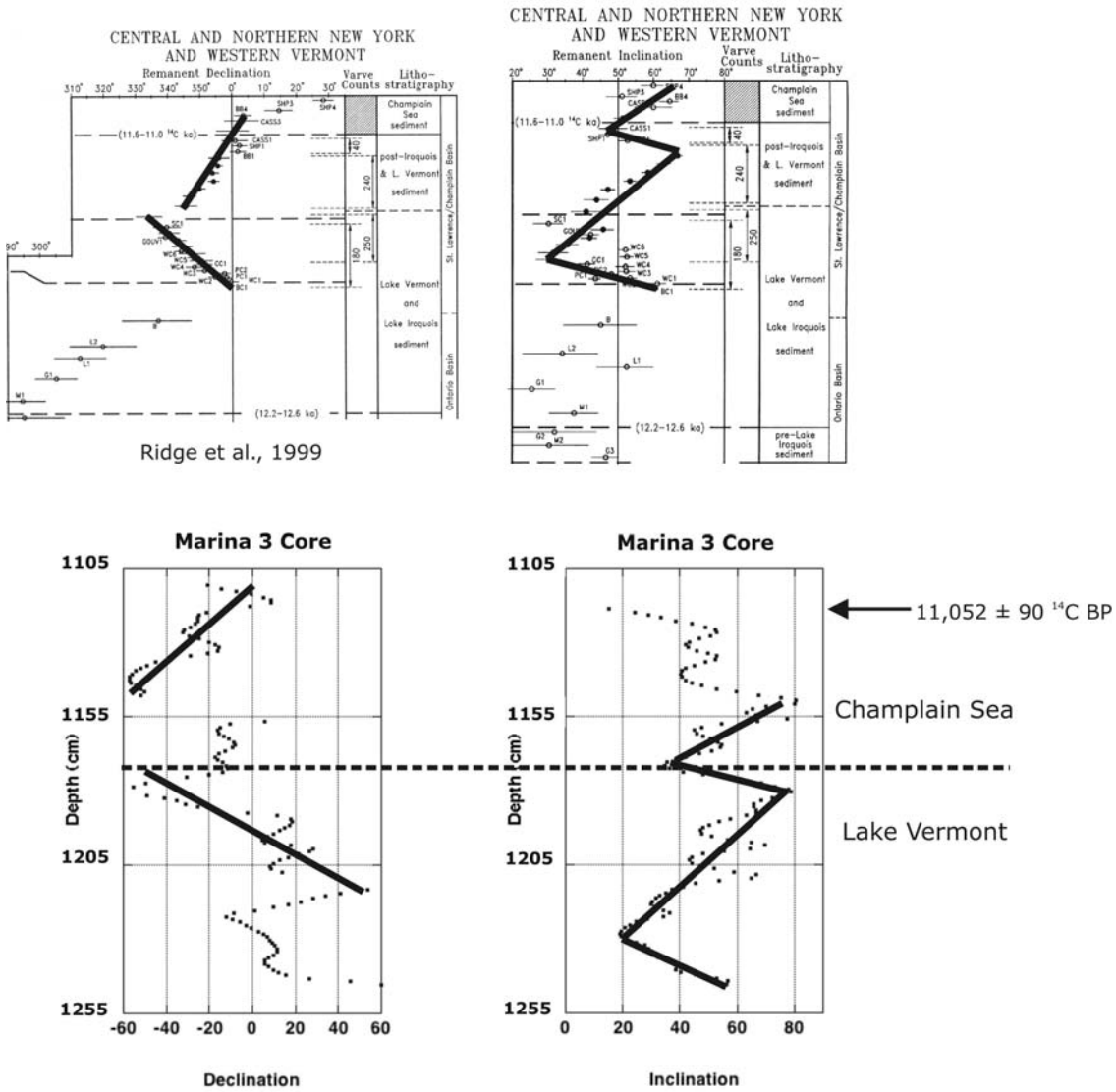

Figure 20. Paleomagnetic inclination and declination in the Marina-3 core (Plattsburgh) compared with data from Ridge et al. (1999) and Ridge (2003). A radiocarbon age of 11,052 ± 50 B.P. was measured near the top of this sequence in the marine transitional phase.

Rates of Ice Recession

Ice recession in the Champlain Lowland between the Elizabethtown and Cobblestone Hill ice margins (Figs. 10 and 12) is bracketed by dates of 13,438–13,020 calibrated years B.P. from the Elizabethtown musk-ox bone and 12,995–12,793 calibrated years B.P. from terrestrial wood at Long Pond (Fig. 22). The wood from Long Pond was collected a short distance above the contact with Coveville varves so should be considered a minimum age for the Iroquois breakout. The distance between the Elizabethtown and Cobblestone Hill ice margins is approximately 65 km, which produces a minimum average retreat rate of about 0.19 km yr^{-1}.

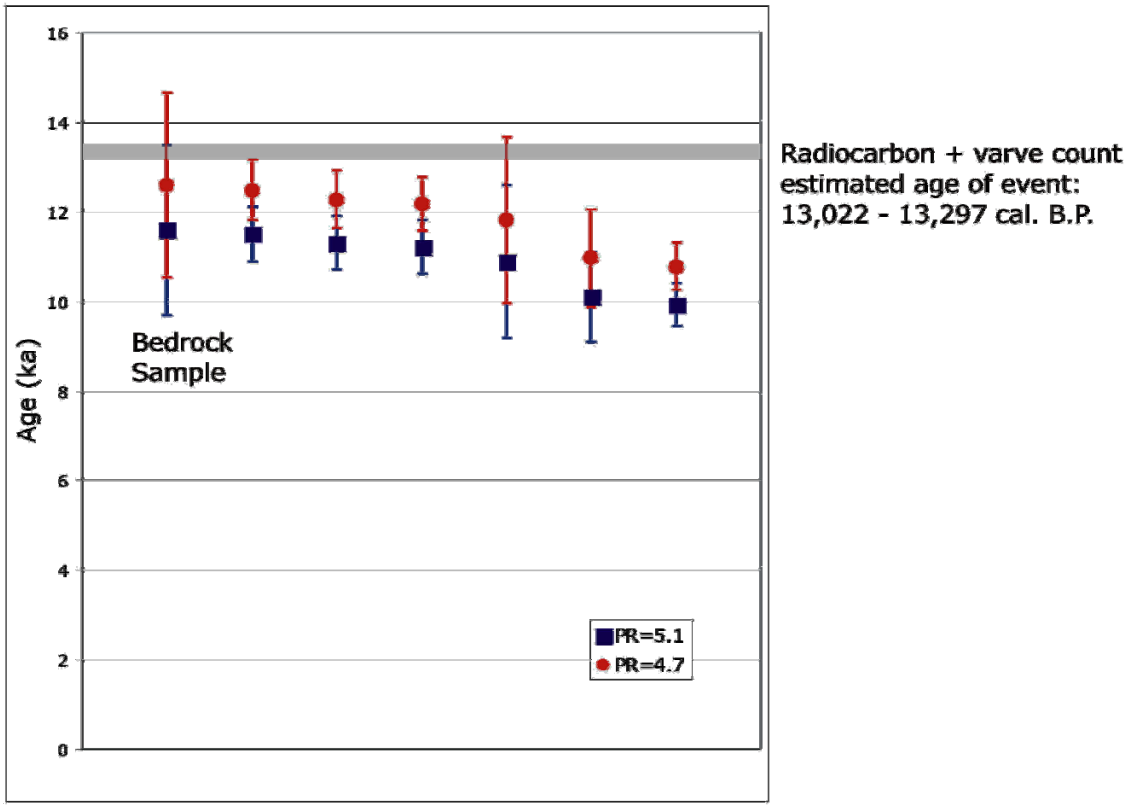

Figure 21. ^{10}Be ages of six boulders on top of Cobblestone Hill and one bedrock sample from the scoured Altona Flat Rock. Ages were adjusted for isostatic rebound (adding about 500 years to the measured age) and are shown for production rates of both 5.1 (Stone, 2000) and 4.7 (Balco and Schaefer, 2006) atoms/year. These ages are well grouped, but consistently younger then the estimated age for the Lake Iroquois break out of about 13,200 B.P.

The rate of ice recession from Keeseville to Cobblestone Hill may be estimated from the varve record at the Keeseville Industrial Park (KIP; Field Trip Stop 12). The KIP site contains of 68 varves deposited in the Coveville Stage of Lake Vermont that are overlain by about 13 m deltaic sand deposited in Lake Vermont at the Fort Ann Stage. The drop from the Coveville to Fort Ann level occurred during the Lake Iroquois breakout when the ice stood at the Cobblestone Hill ice margin. The distance between Keeseville and Cobblestone Hill is approximately 30 km, which represents a retreat rate of approximately 0.44 km/yr.

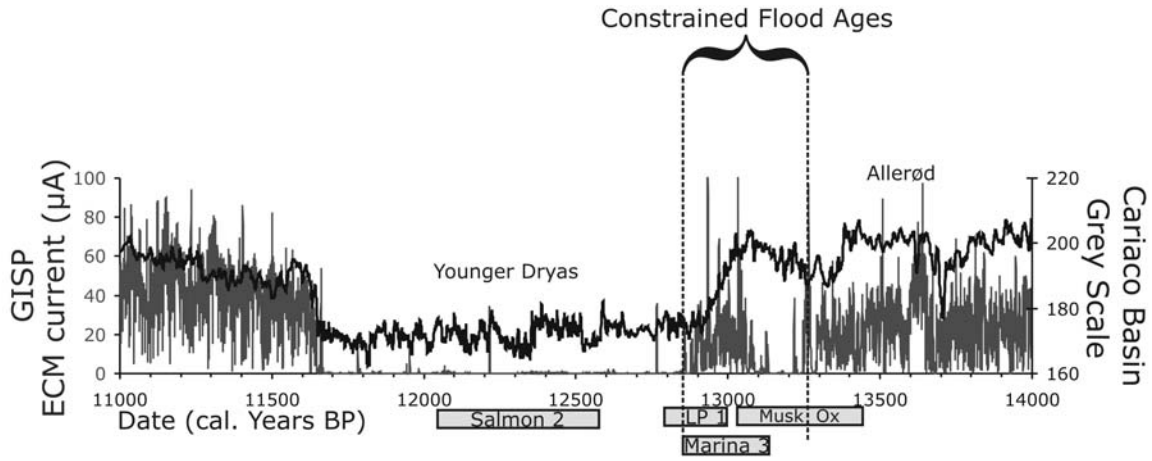

Figure 22. Calibrated ages and paleo-climate indicators. The musk-ox age (13,020–13,438 B.P.) represents the ice margin at Elizabethtown/Westport. The Long Pond (LP1) age (12,793–12,995 B.P.) represents the minimum age of the ice margin at Cobblestone Hill and the break out of Lake Iroquois. The Marina 3 age (12,854–13,129 B.P.) represents the waning of the Champlain Sea freshening event, and is therefore a minimum age for the beginning of the Champlain Sea. The Salmon 2 age (12,054–12,578 B.P.) is well into Marine Phase II of the Champlain Sea (Fig. 8). The constrained age bracket for all four floods is calculated using these radiocarbon ages and varve counts (see text). The GISP electro-conductivity data are after Taylor et al. (1993) and the Cariaco basin grey scale data is from Hughen et al. (2000).

Varve records from the (KIP) site (Field Trip Stop 12) and the Peru and Whallonsburg cores indicate that the Champlain Sea began at least 284 years after the ice receded from the Keeseville–Peru area. The record at KIP contains 68 varves that record ice recession from the Keeseville–Peru area to the Cobblestone Hill ice margin near Altona. The Whallonsburg core contains 48 debris-rich rhythmites (varves) that we interpret to represent the passing of the Iroquois and Frontenac breakout events, during which time the ice receded to a position near the international border. The Peru core contains 168 varves that lie conformably over breakout-flood sand deposits and thus record the interval between the Frontenac breakout and the formation of the Champlain Sea. If a constant ice retreat rate through the northern Champlain Valley and St. Lawrence Lowland is assumed, then a calibrated age of 13,020–13,438 years B.P. from the musk-ox bone at Elizabethtown about 30 km south of Peru, suggests that the beginning of the Champlain Sea occurred between 12,715–13,135 calibrated years B.P. (Fig. 22). This estimate is in good agreement with the calibrated age of 12,854–13,129 years B.P. from the Marina 3 core (Fig. 22).

Franzi et al. (2002) noted that the projection of the Cobblestone Hill ice margin into the Champlain Lowland lies close to the southern limit of the Ingraham Esker (Field Trip Stop 6) near Beekmantown (ice margin 10 of Denny, 1974; Fig. 3 of Franzi et al., 2002). They proposed that esker formation might have been initiated by steepening of hydraulic

gradients within the ice sheet caused by the 24-m drop from the Coveville to Upper Fort Ann level of Lake Vermont that accompanied the Lake Iroquois breakout. Diemer (1988) determined that the esker formed as a time-transgressive series of overlapping subaqueous fans during ice recession. The esker extends approximately 20 km northward to a point about 8.2 km south of the international border in Champlain. The northern end of the esker lies near the eastward projection of the ice margin that stood on the north flank of Covey Hill (ice margin 15 of Denny, 1974) at the time when drainage of Lake Frontenac began. If these interpretations are correct, deposition of the Ingraham Esker correlates in time with the duration of Lake Frontenac in the St. Lawrence Lowland, which from the varve record at Whallonsburg, may have been 48 years. The length of the esker, 20 km (Fisher, 1968; Denny, 1970; 1972,) provides a maximum estimate for the distance for ice recession and yields a maximum retreat rate of 0.42 km yr^{-1} in the northern Champlain Lowland. If half the esker length (10 km) was deposited in the subglacial tunnel at the time esker sedimentation ended, the retreat rate estimate is reduced to 0.21 km yr^{-1}. Although based largely upon indirect evidence, these are reasonable upper and lower estimates for ice recession rates that are similar to rates of retreat determined from other evidence.

Paleoclimatic Implications

Decreases in North Atlantic salinity (Dickson et al., 2002; Antonov et al., 2002; Curry and Mauritzen, 2005) and slowing meridional overturning circulation (MOC) (Häkkinen and Rhines, 2004; Bryden et al., 2005) have heightened concerns about global climate change in recent years. Climate models indicate that MOC is sensitive to small volumes of fresh-water input but oceanic and climate response is highly dependent upon the location and duration of the freshwater forcing (Manabe and Stouffer, 1997; Ganopolski and Rahmstorf, 2001, 2002; Stouffer et al. 2006). Catastrophic glacial lake discharges during the late Wisconsin and early Holocene offer opportunities to examine the role of freshwater forcing of ocean circulation and climate. Broecker et al. (1989) (see also Teller 2002), proposed that the most studied late Wisconsin climate event, the Younger Dryas reversal 12.9-11.5 ka, was caused by the breakout of glacial Lake Agassiz to the North Atlantic through the St. Lawrence Lowland. This hypothesis has undergone close scrutiny since its conception because definitive evidence for the age, routing, and duration of Lake Agassiz and other glacial lake discharges remains elusive (Teller et al., 2005, Tarasov and Peltier, 2005; Broecker, 2006).

Atmosphere-ocean general circulation model (AOGCM) experiments by Rahmstorf (1995, 2000) predict that moderate changes in the flux of freshwater input into the North Atlantic, perhaps less than 0.06 Sverdrup (1 Sv = 10^6 m^3 s^{-1}) sustained over centuries, can lead to disequilibria in North Atlantic Deep Water (NADW) circulation, producing substantial changes in regional climate, such as the Younger Dryas event (Broecker et al., 1989). Rayburn et al. (2005) recognized three large flood pulses to the North Atlantic, each with a magnitude around 0.1 Sv, during deglaciation of the Champlain and St. Lawrence lowlands. The first flood was the combined event that resulted from the breakout of Lake Iroquois across the St. Lawrence–Champlain drainage divide near Covey Hill and the contemporaneous breakout of Lake Vermont from the Coveville to the upper Fort Ann level. This event released around 700 km^3 of freshwater into the

North Atlantic via the Hudson River over an estimated 2-3 months (Rayburn et al., 2005). The second pulse (the Lake Frontenac drainage event) released an estimated 2,500 km^3 of freshwater through the Hudson Lowland.

The third event (final drainage of Lake Vermont and formation of the Champlain Sea) released an estimated 1,500 km^3 of water to the North Atlantic through the Gulf of St. Lawrence. The Hudson–Champlain lowland flood route was never reactivated. Ambient discharge of freshwater from the Great lakes region to the Gulf of St. Lawrence is estimated to have been about 0.06 Sv (Rayburn et al., 2005). Rayburn et al. (2007b) suggest that the Lake Vermont–Champlain Sea flood occurred close to the onset of the Younger Drays (Fig. 22). New evidence presented here indicates that the third flood, which is associated with the Champlain Sea incursion, occurred before 13,000 calibrated years B.P., a finding compatible with Richard and Occhietti's (2005) estimate.

We also present new evidence for a fourth flood event, possibly the largest during the deglaciation of the region, that originated somewhere in the Great Lakes or Lake Agassiz basins. This flood occurred shortly after the initial marine incursion and is responsible for a prolonged return to freshwater conditions (Freshwater Phase of the Champlain Sea; Figs. 8 and 17) in the St. Lawrence–Champlain lowlands. The magnitude of this flood is not known but it was large enough to have maintained freshwater conditions in the Champlain Lowland for more than 100 years. Our preliminary evidence indicates that the flood was waning at the beginning of the Younger Dryas.

Figure 22 illustrates the timeline for these floods in comparison to climate change signals from the GISP core and the Cariaco basin. Our preliminary analyses of varve records preserved in cores from the Champlain Lowland indicate that all of these floods probably occurred within about 350 years. The absolute timing of the floods can be constrained to an approximately 400 year interval preceding the Younger Dryas. The maximum age for the Lake Iroquois breakout flood (about 13,258 calibrated years B.P.) may be determined by subtracting 180 varve years (the minimum number of varves in the Whallonsburg 2 core before the Lake Iroquois flood) from the maximum age of the Elizabethtown musk-ox bone (approximately 13,438 calibrated years B.P.). The minimum age for the flood responsible for the Champlain Sea Freshwater Phase (approximately 12,854 calibrated years B.P.) is determined from the minimum age of a wood sample collected from Transitional Phase deposits (Fig. 17) in the PAFB Marina core. The late glacial floods in this region, therefore, predate the Younger Dryas signal in the GISP core but span the time that marks the end of the Allerød in the Cariaco basin (Fig. 22). The Cariaco Basin grey scale is a direct signal of North Atlantic ocean circulation (Hughen et al., 2000), and thus should be a more sensitive indicator of changing ocean currents. It is possible that the decrease in grey scale in the Cariaco basin from around 13,100–12,900 years B.P. is a result of a combined effect of the four St. Lawrence–Champlain floods in addition to the change in outflow location from the mid-Atlantic region (via the Hudson–Champlain lowlands) to the North Atlantic (via the Gulf of St. Lawrence). These changes may have slowed NADW circulation until a threshold was reached around 12,850 when the GISP record indicates a rapid change in climate at the beginning of the Younger Dryas.

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Road Log and Stop Descriptions

Day 1– The Gulf at Covey Hill (optional)

Day 2–Northwestern Champlain Lowland and Northeastern Adirondack Upland

- Stop 1–Chateaugay Channels
- Stop 2–Ellenburg Moraine
- Stop 3–Altona Flat Rock at Rock Road
- Stop 4–Jericho Moraine
- Stop 5–Altona Flat Rock and Cobblestone Hill
- Stop 6–Ingraham Esker
- Stop 7– Patterson Brook Till Exposure
- Stop 8– Saranac River Till Exposure at Treadwell Mills

Day 3–West–Central Champlain Lowland and East–Central Adirondack Upland

- Stop 9–Plattsburgh Air Force Base Marina
- Stop 10–Ausable Chasm Overlook
- Stop 11–Bedded Diamicton Facies at Mud Brook
- Stop 12–Keeseville Industrial Park
- Stop 13–Whallonsburg Landslide
- Stop 14–Oak Hill Wollastonite Quarry

Road Log for Saturday 2 June 2007

Incr. Distance (mi)	Cum. Distance (mi)	Description
0	0	Leave north parking lot at Hudson Hall and turn right on Broad St. Proceed west on Broad St. for 0.25 mi through the traffic light at Draper Ave. to the Prospect St. intersection.
0.3	0.3	Turn right onto Prospect St. Proceed north through the traffic light at Cornelia St. (0.09 mi) to the intersection with Tom Miller Road (distance to next turn is 0.72 mi).
0.7	1.0	Intersection of Prospect St. and Tom Miller Road. Turn left onto Tom Miller Road and proceed west across the Northway (I-87) overpass for 2.6 mi to the Military Turnpike intersection.
2.5	3.5	Military Turnpike intersection. Turn right and proceed northwest to the intersection with US Route 11 in Ellenburg (total distance = 25.0 mi).
1.0	4.5	Military Turnpike crosses Rte. 374 (Cadyville Expressway)
5.8	10.3	Continue northwest on Military Turnpike past the Recore Road intersection. The road rises onto Lake Vermont strandline deposits and up the western Champlain Lowland border fault-line scarp.
7.8	18.1	Military Turnpike crosses the Great Chazy River. The entrance to Ganienkeh Territory (a native-American community) is just ahead on the right.
5.3	23.4	Military Turnpike crosses the southern margin of the Ellenburg Moraine just before the intersection with Plank Road. Continue northwest on Military Turnpike.
2.7	26.1	Intersection with US Rte. 11. Turn left and proceed west on US Rte. 11 to Châteaugay (total distance = 13.0 mi).
9.6	35.7	Enter Franklin County on US 11. Continue west to Châteaugay.
3.4	39.1	Intersection of Rte. 374 in Châteaugay. Turn left and proceed south to the Pulp Mill Road intersection.
1.5	40.6	Pulp Mill Road intersection. Turn right and proceed west to the steel bridge that spans the Châteaugay River.

Incr. Distance (mi)	Cum. Distance (mi)	Description
0.8	41.4	**STOP 1. Châteaugay Chasm and Channels.** Park on the left shoulder of the road and proceed across the bridge on foot.

The chasm represents post-glacial incision by the Châteaugay River through the surficial deposits and into the underlying Cambrian Potsdam Sandstone. The Châteaugay River descends from 290 m to 170 m above sea level along an 8-km reach through the chasm (Williams et al., in review). High Falls Park is located about 400 meters downstream from the bridge. We will meet in a small clearing on the northwest side of the chasm. We will travel in caravan to observe some of the largest Châteaugay Channels after a brief introduction at this location.

Return to the vehicles and continue west on Pulp Mill Road to the intersection with River Road (total distance = 0.13 mi).

0.1	41.5	River Road (Healy Road) intersection. Turn left and follow River Road south to Hartnett Road. The road crosses the eastern ends of three channels at distances of 0.13 mi, 0.49 mi and 0.83 mi.
0.8	42.3	Hartnett Road intersection. Turn the vehicles around at this location and proceed to Stop 2. Ellenburg Moraine.
0.8	43.1	Pulp Mill Road intersection. Turn right and follow Pulp Mill Road east across the Châteaugay River to Rte. 374.
0.9	44.0	Rte 374 intersection. Turn left and proceed north to US Rte. 11 in Châteaugay.
1.5	45.5	US Rte. 11 intersection. Turn right and follow US Rte. 11 east to Ellenburg Depot (total distance = 15.1 mi).
3.4	48.9	Enter Clinton County on US Rte. 11. Continue east to Ellenburg Depot.
9.5	58.4	Intersection of US Rte. 11 and Military Turnpike in Ellenburg. Continue east on US Rte. 11.
2.6	61.0	Intersection of US Rte. 11 and Plank Road in Ellenburg Depot. Turn right and proceed south on Plank Road.

Incr. Distance (mi)	Cum. Distance (mi)	Description
1.0	62.0	**STOP 2. Ellenburg Moraine,** Turn right into gravel pit entrance.

The gravel pit at this stop is excavated into the eastern (ice-proximal) side of the moraine. The pit contains approximately 10 to 12 meters of interbedded sand, gravel and diamicton that overlies Potsdam Sandstone. Bed thickness generally ranges from about a decimeter to just over a meter but a prominent diamicton layer at the northern end of the exposure exceeds 2 meters in thickness. The maximum elevation of the upper surface of the moraine at this location ranges between 290 and 297 meters above sea level.

An exposure on the west (ice-distal) side of the moraine near the outlet of Lake Roxanne, previously described by Franzi et al. (1994), is no longer well exposed and will not be visited. The Lake Roxanne exposure contains approximately 10 to 12 meters of interbedded fine to medium sand with minor gravel and silt interbeds. Bedsets range from a centimeter to a few decimeters thick and are commonly horizontally laminated or ripple-cross laminated. Thin silt or silty fine sand deposits occur locally as draped laminae. Ripple azimuths and the gentle dip of the strata indicate a westerly paleocurrent. The Ellenburg Moraine was probably deposited in a proglacial lake west of the moraine in the upper North Branch Valley. A small sandplain at an elevation of about 290 meters above sea level at Ellenburg may represent a delta that was built by the North Branch into the western end of the proglacial lake.

Leave the gravel pit and turn left onto Plank Road. Proceed south to Military Turnpike.

0.6	62.6	Military Turnpike intersection. Turn left and follow Military Turnpike southeast to Devils Den Road in Altona.
6.0	68.6	Devils Den Road intersection. Turn left and proceed northeast toward Altona.
1.5	70.1	Intersection of Devils Den Road and Rock Road. Turn right onto Rock Road. The road surface changes from gravel to the exposed rock pavement of Altona Flat Rock a few hundred meters from the intersection. The jack pine barren on this part of the sandstone pavement was damaged extensively during the January 1998 ice storm. Continue about 0.5 mi on Rock Road to Stop 3.

Incr. Distance (mi)	Cum. Distance (mi)	Description
0.5	70.6	**STOP 3. Altona Flat Rock at Rock Road.**

The sandstone at this location consists of medium- to coarse-grained quartz arenite and sub-feldspathic sandstone that probably represents a transitional facies between the AuSable and Keeseville Members of the Potsdam Formation (Landing et al., in press). Glacially polished surfaces, parabolic gouges and friction cracks can be observed on unweathered surfaces. The approximate exposure age of the surface from ^{10}Be dating of sandstone samples from this surface is 11,590 ± 190 yrs. B.P. (Rayburn et al., 2007a).

Return to the vehicles and continue south on Rock Road to Military Turnpike.

1.0	71.6	Military Turnpike intersection. Turn left and proceed southwest for 0.48 mi on Military Turnpike to Duley Road.
0.5	72.1	Duley Road intersection. Turn right onto Duley Road and proceed west to Rand Hill Road.
2.1	74.2	Rand Hill Road intersection. Turn left and follow Rand Hill Road to Jericho.
3.7	77.9	Village of Jericho. Continue south on Rand Hill Road. Stop 3 will be on your right 1.3 mi ahead.
1.3	79.2	**STOP 4. Cadyville–Jericho Moraine.** Turn into the gravel pit entrance on the right side of the road.

The Jericho–Cadyville Moraine is a complex of till and gravel ridges and meltwater channels that were built in the Saranac River Valley (Denny, 1974). Crudely bedded, sandy, red-brown to maroon and light gray diamictons are exposed in this pit. The red color of some of the diamictons is probably derived from a red shale, siltstone and dolostone unit that underlies the Ausable Member of the Potsdam Sandstone. Fisher (1968) referred to the unit as the "basal member" of the Potsdam Formation. Landing et al. (in press) informally propose the name "Altona Formation" for the unit based upon new exposures and well logs in the Little Chazy River watershed near the southeastern margin of Altona Flat Rock.

Return to the vehicles, turn right at the pit entrance and continue south on Rand Hill Road to Jersey Swamp Road.

Incr. Distance (mi)	Cum. Distance (mi)	Description
2.6	81.8	Jersey Swamp Road intersection. Turn left on Jersey Swamp Road. Proceed east on Jersey Swamp Road to Military Turnpike. The road descends the western Champlain Lowland border fault-line scarp over the next 1.5 mi. Gravel pits on the left side of the road at 1.5–1.8 mi expose ice-contact gravel and sand deposited into the Coveville level of Lake Vermont. The road crosses strandline deposits of the Fort Ann level of Lake Vermont at about 1.8–2.3 mi. Continue east on Jersey Swamp Road.
2.9	84.7	Military Turnpike intersection. Turn left and follow Military Turnpike northwest for 3.2 mi to Recore Road.
3.2	87.9	Recore Road intersection. Turn right onto Recore Road and proceed to the Recore Road–West Church Street–Atwood Road intersection.
2.1	90.0	Recore Road–West Church Street–Atwood Road intersection. Follow West Church Street east for 0.1 mi and turn left onto Barnaby Road. Follow Barnaby Road north about 1.0 mi to the end of the paved surface. The road continues north as Blaine Road.
1.0	91.0	Barnaby Road ends. Continue north on Blaine Road (a seasonal gravel road).
1.1	92.1	Continue through the gate at the entrance to property owned by the William H. Miner Agricultural Research Institute. The road rises onto the northeastern flank of Cobblestone Hill. Cobble-gravel beach ridges deposited in the Fort Ann level of Lake Vermont are exposed along the road about 150 meters from the gate.
0.1	92.2	**STOP 5a. Cobblestone Hill and upper Fort Ann Stage Beach Ridges**

The beaches at this location were first described by Woodworth (1905a) and later by Denny (1974). The deposits consist predominantly of moderately rounded to well rounded cobble gravel in multiple, low relief ridges or terraces that extend along the northern and eastern flanks of Cobblestone Hill at elevations between 206 and 175 meters above sea level. The highest ridges lie near the projected highest shoreline of the Upper Lake Fort Ann. Individual ridges are typically 1 to 2 meters high and 10 to 20 meters wide, and often extend laterally for more than 400 meters (Denny, 1974). The gravel is almost exclusively composed of Potsdam Sandstone that was derived from the alluvial boulder gravel that composes Cobblestone Hill.

The large (0.2 to 1.4 meter diameter), angular boulders that comprise the core of Cobblestone Hill can be seen along the road a short distance above the highest beach ridge. The boulders of Cobblestone Hill represent material washed into Lake Vermont from the sandstone pavements by ice-marginal streams from the breakout of glacial Lake Iroquois (Woodworth, 1905a; Denny, 1974; Clark and Karrow, 1984; Pair et al., 1988). Reworking of these alluvial deposits by wave action with relatively little long-shore transport probably formed the beach deposits (Denny, 1974).

Incr. Distance (mi)	Cum. Distance (mi)	Description
0.1	92.3	Turn right at the fork in the road.
0.1	92.4	Road crosses the top of Miner Dam.

Continue northwest on the “Scarpit” Road. N.B.: The “Scarpit” Road presents many hazards, especially for those driving it for the first time. Please drive slowly and cautiously.

The road lies on the southwest flank of Cobblestone Hill following the abandoned shoreline of the former reservoir behind Miner Dam. Miner Dam was part of a failed hydroelectric project initiated by William Miner in 1910 (Gooley, 1980). By the time of its completion in March, 1913, the concrete dam, known locally as the "Million-Dollar Dam", had a maximum height of over 10 meters and stretched more than 700 meters across the Little Chazy River Valley. The design capacity of the reservoir was more than 3.5 million cubic meters.

The inadequate flow of the Little Chazy River and groundwater seepage through Cobblestone Hill, which formed the eastern flank of the reservoir, proved to be major design flaws for the project. A 10 to 15 cm layer of concrete grout was spread over more than 100,000 m^2 along the flank of Cobblestone Hill (the Scarpit) to mitigate the seepage loss. A deep trench was excavated at the base of Cobblestone Hill behind the dam for the purpose of pouring a grout curtain to the underlying sandstone and thereby, presumably, sealing the northeastern flank of the reservoir. The dam and generating station were completed in 1913 but it took almost two years to fill the reservoir to capacity. The grouting effort was partially successful and the power generating plant began operation on January 21, 1915, more than four years from the beginning of the project (Gooley, 1980). The power plant produced electricity intermittently for seven years before mechanical problems forced the abandonment of the project.

Construction of a second dam, the Skeleton Dam (Gooley, 1980), approximately 1.5 km upstream was begun in 1920 to provide supplemental flow to the main impoundment. The Skeleton Dam project, however, ended with the failure of the Miner Dam generating station and was never completed.

Incr. Distance (mi)	Cum. Distance (mi)	Description
0.6	92.9	**STOP 5b. Cobblestone Hill Ice-Marginal Deposits.**

The Cobblestone Hill boulder deposits occur at two distinct elevations at this location. The upper level lies between 225 and 232 meters above sea level and may correspond to cobble and boulder deposits at a similar elevation near Bear Hollow, on the southwestern side of the Little Chazy River Valley. The elevation of these deposits is close to projected elevation of the Coveville Stage if the Coveville shoreline is extended northward from where Chapman (1937) and Denny (1974) mapped the northernmost Coveville shoreline deposits in the Saranac River Valley, assuming a northward isobase gradient of approximately1.2 m/km. The lower level lies between 206 and 215 meters above sea level and corresponds to the boulder deposits observed at Stop 1. The lower level boulder deposits lie close to the elevation of the upper Lake Fort Ann high stand shoreline (Chapman, 1937; Denny, 1970, 1974).

The northeastern flank of Cobblestone Hill contains several large depressions that we interpret to be kettle holes. The northeastern ends of the kettles rise onto a broad terrace composed of beach deposits (Denny, 1970, 1974) at elevations between 201 and 204 meters above sea level. These beach deposits correspond closely to the elevation of lower Lake Fort Ann.

We believe that these data indicate that the ice margin stood at Cobblestone Hill at the time of the Lake Iroquois breakout and that proglacial water levels in the Champlain Lowland dropped during deposition of the Cobblestone Hill boulder deposits.

Return to the vehicles and continue northwest on the “Scarpit” Road.

0.3	93.2	Note the outcrop of Potsdam Sandstone on your left. The largest boulders on Cobblestone Hill have long dimensions that exceed 3m.
0.2	93.4	The Scarpit Road makes a sharp right turn and the concrete pavement ends. The road emerges onto Altona Flat Rock within 30 meters of the turn. The transition from the northern hardwood forest on Cobblestone Hill to the jack pine barrens on Altona Flat Rock is abrupt at this location.

Incr. Distance (mi)	Cum. Distance (mi)	Description
0.1	93.5	**STOP 5c. Altona Flat Rock Sandstone Pavement Jack Pine Barren.** Park at the USGS observation well.

The large areas of sandstone pavement that comprise the Flat Rocks in Clinton County provide habitat for some of the largest jack pine (*Pinus banksiana*) barrens in the eastern United States (Woehr, 1980; Reschke, 1990). Jack pine is a relatively short-lived (<150 years), shade-intolerant, boreal species that maintains communities on the sandstone pavements because of its adaptations to fire and ability to survive in an area with thin (or absent), nutrient-poor soils.

A large proportion of the pine barrens in northeastern New York is owned by a few public and private sector organizations. The William H. Miner Agricultural Research Institute is the largest landowner of pine barrens with almost 1000 ha (hectares) of jack and pitch pine barrens on Altona Flat Rock. New York State owns an additional 600 ha of the Altona Flat Rock barrens, approximately 100 ha of the Gadway barrens and 200 ha of pine barrens at The Gulf near Covey Hill. The Adirondack Nature Conservancy owns 222 ha of the Gadway jack pine barrens at Blackman Rock.

Plattsburgh State University and the William H. Miner Agricultural Research Institute have collaborated in research and teaching initiatives in the Altona Flat Rock pine barrens for more than 30 years. The hydrogeological equipment and instrumentation that can be seen along the roadsides on Altona Flat Rock are part of the Little Chazy River Laboratory, an instrumented watershed dedicated to undergraduate teaching and research in geology and environmental science. The field site offers an excellent geological, hydrological and ecological setting for illustrating the interdependence of natural processes and the effects of human activities on natural ecosystems. The reader is referred to Adams and Franzi (1994) and Franzi and Adams (1993, 1999) for a more detailed description of the Altona Flat Rock pine barrens.

Return to the vehicles and retrace the field trip route back to the gate at the entrance to the Miner Institute property.

1.4	94.9	Gate at the entrance to Miner Institute Property. Continue south on Blaine Road to Barnaby Road.
1.1	96.0	Blaine Road ends. Follow Barnaby Road about 100 meters to the Slosson Road intersection. Turn left on Slosson Road and proceed east.
0.3	96.3	Slosson Road crosses a low-relief Champlain Sea beach ridge. Continue east on Slosson Road.

Incr. Distance (mi)	Cum. Distance (mi)	Description
1.5	97.8	New York Rte. 22 intersection. Continue east on Slosson Road.
1.6	99.4	Rte. 438 (Fiske Road) intersection. Continue east on Slosson Road to Ashley Road.
0.7	100.1	Ashley Road intersection. Turn left and proceed north 0.5 mi to the entrance of the Kalvaitis gravel pit in the Ingraham Esker.
0.5	100.6	**STOP 6. Ingraham Esker at the Kalvaitis Gravel Pit.**

The Ingraham Esker is one of the most conspicuous glacial landforms in the northern Champlain Lowland. This pit contains esker fan deposits, such as described by Diemer (1988), and deposits resedimented by wave action in the Champlain Sea as described by Denny (1972, 1974). Most of the pit is cut into proximal to medial subaqueous fan gravel and sand. The resedimented deposits consist primarily of fossiliferous gravel beds that occur as dipping bedsets on the western flank of the esker. Individual beds are several centimeters to a few decimeters thick and are laterally continuous for several meters.

Return to the vehicles, turn left at the gravel pit entrance and follow Ashley Road south.

0.5	101.1	Slosson Road intersection. Turn right and proceed west to Rte. 348 (Fiske Road).
0.7	101.8	Rte. 348 (Fiske Road) intersection. Turn left and follow Rte. 348 (Fiske Road) to its end at New York Rte. 22.
2.9	104.7	NY Rte. 22 intersection. Continue south on NY Rte. 22 to Beekmantown.
2.3	107.0	Village of Beekmantown. Continue south on NY Rte. 22 for 0.3 mi to Durand Road.
0.3	107.3	Durand Road intersection. Bear right onto Durand Road and proceed south to Jersey Swamp Road.
0.8	108.1	Jersey Swamp Road intersection. Turn right and proceed west to Military Turnpike.
2.9	111.0	Military Turnpike intersection. Turn left and follow Military Turnpike south to Rte. 374 (Cadyville Expressway).
2.7	113.7	Rte. 374 intersection. Turn left and follow Rte. 374 west to Rand Hill Road in West Plattsburgh.
1.6	115.3	Rand Hill Road intersection. Turn left and procced south to NY Rte 3.

Incr. Distance (mi)	Cum. Distance (mi)	Description
0.1	115.4	NY Rte. 3 intersection. Turn right and proceed west on NY Rte. 3 for 0.7 mi to Stop 7.
0.7	116.1	**STOP 7. Patterson Brook Till Exposure.** Park on the right shoulder of the road and proceed to the exposure on the south bank of Patterson Brook.

The till at this location is a red-brown, sandy, matrix-supported diamicton that contains relatively few large clasts. Its red color is probably derived from the red shale and siltsone of the "Altona Formation" (Landing et al., in press), which is exposed in parts of the St. Lawrence and Saranac hills sections of the Adirondack Upland (Fig. 2).

Turn the vehicles around and proceed east on NY Rte. 3 back to West Plattsburgh.

0.7	116.8	Rand Hill Road intersection in West Plattsburgh. Continue east on NY Rte. 3.
2.4	119.2	NY Rte. 22B intersection. Turn left and continue east on NY Rte. 3 to the Military Turnpike intersection.
0.6	119.8	Military Turnpike intersection. Turn right and follow Military Turnpike south to Stop 8.
1.9	121.7	**STOP 8. Saranac River Till Exposure at Treadwell Mills.** Pull off the road as far as possible and proceed on foot to the exposure on the west bank of the river downstream from the dam.

The till at this location is a gray to grayish brown, silty matrix-supported diamicton that contains many cobble and boulder size clasts. The Beekmantown Group dolostone underlies the exposure.

Turn the vehicles around and follow Military Turnpike north.

1.1	122.8	Rugar Street intersection. Turn right and proceed east on Rugar Street to Prospect Street.
1.8	124.6	Prospect Avenue intersection. Turn left and follow Prospect Avenue north to Broad Street.
0.6	125.2	Broad Street intersection. Turn right on Broad Street and proceed east for 0.8 mi to the Hudson Hall parking lot on the SUNY Plattsburgh Campus.

End of Road Log for 2 June 2007

Road Log for Sunday 3 June 2007

Incr. Distance (mi)	Cum. Distance (mi)	Description
0	0	Leave north parking lot at Hudson Hall and turn left on Broad St. Proceed east on Broad St. for 0.21 mi to the Rugar Street intersection.
0.2	0.2	Rugar Street intersection. Continue east (straight) on Broad Street to the US Rte. 9 intersection (distance to next turn is 0.67 mi).
0.7	0.9	US Rte. 9 intersection. Turn right and follow US Rte. 9 south to a roundabout (rotary) near the former Plattsburgh Air Force Base (PAFB).
0.7	1.6	Make a left turn through the roundabout and proceed east into the former PAFB.
0.1	1.7	Ohio Avenue intersection. Turn right onto Ohio Avenue and proceed south 0.25 mi to the PAFB marina.
0.3	2.0	Turn left onto the marina access road. The marina core was collected near the bicycle path just before the access road intersection. Cross the railroad tracks and park in the lot at the edge of the bluff that overlooks Lake Champlain.
0.1	2.1	**STOP 9. Plattsburgh Air Force Base Marina.**

The bluffs along the shore of Lake Champlain extend for more than 1 km north from the former Plattsburgh Air Force Base marina. The bluffs contain a complete late glacial stratigraphic section; however, at present all the units are not exposed at a single location. A massive gray diamicton lies at the base of the glacial section. The diamicton is exposed at the north end of the bluffs where it overlies striated bedrock. The upper contact is not exposed. The base of the section near the marina consists of more than 3 m of dark gray clayey rhythmites, which were probably deposited as varves in glacial Lake Vermont. The rhythmites occur as clay and silty clay couplets that range from a few centimeters thick in the lower part of the section to thin couplets that rarely exceed a few millimeters in thickness near the top of the unit. Soft-sediment deformation structures are common. Rock and sediment clasts are distributed throughout the unit as individual clasts and in discrete layers along bedding planes. A deformed bed of medium sand that is 0 to 0.2 m thick occurs near the base of the exposed section. The lateral extent of this unit is not known from outcrop, but we believe it represents one or both of the Lake Iroquois or Lake Frontenac flood events.

Incr. Distance (mi)	Cum. Distance (mi)	Description

The lacustrine rhythmites are conformably overlain by 1.5 to 2.0 m of laminated to thinly bedded, fossiliferous marine mud. The entire marine transitional sequence is exposed at this location. The marine mud facies grades upward from thinly laminated, fossiliferous silt and clay rhythmites to horizontally bedded silt and fine sand. The silt and sand unit is approximately 7 m thick and the unit coarsens upward. Individual beds range from a few centimeters to a decimeter or two thick and are generally normally graded.

Return to the vehicles and proceed west across the railroad tracks to Ohio Avenue.

0.1	2.2	Ohio Avenue intersection. Turn left and proceed south on Ohio Avenue to Nevada Oval.
0.2	2.4	Nevada Oval intersection. Turn left and continue south on Nevada Oval to US Rte. 9.
0.4	2.8	US Rte. 9 intersection. Turn left and proceed south on US Rte. 9 to Keeseville (total distance = 10.8 mi to Ausable Chasm).
3.2	6.0	US Rte. 9 crosses the Salmon River at Snug Harbor. MIND THE SPEED LIMIT. Continue south on US Rte. 9.
3.7	9.7	US Rte. 9 crosses the Little Ausable River near the Bear Swamp Road intersection. The Peru coring site is 0.9 west of US Rte. 9 on Bear Swamp Road. Continue south on US Rte. 9. A north-south oriented crag-and-tail hill parallel the west side of the road about 0.25 miles south of the Bear Swamp Road intersection.
1.3	11.0	US Rte. 9 crosses the Ausable River. Continue south on US Rte. 9.
2.5	13.5	**STOP 10. Ausable Chasm Overlook.** US Rte. 9 crosses the Ausable River at Ausable Chasm. Cross the bridge and park in the lot on the right (west) side of the road.

Ausable Chasm is one of the most unique scenic spots in the Champlain Lowland. The Ausable River carved a spectacular post-glacial gorge that exposes more than 135 m of the Keeseville Member of the Potsdam Sandstone. The Keeseville Member upstream from the US Rte. 9 bridge generally consists of planar laminated to rippled, thin-bedded, medium-grained orthoquartzites with rare mudstone (Landing et al., in press). Locally, hummocky cross-stratification and polygonally-cracked mudstones are present. The beds also contain a surprising diversity of ichnofossils including scyphomedusae, microbial structures

Incr. Distance (mi)	Cum. Distance (mi)	Description

and horizontal trackways of *Climactichnites* and *Protichnites*. These facies probably represent a shallow subtidal to emergent sand flat paleoenvironment (Landing et al., in press). A delta built by the Ausable River into the Champlain Sea overlies the sandstone at this location (Denny, 1974).

Return to the vehicles and proceed north for 0.22 mi on US Rte. 9 to Rte. 373.

0.2	13.7	Rte. 373 intersection. Turn right and follow Rte. 373 for 0.22 mi to Mace Chasm Road.
0.2	13.9	Mace Chasm Road intersection. Follow Mace Chasm Road for about a mile to Stop 11.
0.9	14.8	**STOP 11. Bedded Diamictons at Mud Brook.** Park well onto the shoulder of the road and proceed carefully to the base of the Mud Brook exposure.

More than 30 m of crudely bedded diamicton are exposed at this location. Diamicton beds are lenticular to planar in shape and range from a few centimeters to a few meters thick. Stratified interbeds range from thin, discontinuous sand, silt and clay laminae to massive, planar bedded and cross-stratified sand and gravel beds more than a meter thick.

Return to the vehicles and continue south on Mace Chasm Road for about 0.15 mi.

0.2	15.0	Turn right on Soper Rd. and proceed west to Keeseville.
1.6	16.6	US Rte. 9 intersection. Turn left and proceed south on US Rte. 9 to Augur Lake Road. US Rte. 9 rises out of the Ausable River Valley and onto the surface of a delta that was built into Lower Lake Fort Ann. The delta surface elevation is approximately 156 meters above sea level.
0.4	17.0	Augur Lake Road intersection. Turn right and follow Augur Lake Road 0.15 mi west to the entrance to Keeseville Industrial Park.
0.2	17.2	Keeseville Industrial Park entrance. Turn right and proceed 0.2 mi to Stop 12.

Incr. Distance (mi)	Cum. Distance (mi)	Description
0.2	17.4	**STOP 12. Keeseville Industrial Park.**

The Keeseville Industrial Park section is exposed in a landslide scar on the south bank of the Ausable River. The river is deeply incised into a deltaic terrace graded to Lower Lake Fort Ann. The surface elevation of the delta surface is approximately 155 m.

A massive to crudely bedded, dark gray diamicton forms the base of the section. The diamicton is overlain by approximately 2 m of rhythmically laminated silt and clay couplets. Clay laminae are generally 1 cm or less thick and the silt laminae or beds range from about 0.5 to 4 cm thick. The silt beds are commonly internally laminated. The rhythmite section contains about 68 couplets. The rhythmite section is conformably overlain by approximately 7 m of deltaic silt and sand that coarsen upward to sand and gravel.

The sediments at Keeseville Industrial Park record ice recession from the Ausable Valley. The basal diamicton is interpreted to be a till and thus represents ice cover. The rhythmites are probably varves and thus record inundation of the lower Ausable Valley by proglacial Lake Coveville. Assuming that the entire varve sequence represents proglacial Lake Coveville and the overlying silt and sand record the drop of proglacial lake level to Upper Lake Fort Ann, then Coveville occupied the lower Ausable Valley for approximately 68 varve years before proglacial lake levels dropped to the Upper Lake Fort Ann level. The ice front would have receded about 30 km north to the Cobblestone Hill Ice margin over this time interval at an average retreat rate of approximately 0.44 km/yr.

Return to the vehicles and return to Augur Lake Road.

0.2	17.6	Augur Lake Road intersection. Turn left on Augur Lake Road and proceed west to US Rte. 9.
0.2	17.8	US Rte. 9 intersection. Turn right and follow US Rte. 9 to its intersection with NY Rte. 22.
4.1	21.9	NY Rte. 22 intersection. Turn left and follow NY Rte. 22 south to Willsboro.
3.8	25.7	NY Rte. 22 crosses the outlet of Long Pond, where the Long Pond core was collected. Continue south on NY Rte. 22.
4.4	30.1	NY Rte. 22 crosses the Boquet River in Willsboro. Continue south on NY Rte. 22 to the Middle Road intersection.
0.4	30.5	Middle Road intersection. Bear right onto Middle Road and proceed south to NY Rte. 22.

Incr. Distance (mi)	Cum. Distance (mi)	Description
0.8	31.3	The Willsboro core site is on the right. Continue south on Middle Road.
2.7	34.0	NY Rte. 22 intersection. Turn right onto NY Rte. 22 and proceed west then south to the Cook Road intersection in Whallonsburg.
3.9	37.9	NY Rte. 22 crosses the Boquet River. The Cook Road intersection is just ahead. Turn right onto Cook Road and proceed north to Stop 13.
0.9	38.8	**STOP 13. Whallonsburg Landslide and Core Site.**

The Whallonsburg slump occurred during the night on 28 July, 1987, following localized light to moderate thunderstorm activity. The slump involved the eastward displacement of 0.9 ha of Pleistocene lacustrine sediment on a cut bank of the Boquet River. This portion of the Bouquet Valley has a history of landslide activity (Newland, 1938; Whitcomb, 1938; Buddington and Whitcomb, 1941).

The 1987 Whallonsburg slump was described by Franzi et al. (1988). The slump is roughly rectangular in plan with an average length of 85 m and an average width of 110 m. The crown is 16 m above stream level. A 0.2 ha mass of highly plastic clay and alluvial sediment was raised to a height of 4.5 m in the stream channel at the toe of the slide. The slump consists of a large primary mass that was involved in the initial slump and two smaller retrogressive slumps.

The slump exposed 0.5 to 1 m of coarse sand and gravel over 4 m of thinly bedded to rhythmically laminated clay and silt at the head scarp. Rhythmites at the base of the head scarp exposure contain IRD and are interpreted to be fresh water varves. The clays are characterized by high plasticity and natural water content with low bulk density and shear strength. The lacustrine unit grades upward from soft, blue-gray, thinly laminated, clay and clayey silt rhythmites to stiff, brown, thinly bedded silt and clay. The textural variation probably reflects the effects of shoaling or infilling during the waning stages of the lake's history or they may represent early Champlain Sea deposits. The overlying gravel is part of a large terrace that is graded to late Pleistocene marine deltas near Willsboro.

Incr. Distance (mi)	Cum. Distance (mi)	Description

Hillslope reconstructions based on slide-deposit morphology and laboratory analysis of the long-term shear strength of undisturbed clay samples were used to determine the geological controls on slope stability (Franzi et al., 1988). The factor of safety (FS = resisting forces/driving forces) of the reconstructed pre-slide slope was determined using the slope stability model STABR (Duncan and Wong, 1985) under a range of "likely" pore pressure conditions. The results indicate that the pre-slide FS was close to the stability threshold value of 1.0 (range 1.0 to 1.2). Although the slump may have been triggered by rainstorm activity, the actual cause is probably related to pore-pressure conditions at the clay-bedrock contact and long-term processes such as cut-bank erosion at the toe of the slope or fissure development in the upper, overconsolidated clay.

Ridge (pers. comm., 2006) visited this site shortly after the slump and counted about 250 thinly laminated varves. He also collected oriented samples from the face of the exposure for remnant paleomagnetic analysis. Ridge et al. (1999) correlated the paleomagnetic record of these varves to NEVC years 7200–7600 (Fig. 20). The age of these varves was estimated to be 11,800–11,400 ^{14}C years B.P. or about 13,800–13,400 calibrated years B.P. (Ridge et al., 1999).

We collected two cores in the vicinity of the slump in June, 2006 using a USGS truck-mounted vibra-core. Approximately 1.5 m of medium sand overlies more than 22 m of silt and clay that becomes finer downward. We were not able to penetrate the entire Pleistocene section but seismic profiling in 1987 (Franzi et al., 1988) indicated that a dense layer, presumably bedrock, lies at a depth of about 23 m at the landslide head scarp. The base of the measurable varve section starts at about 9.45 m depth and continues upward for 180 couplets to about 8.25 m depth, where a significant change in varve texture, composition and bedding occurs (Fig. 12). We suggest that the change in varve character marks the beginning of the first flood event. There are 48 varves in the flood sequence, which also show a significant increase in the presence of *Candona*. A second interval of IRD-rich rhythmites about 6 cm thick occurs at a depth of 7.9 m in the core. This interval may mark the second flood event and is equivalent to the sand layer at the base of the 168 varves in the Peru-3 core. The record above 7.75 m in this core is truncated by coarser material as the existing Fort Ann phase sediments were probably removed after the drop to Champlain Sea level.

Return to the vehicles and proceed west on Cook Road.

0.8	39.6	Leaning Road intersection. Turn right and proceed north to County Rte. 12.
2.0	41.6	Rte. 12 intersection. Turn left and proceed west on Rte. 12.

Incr. Distance (mi)	Cum. Distance (mi)	Description
2.3	43.9	Rte. 12 turns sharply right. Continue north.
0.4	44.3	Rte. 12 turns sharply left. Continue west.
4.9	49.2	Rte. 12 crosses Interstate 87 at the Exit 32. Continue west to the US Rte. 9 intersection in Lewis.
1.7	50.9	US Rte. 9 intersection. Turn right and proceed north on US Rte. 9 to Pulsifer Lane.
1.1	52.0	Pulsifer Lane intersection. Turn left and proceed to the entrance of the NYCO Wollastonite quarry.
0.6	52.6	**STOP 14. NYCO Wollastonite Quarry.**

The quarry operators excavated deeply into ice-marginal stratified drift and diamicton on the flank of Oak Hill. The composition, stratification and texture of these deposits are highly variable. The stratified sediment probably represents sedimentation by ice-marginal streams flowing from the "Channel Belt" (Kemp and Alling, 1925) and local impoundments. The diamictons are probably till or sediment flow deposits.

Return to the vehicles and retrace the field trip route back to Interstate 87.

0.6	53.2	US Rte. 9 intersection. Turn right and proceed south to Lewis.
1.1	54.3	Turn left at Lewis and proceed east to Interstate 87.
1.7	56.0	Interstate 87—End of Trip. The SUNY vans will return to Plattsburgh via Interstate 87.

End of Road Log for Sunday 3 June 2007

Northeast Friends Of The Pleistocene Reunions (1934-2007)

Reunion		Leaders	Area
1.	1934	George White, J.W. Goldthwait	Durham to Hanover, NH
2.	1935	Dick Flint	New Haven to Hartford, CT
3.	1936	Kirk Bryan	SE Rhode Island to Cape Cod, MA
4.	1937	J.W. & Dick Goldthwait, Dick Lougee	Hanover to Jefferson, NH
5.	1938	Charlie Denny, Hugh Raup	Black Rock Forest, NY
6.	1939	Paul MacClintock, Meredith Johnson	Northern NJ (drifts)
7.	1940	Kirtley Mather, Dick Goldthwait	Western Cape Cod, MA
8.	1941	John Rich	Catskill Mtns., NY
		1942-45 no meetings during war years	
9.	1946	Lou Currier, Kirk Bryan	Lowell-Westford area, MA
10.	1947	Earl Apfel	Eastern Finger Lakes, NY
11.	1948	D.F. Putnam, Archie Watt, Roy Deane	Toronto to Georgian Bay, ONT
12.	1949	Paul MacClintock, John Lucke	NJ ("Pensauken problem")
13.	1950	O.D. Von Engeln	Central Finger Lakes, NY
14.	1951	John Hack, Paul MacClintock	Chesapeake, MD (soils/stratigraphy)
15.	1952	Dick Goldthwait	Central OH (tills)
16.	1953	Lou Currier, Joe Hartshorn	Ayer quad, MA (outwash sequences)
17.	1954	Charlie Denny, Walter Lyford	Wellsboro-Elmira-Towanda, PA–NY
18.	1955	Paul MacClintock	Champlain lake and sea, NY
19.	1956	Nelson Gadd	St. Lawrence Lowland, QUE
20.	1957	Paul MacClintock, John Harris	St. Lawrence Seaway, NY
21.	1958	John Hack, John Goodlett	Appalachians, Shenandoah, VA
22.	1959	Alexis Dreimanis, Bob Packer	Lake Erie, ONT (till bluffs)
23.	1960	Ernie Muller	Cattaraugus Co., western NY
24.	1961	Art Bloom	SW Maine (marine clay; ice margins)
25.	1962	Cliff Kaye, Phil Schafer	Rhode Island (Charleston Moraine etc.)
26.	1963	Hulbert Lee Lower	St. Lawrence Lowland, QUE
27.	1964	Cliff Kaye	Martha's Vineyard, MA
28.	1965	Joe Upson	Northern Long Island, NY
29.	1966	Nick Coch, Bob Oaks	Southeast VA (scarps; stratigraphy)
30.	1967	Hal Borns	Eastern ME (moraines; glaciomarine)
31.	1968	Carl Koteff, Bob Oldale, Joe Hartshorn	Eastern Cape Cod, MA
32.	1969	Nelson Gadd, Barrie McDonald	Sherbrooke area, QUE
33.	1970	Dick Goldthwait, George Bailey	Mt. Washington area, NH
34.	1971	Gordon Connally	Upper Hudson Valley, Albany, NY
35.	1972	Art Bloom, Jock McAndrews	Central Finger Lakes, NY
36.	1973	Don Coates, Cuchlaine King	Susquehanna-Oswego Valleys, NY–PA
37.	1974	Bill Dean, Peter Duckworth	Oak Ridges-Crawford Lake, ONT
38.	1975	George Crowl, Gordon Connally Bill Sevon, Les Sirkin	Delaware Water Gap to Poconos, PA
39.	1976	Bob Jordan, John Talley	Coastal Plain, DE
40.	1977	Bob Newton,	Ossipee area, NH
41.	1978	Denis Marchand, Ed Ciolkosz, Milena Bucek, George Crowl	Central Susquehanna Valley, NY
42.	1979	Jesse Craft	NE Adirondack Mtns., NY
43.	1980	Bob LaFleur, Parker Calkin	Upper Cattaraugus, Hamburg, NY
44.	1981	Carl Koteff, Byron Stone	Nashua Valley, MA

Reunion		Leaders	Area
45.	1982	Pierre LaSalle, Peter David, Michelle Bouchard	Drummondville, QUE
46.	1983	Woody Thompson, Geoff Smith	Augusta–Waldoboro area, ME
47.	1984	Peter Clark, J.S. Street	St. Lawrence Lowland, NY
48.	1985	Ed Evenson, Jim Cotter, Dave Harper Carl Koteff, Jack Ridge, Scott Stanford, Ron Witte	Great Valley, NJ–PA
49.	1986	Tom Lowell, Steve Kite	Northernmost ME
50.	1987	Carl Koteff, Janet Stone, Fred Larsen, Joe Hartshorn	Connecticut Valley–Lake Hitchcock CT–MA
51.	1988	Ernie Muller, Duane Braun, Bill Brennan Dick Young	Genesee Valley, NY
52.	1989	Pierre LaSalle, Andree Blais, Denis Demers, Michel Lamothe, Bill Shilts	Mid St. Lawrence Lowland, QUE
53.	1990	Ralph Stea, Bob Mott	Halifax region, NS
54.	1991	Jack Ridge	Western Mohawk Valley, NY
55.	1992	Bob Dineen, Eric Hanson, Bob LaFleur, Dave DeSimone	Lower Mohawk Valley, NY
56.	1993	Carol Hildreth, Richard Moore	Contoocook–Souhegan–Piscataquog Valleys, NH
57.	1994	Duane Braun, Ed Ciolkosz, Jon Inners Jack Epstein	Eastern PA
58.	1995	Woody Thompson, Tom Davis, John Gosse, Bob Johnston, Bob Newton	Portland–Sebago Lake–Ossipee Valley ME
59.	1996	Hal Borns, Chris Dorion, Joe Kelley, Karl Kreutz , Dave Smith, Woody Thompson, Rick Will	Glaciomarine deposits, eastern ME
60.	1997	Scott Stanford, Ron Witte	Northern and central New Jersey
61.	1998	Les Sirkin	Long Island, NY
62.	1999	Ben Marsh	Periglacial landscapes, central PA
63.	2000	Julie Brigham-Grette, Tammy Rittenour, Janet Stone, Jack Ridge, Al Werner, Dena Dincauze, Ed Klekowski, Richard Little	Glacial Lake Hitchcock, MA
64.	2001	Najat Bhiry, Jean-Claude Dionne, Martine Clet, Serge Occhietti, Jehan Rondot	Quebec City region, QC
65.	2002	Woody Thompson, Carol Hildreth, Dick Boisvert. Chris Dorion, Brian Fowler	Northern White Mtns., NH
66.	2003	Fred Larsen, Steve Wright, George Springston, Richard Dunn	Central VT
67.	2004	Duane Braun, Jon Inners, Jack Ridge	Great Bend–Tunkhannock, northeast PA
68.	2005	Don Pair, Bill Kappel	Onondaga Valley, NY
69.	2006	Roger Hooke, Alice Kelley, Hal Borns, George Jacobson, Brian Robinson, David Sanger	Penobscot Lowland, Central Maine
70.	2007	David Franzi, John Rayburn, Peter Knuepfer, Tom Cronin	Champlain Lowland, NY