PROGLACIAL SEDIMENTOLOGY AND PALEOECOLOGY OF THE TONAWANDA BASIN, WESTERN NEW YORK: IMPLICATIONS FOR A LATE WISCONSINAN PROGLACIAL ENVIRONMENT.

By

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THESIS

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ABSTRACT

PROGLACIAL SEDIMENTOLOGY AND PALEOECOLOGY OF THE TONAWANDA BASIN, WESTERN NEW YORK: IMPLICATIONS FOR A LATE WISCONSINAN PROGLACIAL ENVIRONMENT.

by HEIDI HARLENE NATEL

Proglacial lakes occur in ice contact environments along the terminus of glaciers. Water levels in proglacial lakes are ephemeral owing to the varying ablation rates of glaciers. Proglacial lakes can be overridden by advancing glaciers, in-filled with sediments by glaciofluvial deposition, or persist for hundreds to thousands of years. In modern glacial environments, proglacial lakes exist along the terminus of the Bering Glacier; a temperate, piedmont glacier in Alaska. Rates of sedimentation in modern proglacial lakes such as those at the Bering Glacier are used as a reference to sedimentation rates along the Ontario Lobe of the Laurentide ice sheet in western New York.

Recent coring of eastern parts of glacial Lake Tonawanda suggest its existence as an ice-contact lake which coincided with beaded-esker and delta development. A mollusk assemblage has been recovered from cores with depths that extend from 2 to 8 meters, and grade upward from a basal till and diamicton through several meters of lacustrine clays and marls, overlain by a meter of peat. Cores were sampled at 10 centimeter intervals from strategic locations of the Holley Embayment of glacial Lake Tonawanda. A volume fraction was taken from undisturbed cores, yielding a rich shelly fauna from marl horizons within each core. Textural analyses and percent organic matter suggest a transition at 140-150 cm depth interval, coinciding with an environmental transition from lacustrine to muckland.

Gastropod species identified represent lacustrine, pond and near-shore, and riverine species, representative of a natural lacustrine evolution to a bog or muckland. A similar species development exists in abandoned lake beds at the Matanuska and Bering Glaciers, Alaska. Appearance and disappearance of species coincide with transitions in the depositional environment from deep-water, ice-contact, to pond and bog stages.

Geochemical analysis of Tonawanda sediments (including rare earth elements (REE)) shows chemical signatures consistent with source rocks overridden by the Laurentide Ice Sheet. REE data suggest a correlation of La/Sm ratios with carbonate rock of the Lockport Formation, and high concentrations of Al that can be correlated with clay concentrations, thus suggesting a glacial flour signature associated with lacustrine sedimentation.

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INTRODUCTION

The chronology of the glacial landscape of western New York has been described by many workers (Calkin and Muller, 1993; Fairchild, 1934; Muller and Calkin, 1993; Muller and Prest, 1985; and others; Figure 1). Sediments in the Tonawanda Basin have been previously characterized in several ways. Studies by Kindle and Taylor (1913) characterized the sediments as fluvial deposits associated with spilling of the ancestral Niagara River onto the Tonawanda plain. Later studies referred to the Tonawanda basin as a non-glacial lake basin and relict of such glacial lakes as Warren, Lundy, and Dana (D'Agostino, 1958; Muller, 1977b, b; Muller and Prest, 1985; Calkin and Fenestra, 1985; Fairchild, 1895). The purpose of this study is to characterize the sediments in the Tonawanda Basin with respect to adjacent glacial morphology to better understand the paleoecology associated with the retreating Laurentide Ice Sheet in western New York. To characterize the sediments, hand-augered cores were collected and analyzed for particle size, organic content, carbonate content, trace element geochemistry (including rare earth elements), macrofossils, and magnetic susceptibility. Macrofossil and landform analyses were conducted at the Bering Glacier and Matanuska Glacier, Alaska. These provide a modern analog for ice-contact, proglacial and periglacial environments for the purposes of comparison.

The geomorphology adjacent to the Tonawanda basin has been well documented (Muller, 1977b; Cadwell et al., 1988). The analyses suggest a different depositional environment, then previous studies. The environment evolved from a short-lived ice-contact lake with dynamic outbursts by subglacial conduits with an actively retreating ice-front that lowered lake level to a lower, pond stage in the Tonawanda Basin (Figure

FIGURE 1: STUDY AREA, WESTERN NEW YORK.

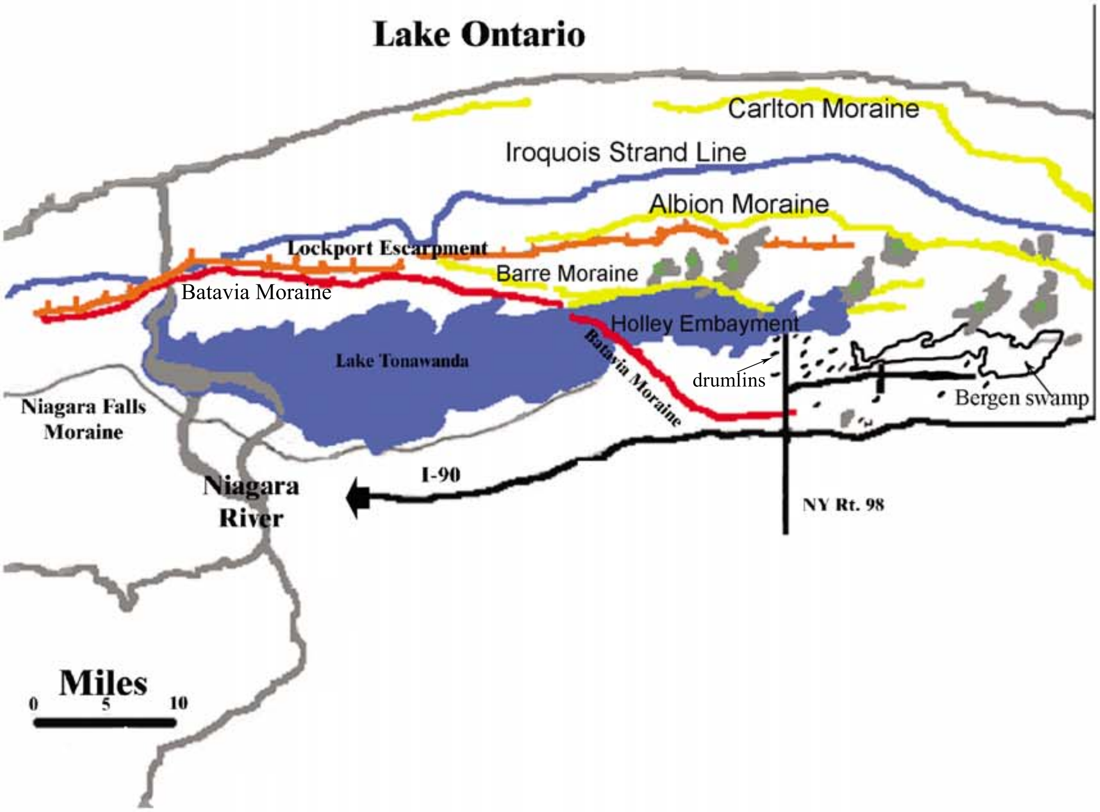


Figure 1a. Map showing the Tonawanda Basin and the Holley Embayment. Field area is within the Holley Embayment. Grey patches between Albion Moraine and Barre Moraine are eskers (with green e on them). Modified version of Muller and Calkin, 1988.

Figure 1b. Photograph of the Holley Embayment of the Tonawanda Basin looking North from Sample Site 1. This region is used for corporate agriculture (crops include: onions, potatoes and corn).

1). The pond stage is represented by development of an extensive biological community associated with warmer, shallow water depths and increased plant growth. The final stage was eutrophication and bog development that continues today in the Tonawanda Basin as a muckland (drained bog environment).

GEOLOGIC SETTING

THE TONAWANDA BASIN, WESTERN NEW YORK

The Tonawanda Basin is located in Western New York (Figure 1). The basin ranges in size from 2-5 miles across (North to South) and stretches 58 miles from East to West, just north of the New York State Thruway (Interstate 90). To the east, the basin is bound by the Clarendon-Linden fault scarp and to the west by the Niagara River. Western portions of the basin are primarily urban, while muckland areas to the east (referred to herein as the Holley Embayment) are either part of the Iroquois Wildlife preserve or are used for corporate agriculture (Figures 2 and 3). The thickness of lacustrine sediments ranges from 1 to 6 m. They are represented by lacustrine silts, clays and marls overlain by about 1 m of organic peat (muck). Farming of the mucklands since the 1920's has resulted in a sediment loss of nearly a 3.2 mm of peat annually by wind erosion (Craig Yonker, personal communication, 2000).

Field work began during the summer of 2000 and was concluded during the fall 2003. The study area encompasses the Holley Embayment (Figure 1). Field work consisted of general mapping and observation of glacial landforms and their structures and hand-augered coring and collection of lacustrine sediments (Figures 2-5).

Figure 2. The Tonawanda Basin.

Figure 2 shows extent of known sediments associated with Glacial Lake Tonawanda. Shoreline based on Muller (1977a) and Cadwell et al. (1988). Orange lines represent esker deposits. Purple lines represent the Barre Moraine and blue lines the limits of Glacial Lake Tonawanda. Red-dashed lines represent the limits of the Batavia Moraine. Map scale is 1:500,000; contour interval is 10 m.

Figure 2. The Tonawanda Basin, western New York State.

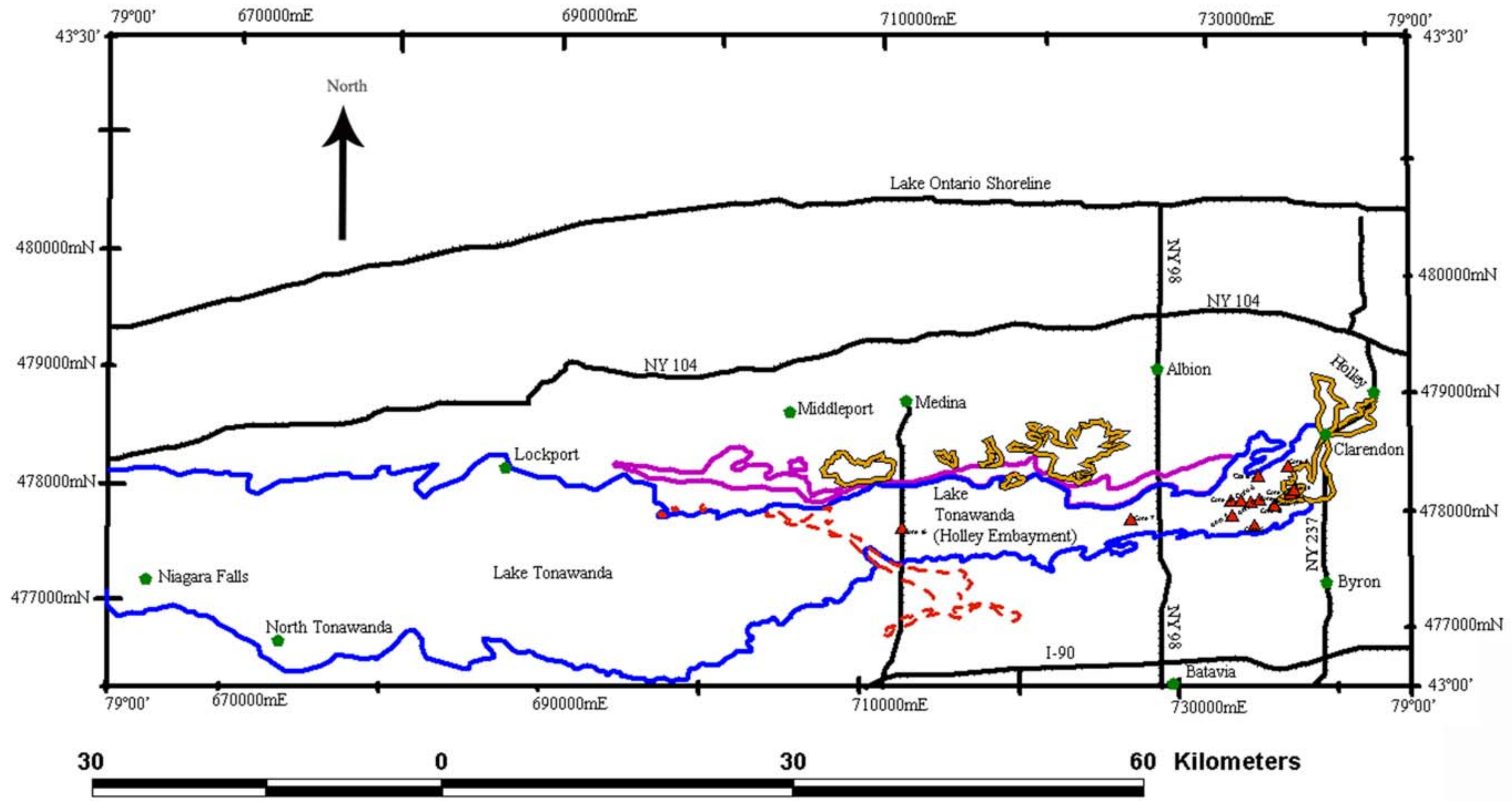


Figure 3. Map of the Holley Embayment.

Figure 3 shows enlarged view of eastern portions of the Tonawanda Basin. Shown are locations of coring sites within the Holley Embayment. Orange lines represent esker deposits. Purple lines represent portions of the Barre Moraine and blue lines the limits of Glacial Lake Tonawanda. Red dashes outline the extent of the Batavia Moraine after Muller, 1977b. Map scale: 1:250,000; contour interval is 10 m.

Figure 3. The Holley Embayment, western New York.

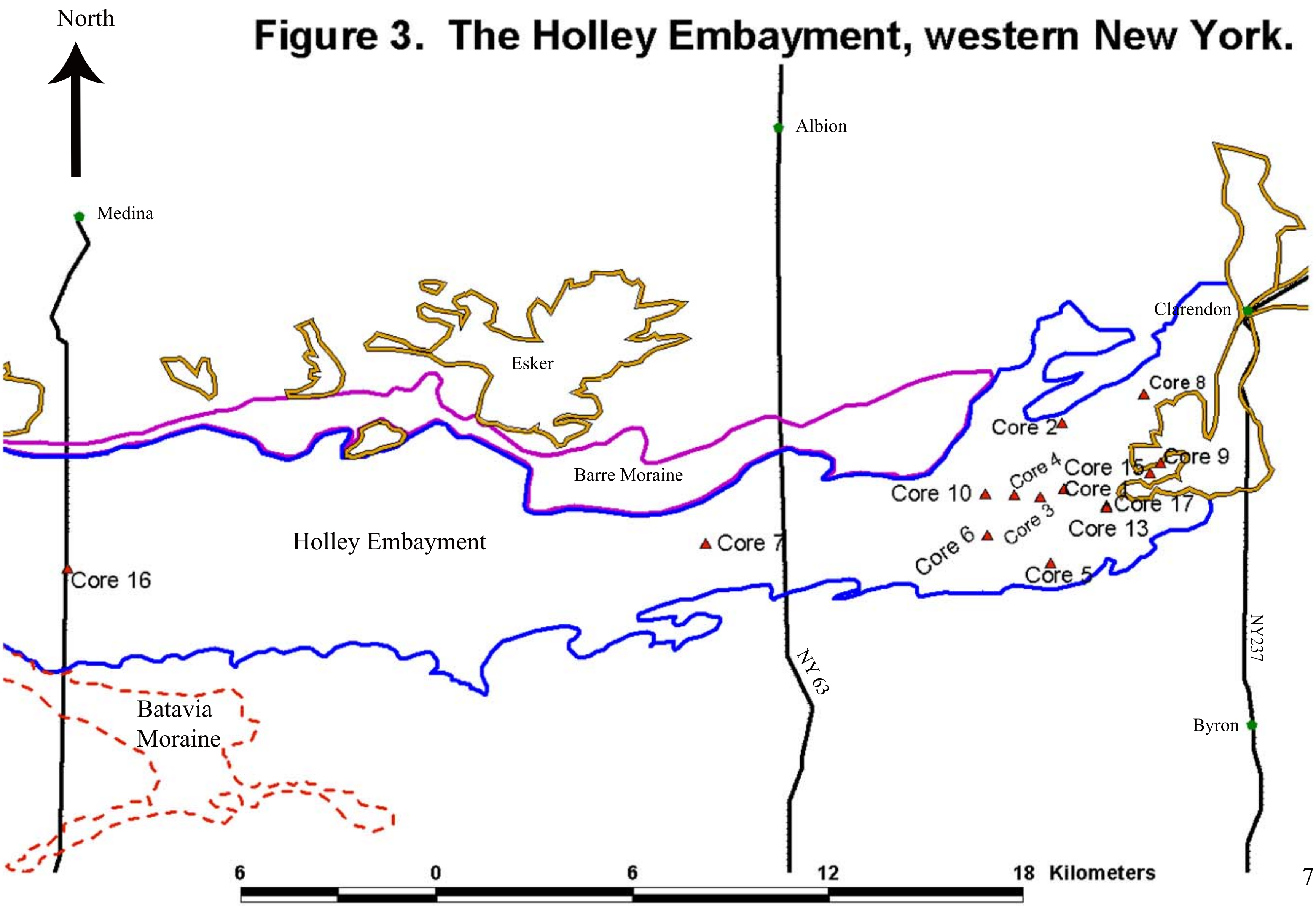


Figure 4. Core Locations.

Figure 4 shows core locations in the eastern Holley Embayment. Not all cores were collected, in some cases only field observations were noted. Locations where cores were collected have been analyzed for corresponding particle size, magnetic susceptibility, loss-on-ignition and/or trace element geochemistry. Where possible, all analyses were conducted for each core. See appendices for raw data associated with these core locations. Orange line represents limits of esker deposit and associated delta. Purple lines represent portions of the Barre Moraine and blue lines the limits of Glacial Lake Tonawanda. Map scale is 1:115,534; contour interval is 10 meters.

Figure 4. Core Locations.

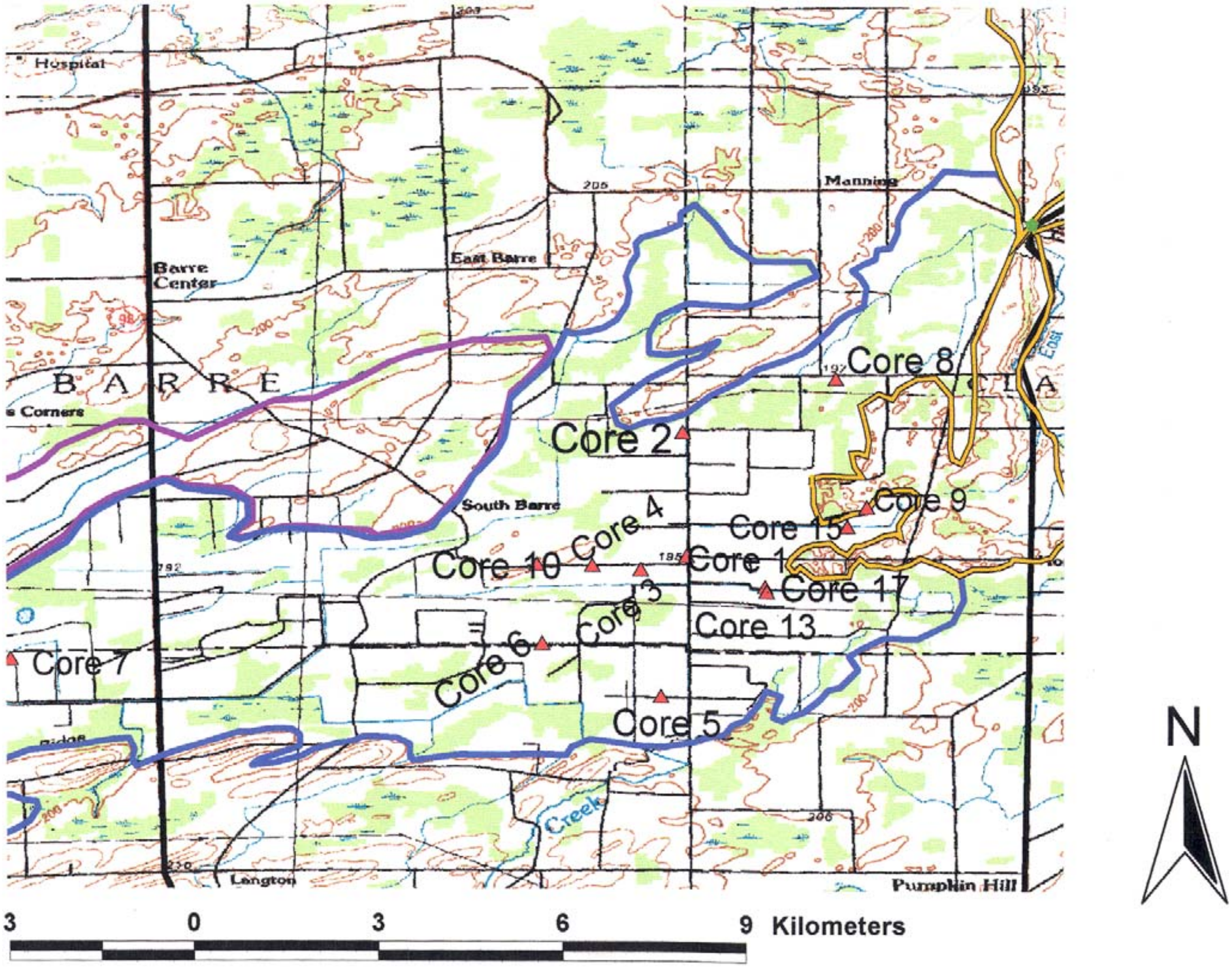
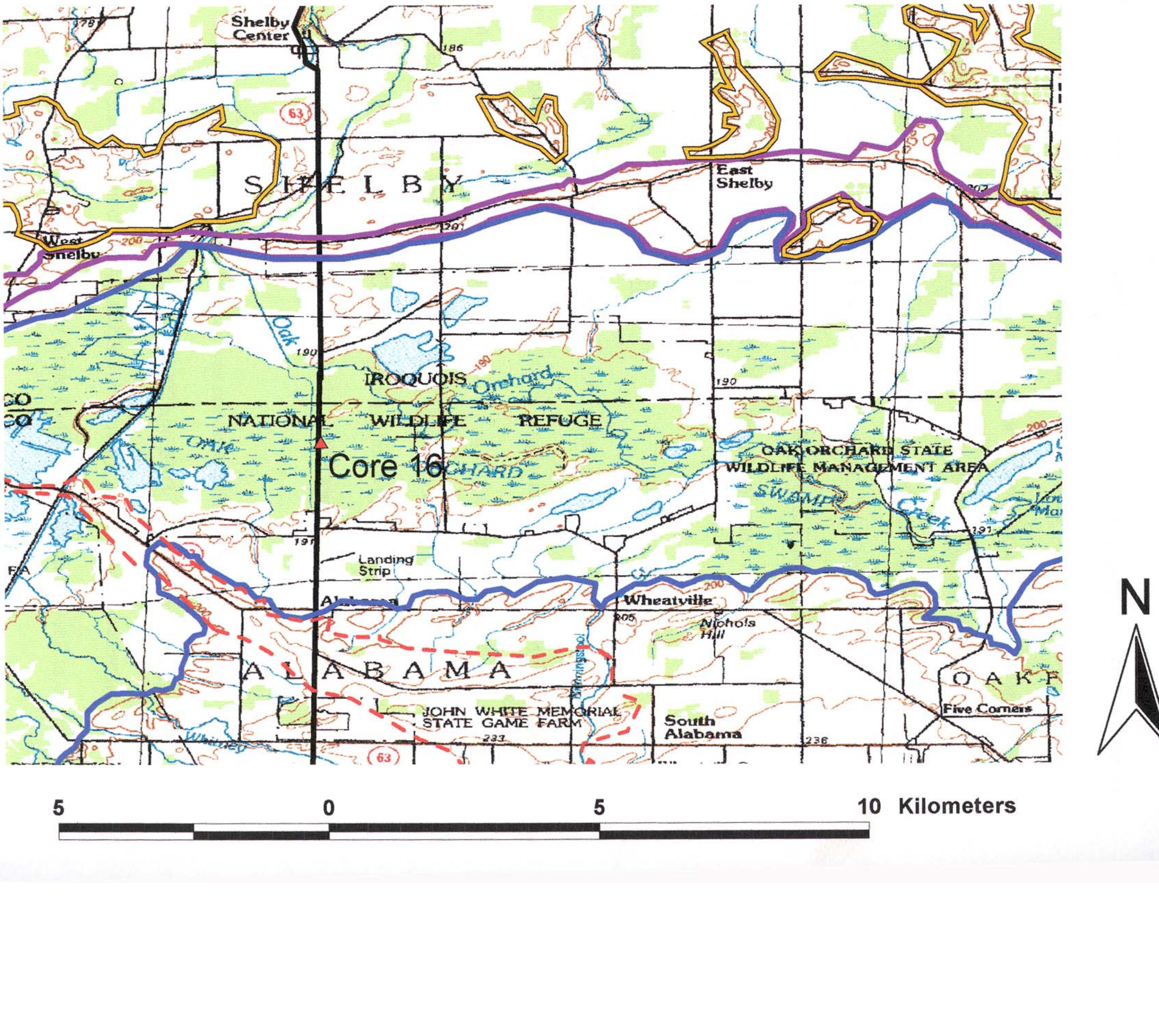


Figure 5. Core Locations.

Figure 5 shows the location of Core 16, which is located on Route 63 just south of the Orleans/Genesee County lines in western New York (N 43º07’24.5”, W78º23’22.5”). The Batavia Moraine runs along the southern edge of the basin and cuts through the basin in a NW direction (marked with red dashes). Calkin and Fenestra (1985) suggest that the Batavia and Vinemount moraines (in Ontario, Canada) are synchronous with one another.

Orange lines represent limits of esker deposits. Purple lines represent portions of the Barre Moraine and blue lines the limits of Glacial Lake Tonawanda. Map scale is 1:100,000; contour interval is 10 meters.

Figure 5. Core Locatio



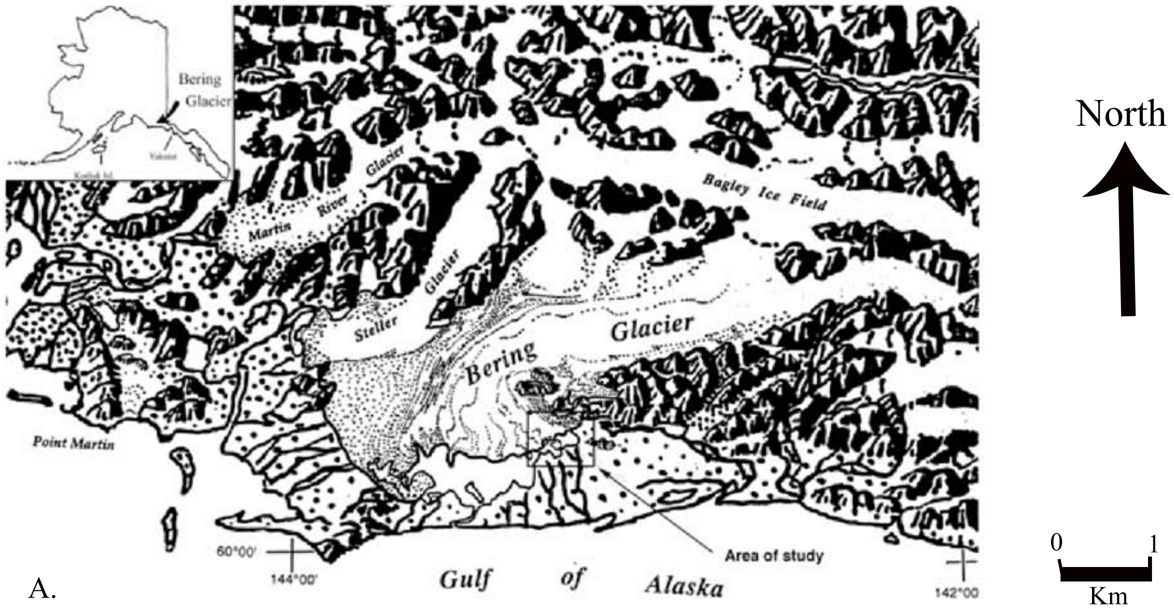
The Tonawanda Basin lies between the (Middle) Devonian Onondaga Escarpment and the (Middle) Silurian Lockport Escarpment to the north (Figure 1). Bedrock beneath the lake sediments are primarily represented by the Silurian Lockport Group, Guelph dolostone, near Clarendon and the Silurian Salina Group, Vernon Shale, towards Elba and points westward (Autin et al., 2001; Natel et al., 2001; Rogers et al., 1990). The Lockport dolostone unit of the Lockport Group surrounds the Tonawanda basin to the north and north-east forming the Lockport Escarpment. Near Lewiston, New York, the Lockport Escarpment (a cuesta) exhibits nearly 240 feet (195 m) of relief. The Onondaga Escarpment to the south is capped by the Onondaga Limestone which locally exhibits about 30 feet (9.14 m) of relief.

THE BERING GLACIER, ALASKA

The Bering Glacier is located along the south-central coast of Alaska, where its melt water discharge forms several ice-contact, proglacial lakes that drain into the Gulf of Alaska (Figure 6). The terminus is located 160 Km east of Prince William Sound and 400 Km southeast of Anchorage. The Bering Glacier is fed by ice that travels 200 km from the Bagley Ice Field in the Chugach Mountains to the terminus of the piedmont lobe, where it coalesces with the Steller Glacier to the west (Muller and Fleisher, 1995; Figures 6).

Field work at the Bering Glacier was conducted from a base camp on Weeping Peat Island along the terminus during the 2002 and 2003 field seasons with members of the Bering Glacier Research Group (BERG; Figure 6). BERG members are faculty and students of the State University of New York, College at Oneonta, as well as others from

FIGURE 6. BERING GLACIER, ALASKA.



A.

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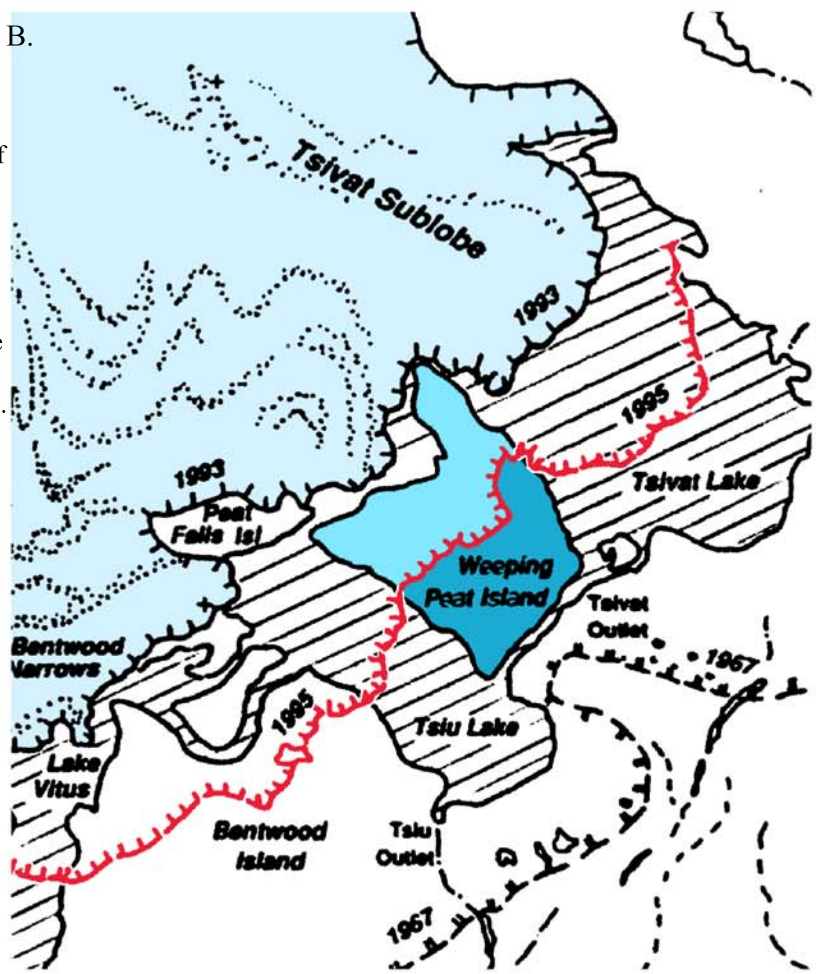


Figure 6a. Map showing the Bering Glacier with an inset map showing its location along the central coast of Alaska. From Fleisher et al., 2003.

Figure 6b. Map of the Bering Glacier and Weeping Peat Island. Also shown are the 1967 and 1995 surge limits, and the 1993 ice front position. From Fleisher et al., 2003.

neighboring universities and members of CREL (Cold Regions Research Engineering Lab). Landform and ice front mapping was conducted during the 2002 field season and was completed during the 2003 as retreat uncovered new landforms (Figures 7-9). Shore lines of water bodies were examined for mollusks. One collection site was located in an ephemeral pond that was about 20-25 cm deep (Figures 7 and 9).

THE MATANUSKA GLACIER, ALASKA

The Matanuska Glacier is a 27 mile-long alpine valley glacier that trends to the northwest descending nearly 12,000 feet from its origins in the Chugach Mountains (www.alaskascenes.com; www.matanuskaglacier.com). Its terminus is nearly 4 miles wide and is located 46 miles northeast of Palmer, Alaska (Figure 10). Drainage from the Matanuska Glacier forms the Matanuska River which drains into the Matanuska-Susitna Valley.

The study area at the Matanuska Glacier is along its northeast flanks which are characterized by a debris covered, passive ice margin. Field work at the Matanuska Glacier took place during June, 2003, during which time selected locations were sought for the collection of mollusk specimens. Two locations were identified, one in an abandoned ice contact, proglacial lake basin that had previously existed for several years (Staci Goetz-Ensminger, personal communication, 2003; Figure 11b). The second location was in a stream that drained melt water from debris covered, ice-cored terrain (Figure 11a). Mollusca from the lake basin were not living; whereas those from the ice-contact drainage were living when collected.

Figure 7. Bering Glacier, Alaska.

Figure 7 shows the location of Weeping Peat Island, Bering Glacier, Alaska. Purple perimeter represents island in 2003. Areas in orange and shades of orange are newly exposed land in 2003. Orange over purple lines represent the boundary of the 1995 push moraine; the maximum extent of the 1993-1995 surge. The Icewall Sandur was deposited in 1995 following a subglacial outburst which interrupted the surge. The Splitlake and Riverhead Sandars were deposited from subsequent outbursts following minor readvances. This glacier has a 25 year surge cycle, with the previous surge occurring in 1966-1967. See Muller and Fleisher (1995) for additional details pertaining to the surge history of the Bering Glacier.

Mollusk samples were collected during June, 2003 on Umlaufland, noted with a green circle and arrow indicating the site. Gastropods collected were living at time of collection. This map was constructed using GPS data overlain on a 1998 Bureau of Land Management aerial photograph.

FIGURE 7: WEEPING PEAT ISLAND, BERING GLACIER, ALASKA, 2003.

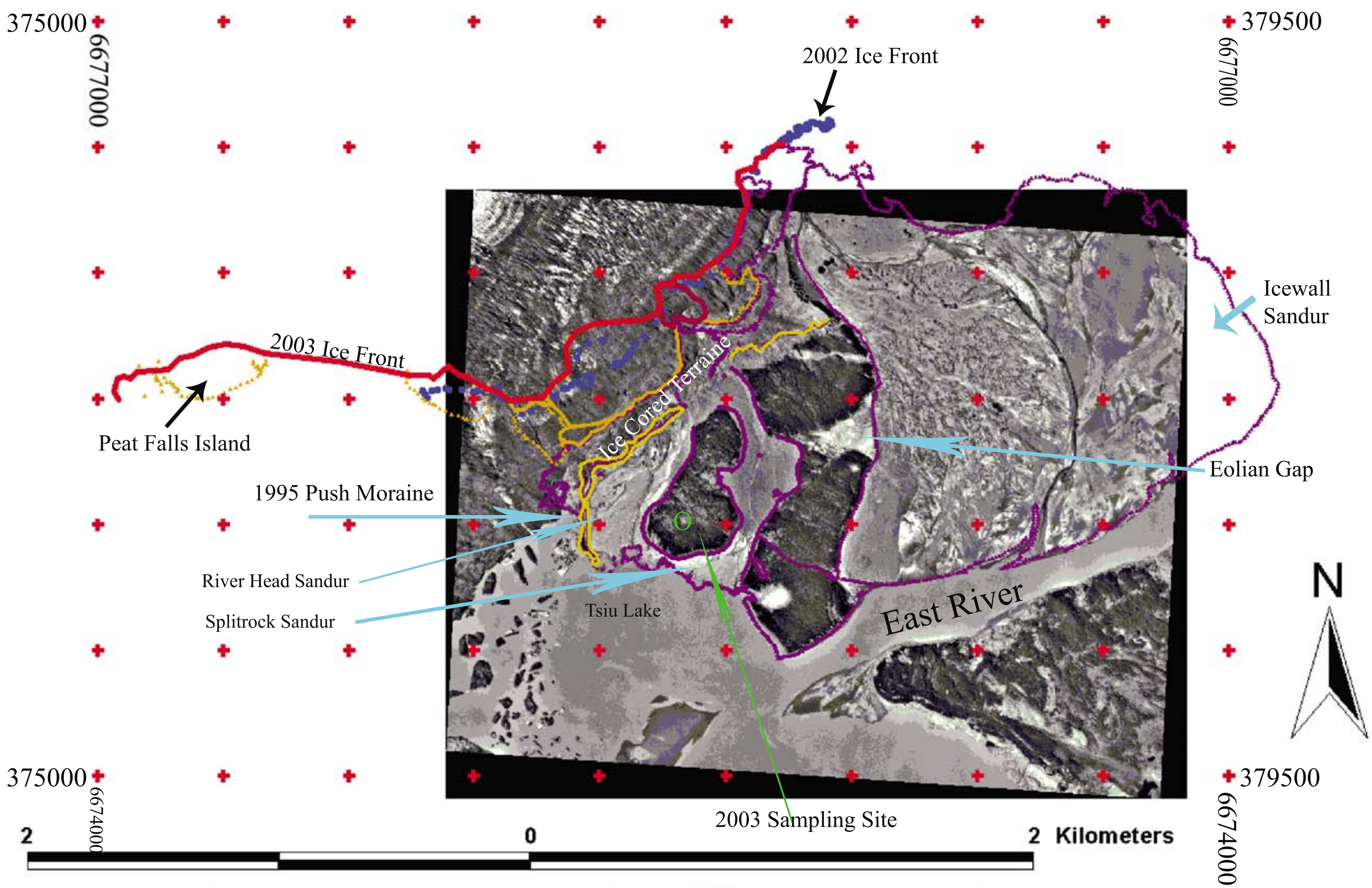


FIGURE 8: BERING GLACIER, ALASKA

Figure 8a. Weeping Peat Island, Bering Glacier Alaska, June, 2003. Looking N-NE across island. Ice front is debris covered and appears dark.

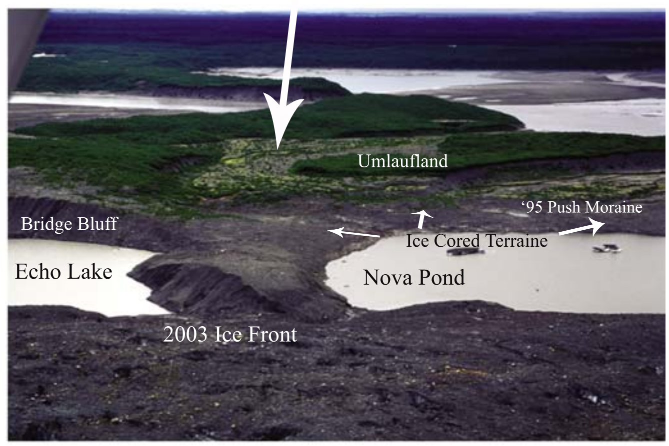
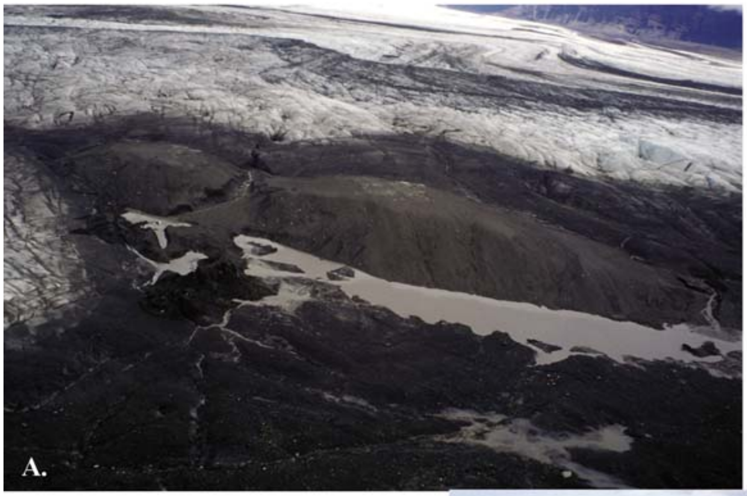


Figure 8b. Looking SE across island. Shown is the locations of the 1995 Push Moraine. Collection site for mollusks was on top of Umlaufland.

Figure 8c. Looking across island to the N-NE. Tsiu Lake is shown in front of Weeping Peat Island.

FIGURE 9. BERING GLACIER, ALASKA.



9a. Peat Falls Island, Bering Glacier, Alaska. Newly uncovered in 2003 by ice retreat from 1995 surge limit. Photo by Eric Natel, 2003.

9b. A view SE across Weeping Peat Island (Riverhead Sandur), Bering Glacier, Alaska. Ice front in foreground. Photo by Eric Natel, 2003.

FIGURE 10. MATANUSKA GLACIER, ALASKA

Figure 10a. Matanuska Glacier, Alaska, during June 2003.

Figure 10b. Map showing location of Matanuska Glacier in Alaska. Modified from image on website (www.matanuskaglacier.com).

North

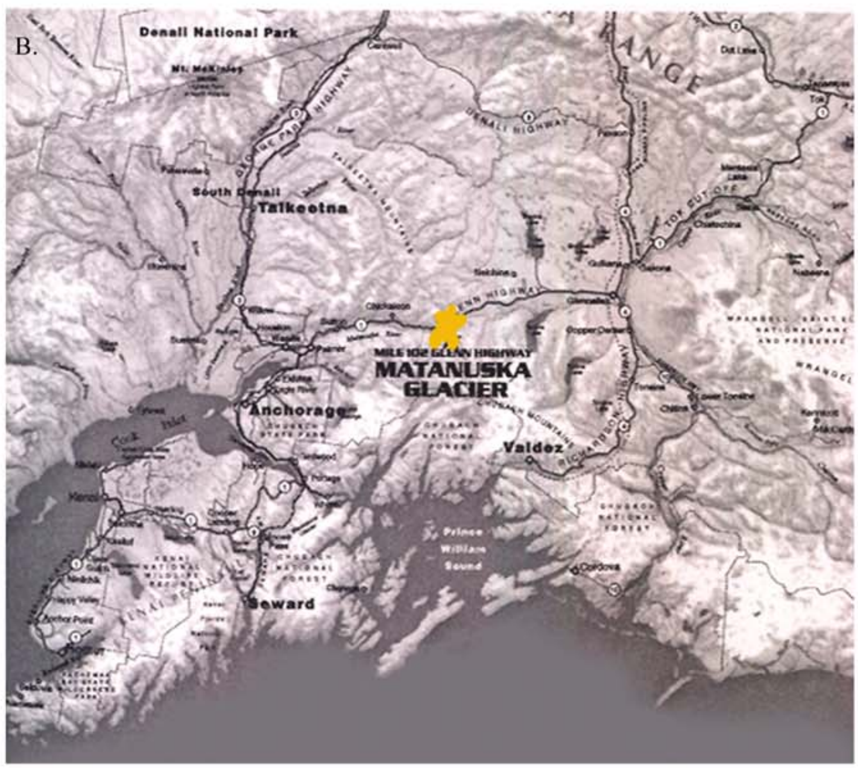
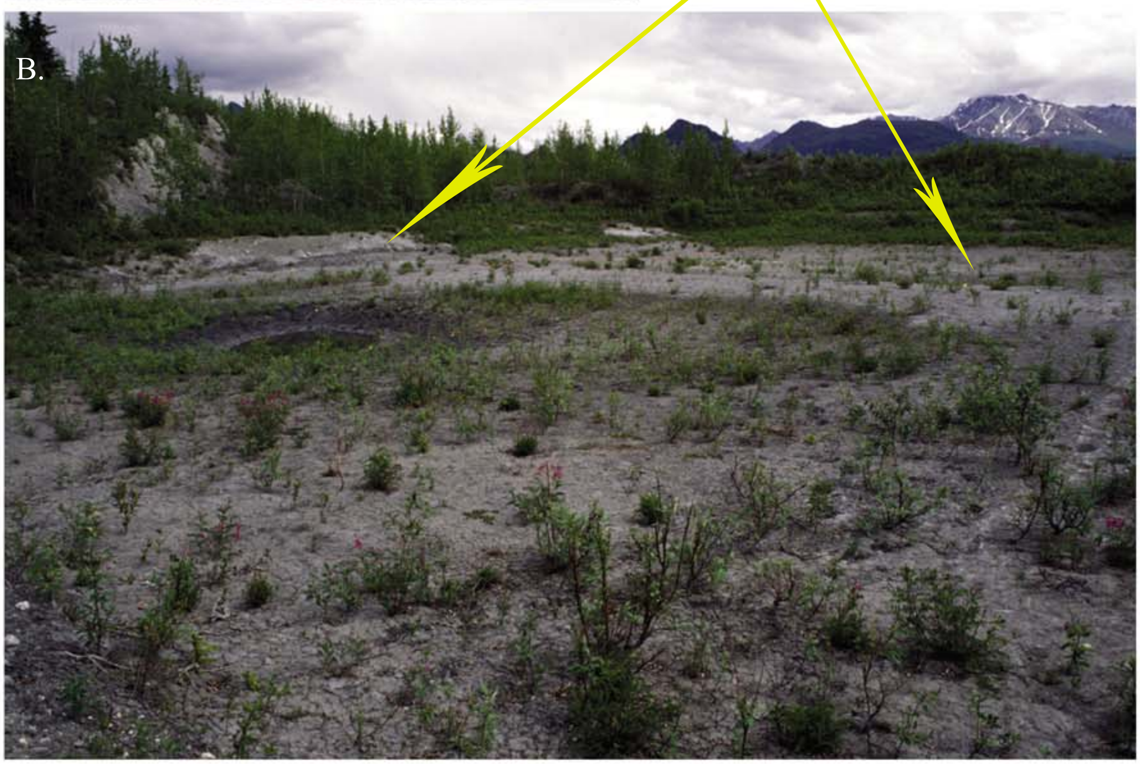


FIGURE 11. MATANUSKA GLACIER, ALASKA.

Figure 11a. Photograph of collection site #1; Matanuska Glacier, drainage from ice corred terraine. June 26, 2003. Living specimens collected. Gastropods were in shallow, cold water.

Figure 11b. Photograph of proglacial lake bed at the Matanuska Glacier, June 27, 2003. Gastropod shells were collected around the perimeter of the lake bed and on banks surrounding the lake.



PREVIOUS WORK

GLACIAL HISTORY OF WESTERN NEW YORK

Glacial landforms in western New York State are the result of several re-advances of the Ontario Lobe of the Laurentide Ice Sheet during the Late Wisconsinan. Calkin (1970) states that the Port Huron Phase of the Laurentide Ice Sheet, Late Wisconsinan, formed the Huron Moraine in southwestern New York approximately 13,000 years BP. Muller and Calkin (1993) suggest that the Huron Moraine formed in the lake bed of Glacial Lake Whittlesey, cross-cutting strand lines, and that as ice retreated to the position of the Marilla Moraine, Glacial Lake Whittlesey gave way to various levels of Glacial Lake Warren. Muller and Calkin (1993) also suggest that the Port Huron maximum is younger than 13,100 years BP based on the dating of the St. Joseph Till, Ontario, Canada, by Gravenor and Stupavsky (1976; Karrow, 1987). The Huron Moraine formed prior to lowering of Glacial Lake Whittlesey to Glacial Lake Warren water levels, and following ice retreat to the position of the Marilla Moraine. Glacial Lake Whittlesey formed in an ice-contact environment during formation of the Marilla Moraine (Barnett, 1985; Muller and Prest, 1985). Cross-cutting relationships suggest that a strand line can overlay a moraine, but not vice versa, as an advancing ice front forming a moraine would otherwise destroy evidence of preexisting strand lines. Thus it is more likely that ice retreated to the position of the Marilla Moraine allowing Glacial Lake Whittlesey to flow over the Huron Moraine at its high water level, before lowering to high water levels of Glacial Lake Warren (Muller and Prest, 1985; Muller and Calkin, 1988).

Following ice retreat from the position of the Marilla Moraine, the Alden Moraine formed in western New York State and the Fort Erie Moraine in Ontario (Muller and Calkin, 1993). Further active ice retreat is recorded by positions of the Buffalo and Niagara Falls Moraines in western New York. Strand lines associated with Glacial Lake Warren III suggest that the lowest lake levels (6 m below phase II) existed when the ice front was positioned at the Batavia Moraine (Muller and Prest, 1985; Muller and Calkin, 1988). The Batavia Moraine represents an end moraine associated with a re-advance of the Ontario Lobe into western New York (Muller and Prest, 1985; Muller and Calkin, 1988). The Batavia Moraine is an arcuate moraine that stretches from about 11 Km northwest of Batavia, New York, to Lockport, New York (Figure 1). Calkin and Fenestra (1985) suggest that the Batavia Moraine is contemporaneous with the Vinemount Moraine in southern Ontario, implying that the Batavia Moraine may have draped itself along the Lockport Escarpment, where Muller (1977a) maps the western extension of the Barre Moraine. Ice retreat northward from the Batavia Moraine is marked by the positions of minor readvances represented by the Barre, Albion and Carlton Moraines. The Barre Moraine cross-cuts the Batavia Moraine at South Lockport, New York, and extends west to the Niagara River (Muller, 1977a). The re-advances of the Laurentide Ice sheet forming the Barre, Albion and Carlton Moraines inhibited eastward drainage in the Ontario Basin (Muller and Prest, 1985; Muller, 1977b). According to Muller and Prest (1985), the Barre Moraine is in the lake bed of Glacial Lake Lundy (located in the Erie Basin), which drained eastward into Glacial Lake Avon (in the Genesee Valley). Similarly, during the formation of the Albion Moraine, eastward drainage was again redirected down the Genesee Valley (Glacial Lake Dawson; Muller, 1977a).

High water level strand lines for Glacial Lake Iroquois indicate its existence between formation of the Albion and Carlton Moraines (12,600 to 12,100 years BP; Muller and Calkin, 1993; Muller and Prest, 1985). Glacial Lake Iroquois gave way to the proto-Lake Ontario following ice retreat sufficient to uncover a lower drainage outlet in Rome, New York, allowing drainage to shift down the Mohawk River Valley (Muller and Calkin, 1993). Eastward drainage down the Mohawk Valley began approximately 11,400 years BP when water levels in Glacial Lake Iroquois dropped to 100 meters below modern day Lake Ontario levels (Coakley and Karrow, 1994; Karrow and Calkin, 1984; Mullins, 1998). The onset of eastward drainage along the ice front would have allowed upland areas on the Lockport Escarpment to drain, causing the lowering of water in the Tonawanda basin to a pond stage in the Holley Embayment (Figures 1-4). Admittedly, Muller and Calkin (1993) suggest that there is poor chronostratigraphic control of Late Wisconsinan ice retreat from western New York, but that chronology depends heavily on end moraine and shore line relationships. This further suggests that rates of sedimentation in ice contact, proglacial lakes along the Laurentide Ice Sheet are based on a chronology that is often contradictory (Muller and Calkin, 1993).

BERING GLACIER, ALASKA

The Bering Glacier is a temperate, alpine glacier whose terminus expands forming a low-relief 40 km-wide piedmont lobe (Muller and Fleisher, 1995). According to Ashley (1987) a temperate lithofacies is associated with the outer 10-100 km of a continental ice lobe during the waning phase. Ashley's (1987) temperate facies consists of four diachronous, superimposed assemblages: a) subglacial lithofacies (deposits consisting primarily of unsorted tills or diamicton); b) ice-contact lithofacies (unsorted and sorted remobilized diamictons, and lacustrine and glaciofluvial deposits); c) ice contact, proglacial lithofacies (sorted fluvial-deltaic and deltaic-lacustrine deposits); d) postglacial lithofacies (well sorted eolian deposits of sand and silt (loess)). Because the Bering Glacier expands into a low relief piedmont lobe (below sea level in places), and its lithofacies compare well with those outlined by Ashley (1987), and is considered to be a temperate glacier, it is used herein as a modern analog to the Ontario Lobe of the Laurentide Ice Sheet in western New York. The piedmont lobe of the Bering Glacier is 40 Km wide, whereas the Ontario Lobe of the Laurentide Ice Sheet was approximately 200 Km wide; consequently the Bering Glacier is a small scale analog to the Ontario Lobe (Muller and Fleisher, 1995; Muller and Prest, 1985).

Muller and Fleisher (1995) provide an accurate account of the surge history of the Bering Glacier. The Bering has surge cyclicity, interpreted from U.S.G.S. aerial photos. Surges are known to have occurred in 1940, 1960, 1966 and most recently in 1993 (Muller and Fleisher, 1995). Figure 7 shows recently mapped landforms adjacent to the eastern margin of the Bering piedmont lobe. Facies comparable with the temperate glacier model of Ashley (1987) are: A) Subglacial lithofacies: represented by subglacial sediments remaining in the former conduit that discharged sediments into the Tsivat Lake basin forming the Icewall sandur. The subglacial sediments (referred to as the “Dogleg” sediments by members of BERG), are primarily stratified sands (Fleisher et al., 2004). The subglacial outburst that formed the Icewall sandur was subsequently followed by another outburst that created the Splitlake and Riverhead sandars. Sandur deposits are composed of stratified sands, gravels and cobbles, with rare lag boulders, and are presently being incised by melt-water drainage. The subglacial, “Dogleg” sediments represent waning flow in the conduit following the outburst. Ashley (1987) identifies unsorted tills and diamictons as representative of subglacial deposits; much of the newly uncovered terrain on Weeping Peat Island is covered by a veneer of till. B) Ice-contact lithofacies: here are represented by the region between the ice front and that labeled "ice cored terrain" (Figure 7). The ice contact terrain is associated with two small ice-contact ponds (Echo Lake and Nova Pond; Figures 7-9). The ice cored terrain is between the ice-contact ponds and the 1995 push moraine (Figure 7). The 1995 push moraine represents the maximum position of ice advance during the 1993/95 surge. C) Proglacial lithofacies is represented by the former Tsivat Lake basin, now the Icewall sandur, and the Tsiu Lake basin and delta. A third, larger ice-contact lake, Lake Vitas, is located farther to the south and southwest along the ice front. D) Eolian lithofacies are represented by a region on high ground terrain adjacent to the Icewall sandur referred to as Eolian Gap (Figure 7). The area referred to as Eolian Gap, is dominated by eolian sheets (Figure 7).

RATES OF SEDIMENTATION IN ICE-CONTACT PROGLACIAL LAKES

Proglacial, ice-contact lakes occur along margins of continental and alpine glaciers (Teller, 1987). During the Late Wisconsinan, ice contact, proglacial lakes developed along the margins of the Laurentide Ice Sheet in Western New York. Modern ice-contact lakes are found along the terminus of the Bering Glacier, Alaska. They are Vitas, Tsivat and Tsiu lakes (Figures 7-9). Glaciolacustrine sediments are often identified by diagnostic rhythmic or varved bedding that indicates cyclic seasonal deposition. Glaciolacustrine sediments are typically fine-grained. However, larger sediment sizes are encountered owing to glaciofluvial processes, bank slumping, drop stones, turbidity currents, under flows, and various forms of sediment gravity flows

(Rovey and Borucki, 1995; Syverson, 1998). Sediment sizes are largest proximal to the ice and decrease distally across a basin. Ice-contact sedimentation rates are influenced by glacial discharge, which is directly related to ablation rates, turbidity currents and various forms of gravity flows.

The Laurentide Ice Sheet is thought to be temperate, indicating temperature gradients that promoted summer ablation and represented by the lithofacies model of Ashley (1987). Glacier characteristics affecting rates of sedimentation in ice-contact lakes include glacier size, subglacial hydrology, sediment sources, and rates of ablation and/or advance. Ice-contact lakes can infill rapidly by sub- or englacial discharge containing high sediment loads. This has been documented at the Bering Glacier, when a subglacial conduit ruptured at the ice front in 1994, effectively interrupting the 1993-95 surge and infilling the southern third of the Tsivat Lake basin during a three day period (Fleisher et al., 1993; Muller and Fleisher, 1995; Figures 6 and 7). Ice-contact lakes may also be displaced by overriding ice (Smith, 1990). Ice-contact, proglacial lake development promotes backwasting of ice fronts by calving, which in turn promotes sedimentation by drop stones and gravity flows along the ice front. When freshwater lakes mix with salt water, sedimentation rates are enhanced by flocculation of suspended sediments. In modern glacial environments, rates of sedimentation can be measured directly or calculated indirectly (Table 1; Leonard and Reasoner, 1999; Pickrill and Irwin, 1983; Fleisher et al., 2003; Gustavson, 1971; Smith, 1990). Ice-contact sedimentation rates at the Bering Glacier were 9.7 m year^{-1} during the early surge period and remained high at 3.3 m year^{-1} post surge (Fleisher et al., 2003). A decade of sedimentation (1991-2000) in Tsiu and Tsivat lakes, resulted in the deposition of 227 m^3 of sediment (Fleisher

TABLE 1: SUMMARY OF LITERATURE SOURCES DEPICTING RATES OF SEDIMENTATION IN ICE-CONTACT, PROGLACIAL LAKES.

Authors	*Proximal Sed. Rate*	*Distal Sed. Rate*	*Suspended Sed. Conc.*	*Rythmites*	*Lake Volume*	*Notes*
Syverson, 1998	1000 mm/yr (9 year average)	14 mm/yr	Alpine	Not related to varves; located in distal portions of lake due to density differences	7.5×10^6 m^3 3.6×10^3 m^3 (min) or 8×10^5 m^3 (max)	Burroughs Glacier, Alaska Calving Lake (l: 1000 m; w: 500 m; d: 10-15 m). Bruce Hills Lake (l: 180 m; w: 7-10 m; d: < 2m in ‘88-90; was l: 400 m; w: 100 m; d: 20 m in 1986).
Leonard and Reasoner, 1999		Hector: 5 mm/yr Crowfoot: 0.7 mm/yr *(calc.)*	Alpine 18 mg/L* (Hector Lake)	4500 yr varve record	3.3×10^8 m^3 1.8×10^6 m^3 1.4×10^8 m^3	Hector Lake (area: 5.5 km^2; d: 87 m); 60 m deep/11300 yrs. Crowfoot Lake (area: .23 km^2, d: 8 m); 8 m deep/11300 yrs. Bow Lake (2.7 km^2; d: 51 m). S to N direction. 45 m deep.
Pickrill and Irwin, 1983	20 mm/ yr	5 mm/ year	Alpine (fjord) 57 mg/L	Rythmites associated with periodic floods	1.9×10^9 m^3	Lake Tekapo, New Zealand;. Rate from Pb^{210} dating and varve counting. Author interchangeably used varves/rythmites in paper l: 6 km; d: 105 m; w: 3 km.
Fleisher and Bailey, 1993	180 mm/ year	---	Alpine; 700-2000 mg/L**	Varves	N/a	Tsiu Lake; Bering Glacier, Alaska.
Fleisher et al., 1993	1789 mm/yr 3869 mm/yr	---	Alpine, piedmont	N/a	N/a	Former Tsiu Lake, Bering Glacier, Alaska Tsivat Lake, Bering Glacier, Alaska
Fleisher et al., 2003	970 mm/yr		Alpine, piedmont	N/a	N/a	Tsivat Lake, Bering Glacier, Alaska
Gustavson, 1971	---	---	Alpine 723 mg/L*	Varves	4.1×10^9 m^3	Malaspina Lake, southeastern Alaska (area: 35 mi^2); d: 45 m).
Gilbert and Shaw, 1981	900 mm/yr	---	Alpine 380 mg/L	Rythmites assoc. with 1 to several day flood events	Small	Sunwapta Lake, Alberta Rocky Mtns. (Volume of lake not calculated by author; referred to as small).
Smith, 1990	8111 mm/yr		Alpine 2370 mg/L	N/a	N/a	Carroll Glacier**, Glacier Bay, Alaska; depth: 8-9.5 m; area: 1.3 km^2 .
Reasoner & Rutter, 1988	0.19mm/yr	0.06mm/yr	Subalpine	N/a	N/a	Area: 0.5 Km^2, depth: 42 m; fed by Oesa and Opabin cirque glaciers.

* Data collected and reported in Smith, 1990, from various sources. All rates of sedimentation converted to mm/year for interpretative purposes. ** Carol Glacier is a marine glacier; sedimentation rates enhanced by flocculation of suspended sediments due to salt water mixing.

et al. ,2003; Table 1). Ice-contact, proglacial lakes along the Laurentide ice front represent a challenge for determining rates of sedimentation. Depth and volume of lacustrine sediments coupled with radiometric ages can be used to determine rates of sedimentation along the Laurentide Ice Sheet (Pickrill and Irwin, 1983). Lacustrine sedimentation in the Tonawanda basin deposited 2-6 meters of silt and clay across the basin. Assuming high sediment rates and turbidity similar to that seen at the Bering Glacier, ice-contact sedimentation in the Tonawanda Basin likely represents a very short period of time, certainly less than a decade, given the size of the Ontario lobe of the Laurentide Ice Sheet compared to that of the Bering Glacier and the depth of sediment (Fleisher et al., 2003; Figure 12).

GEOLOGIC SETTING OF ICE-CONTACT, PROGLACIAL LAKES

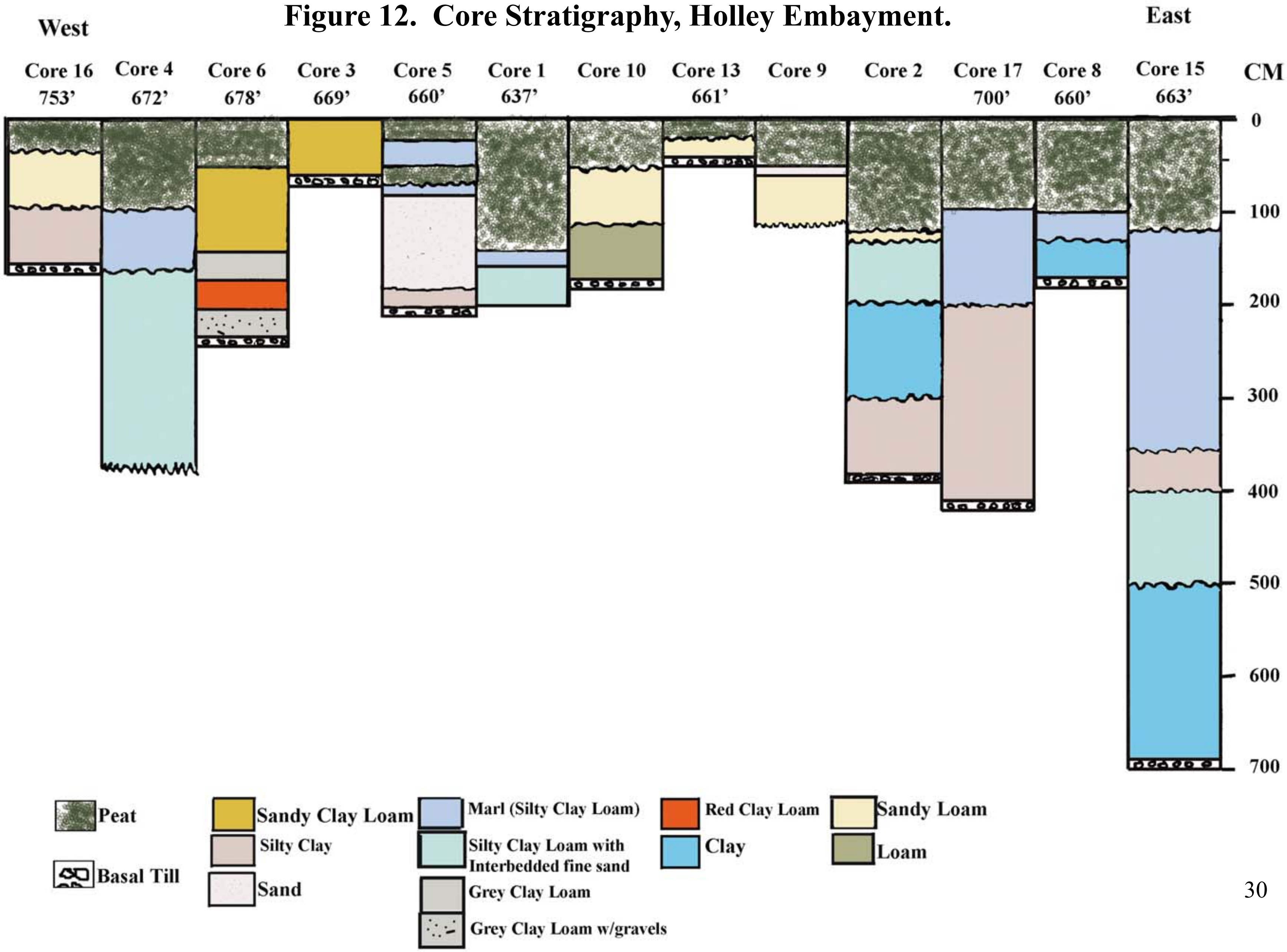
Ice-contact, proglacial lakes in alpine settings often form as a result of ice jams owing to calving of active glaciers, or stagnant ice that dams up narrow alpine valleys (Benn and Evans, 1998). Lakes can also form in glacial scours, bedrock depressions, or behind moraines, drumlins and scarps, such as the Onondaga and Lockport Escarpments (Syverson, 1998). Ice-contact, proglacial lakes that form as a result of ice jams are prone to seasonal depth changes related to jökulhlaup or smaller discharge events.

Ice-contact, proglacial lakes in continental glaciated terrains are commonly located in scoured bedrock basins, where less resistant bedrock has been removed, or are oriented along structural features of the bedrock (Teller, 1987). Lakes can also develop between the ice front and upland topography, as suggested by Francek (1990) regarding the early formation of Glacial Lake Iroquois between the retreating Laurentide Ice Sheet in the Ontario Basin and the Lockport Escarpment of the Appalachian Plateau in New

Figure 12. Core Stratigraphy, Holley Embayment.

Figure 12 shows stratigraphy of cores from the Tonawanda basin, east of the Batavia Moraine to the Clarendon-Linden fault scarp. Most cores terminated in a reddish till, based on refusal of hand augers. Cores 2, 3, 4, 9, 10, and 13 were test holes and data are from field observations and notes only, as no cores were collected at these sites. Stratigraphy for remaining cores is based on collected 10 cm intervals. See appendices for details of field data and laboratory data related to these cores. Figure shows that lacustrine sediments range from 2-6 m, based on the occurrence of silty clay loam, silty clay and clay in the Holley Embayment.

Figure 12. Core Stratigraphy, Holley Embayment.



York State. Shoreline positions suggest that proglacial, ice-contact lakes in the Tonawanda Basin of western New York most likely have been the result of ice retreat to the position of the Barre Moraine and the upland topography to the south (Figures 3 and 4).

Ice-contact, proglacial lakes can be short-lived, existing for a few days, or can persist for thousands of years (Smith and Ashley, 1985). Proglacial, ice-contact lakes may infill rapidly by subglacial and englacial discharge (Martini et al., 2001; Fleisher et al., 2003). Relicts of such lakes along the Laurentide Ice Sheet still persist as lakes, swamps and mucklands (e.g., The Great Lakes, Glacial Lake Tonawanda and the Montezuma wet lands; Teller, 1987; Smith and Ashley, 1985; Natel and Autin, 2001). Ice-contact, proglacial lakes are recorded by the presence of lacustrine sediments and landform expression.

COMPARISON OF ALPINE VERSUS CONTINENTAL ICE-CONTACT, PROGLACIAL LAKE DEVELOPMENT

Rates of sedimentation have been studied in several modern ice-contact, proglacial lakes (Fleisher and Bailey, 1993; Fleisher et al., 2003; Leonard and Reasoner, 1999; Pickrill and Irwin's, 1983; Syverson, 1998). These studies depict alpine environments. Little has been published regarding ice-contact proglacial sedimentation rates of the Laurentide Ice Sheet (Gustavson and Boothroyd, 1987). The primary difference between studies of alpine and continental rates of sedimentation is the manner in which they are determined. Several studies base their rates on measured suspended sediment concentrations (Gilbert and Shaw, 1981; Smith, 1990; Syverson, 1998). Table 1 summarizes current literature sources of sedimentation rates, as well as suspended sediment concentrations (mg/L). Smith (1990) suggests that higher suspended sediment concentrations (~ 9 g/L collected at the margin of the Carroll Glacier, Alaska) are typical early in the summer when melt rates are at their greatest. However, the suspended sediment load of 9 g/L at the Carroll Glacier was enhanced due to salt water mixing with fresh water. Other studies base sedimentation rates on measurements of sediment thickness calibrated with radiometric ages (Pickrill and Irwin, 1983). Differences between calculated and measured sedimentation rates have not been previously discussed in the literature. Sedimentation rates are controlled by glacier ablation and flow, plus availability of sediment to the glacier, as well as characteristics of the glacial lake, such as depth, volume and salinity. Factors such as whether the glacier is surging or passive, season of the year, and texture of glacier bed are important (e.g., type of bedrock and its resistance to glacial abrasion and/or type of sediment). During the summer of 1993, members of BERG recognized the onset of surge conditions at the Bering Glacier (Muller et al., 1993; Muller and Fleisher, 1995; Fleisher et al., 2003). The Bering Glacier pressed ice 1.0 to 1.5 Km onto Weeping Peat Island before a subglacial outburst, in July, 1994, temporarily interrupted ice advance. Outburst sediments infilled the southern third of the Tsivat Lake Basin, along the northern flank of Weeping Peat Island during a three day period (Figures 2 and 3; Fleisher and Bailey, 1993; Fleisher et al., 2003, 2004). At the time of infilling, the Tsivat Basin was approximately 1.5 km by 2.5 Km with an irregular basin topography and depths ranging to 58 m (Fleisher et al., 1993). Rates of sedimentation in Tsivat Lake during June, 1992 were recorded by Fleisher et al. (1993) to be on the order of 3,869 mm year^{-1} (Figure 6; Table 1). Tsiu Lake had sedimentation rates of 1,789 mm year^{-1} during 1992 (Fleisher et al., 1993) and 180 mm year^{-1} during

1993 after the surge had begun, associated with an increase in discharge as the surge began (Fleisher and Bailey, 1993; Muller et al., 1993; Figures 5 and 6; Table 1).

Gustavson and Boothroyd (1987) used the Malaspina Glacier as a modern analog for the Laurentide Ice Sheet in Lake Narragansett, Rhode Island. This study uses measured sedimentation records of depth and calibrated radiometric ages to determine a sedimentation rate.

ALPINE ICE-CONTACT, PROGLACIAL LAKE STUDIES

Syverson (1998) reports a sedimentation rate of 1,000 mm year^{-1} near the ice margin of the Burroughs Glacier, compared to less than 20 mm year^{-1} distally, based on stratigraphy (Table 1). Leonard and Reasoner (1999) used radiometric ages and depths of lake basins in Banff Park, Alberta, Canada (Hector, Crowfoot and Bow Lakes) to calculate sedimentation rates. All three lakes were characterized as evolving from an ice-contact, proglacial, lacustrine environment to a swamp environment, based on the increased organic content. The authors used bathymetric profiles and AMS dates of plant materials to calculate average sedimentation rates (Table 1). Reasoner and Rutter's (1988) study of Lake O'Hara reports sedimentation that dates from 10,100 and 8,530 years BP, with rates of 0.06 and 0.19 mm year^{-1}. The marked differences in sedimentation rates are attributed to shifts in sediment delivery to Lake O'Hara by the glacier (Table 1).

Where alpine glacial lakes mix with marine waters, primarily in fjords, saline-induced flocculation significantly enhances deposition of the suspended sediment load (Table 1; Pickrill and Irwin, 1983; Smith, 1990). Studies that measure suspended sediment concentrations from glacial discharge directly into the proglacial lake do not encounter potential errors in data due to mixing with marine waters (Leonard and Reasoner, 1998; Pickrill and Irwin, 1983; Gilbert and Shaw, 1981; and Smith, 1990). Sediment concentrations from discharge into proglacial lakes range from 18 mg/L (Leonard and Reasoner, 1998) to very turbid rates of 1,700-2,000 mg/L (Fleisher and Bailey, 1993). Increased rates of discharge into the lakes implies greater the rates of sedimentation, though very high discharge rates are likely to be affected by seasonal warming and increased ablation during warm periods.

Historic rates of sedimentation from ice-contact proglacial lakes associated with the Laurentide retreat are poorly understood, although many persist today (the Great Lakes, Glacial Lake Tonawanda; Gustavson and Boothroyd, 1987; Smith and Ashley, 1985). Gustavson and Boothroyd (1987) use the temperate glacial environment of the Malaspina Glacier, Alaska, to reconstruct the basin hydrology of Glacial Lake Narragansett along the Laurentide Ice Sheet in Rhode Island.

Sediments of Glacial Lake Tonawanda in western New York are primarily silt with rhythmically bedded fine sands based on data and observations presented herein (Figures 12 and 13). In shallow, ice-contact, proglacial environments such as Glacial Lake Chicago, which was approximately 20 m deep, Robey and Borucki (1995) suggest that the sediment deposits are gradational over a 90 m distance, where a diamicton grades laterally to varved mud and sand. This is consistent with coarse glacial outwash sediments that were deposited proximal to a retreating ice front as suspended sediments settled in distal portions of the lake (Robey and Borucki, 1995; Pickrill and Irwin, 1983; Leonard and Reasoner, 1999).

Figure 13. Percent sand, silt and clay.

Figure 13 shows the percent sand, silt and clay versus depth for cores 8, 11, 15, 16, and 17 in the Tonawanda Basin, western New York. Notice that silt and sand were deposited in concentrations nearly opposite from one another. Silt is the primary constituent of the cores through lacustrine and muckland deposits. Cores typically coarsen upward with decreasing clay content.

Figure 13A. Percent sand, silt and clay.

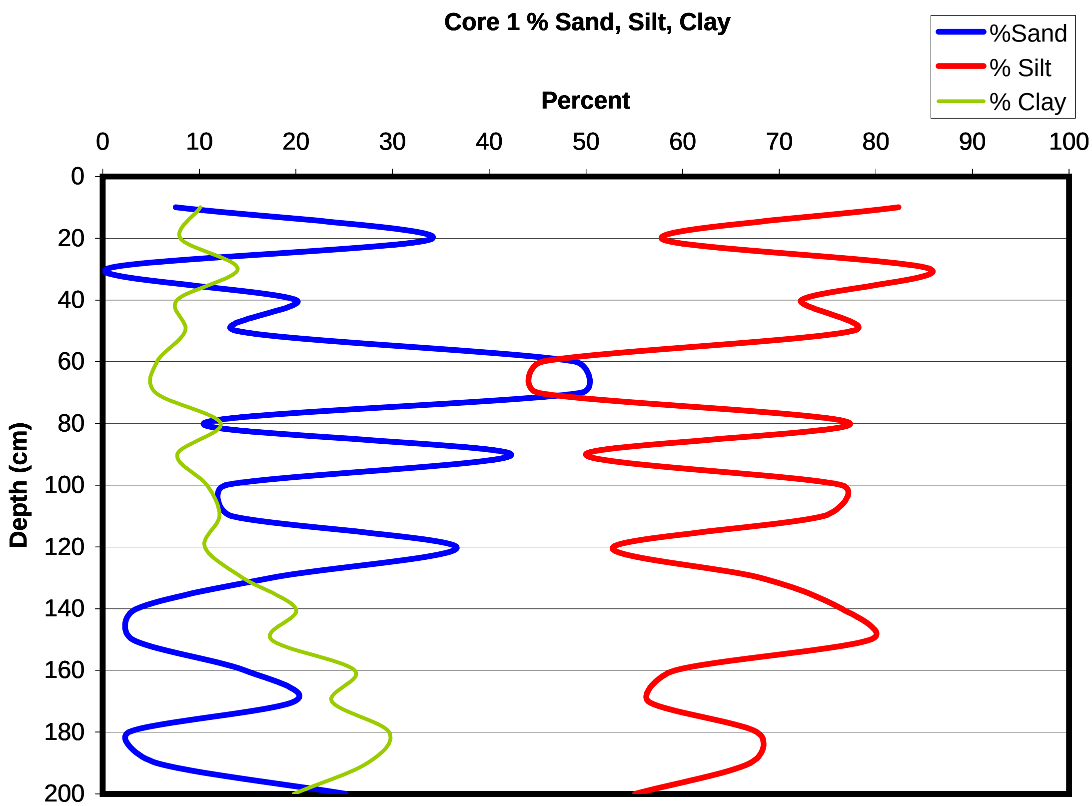
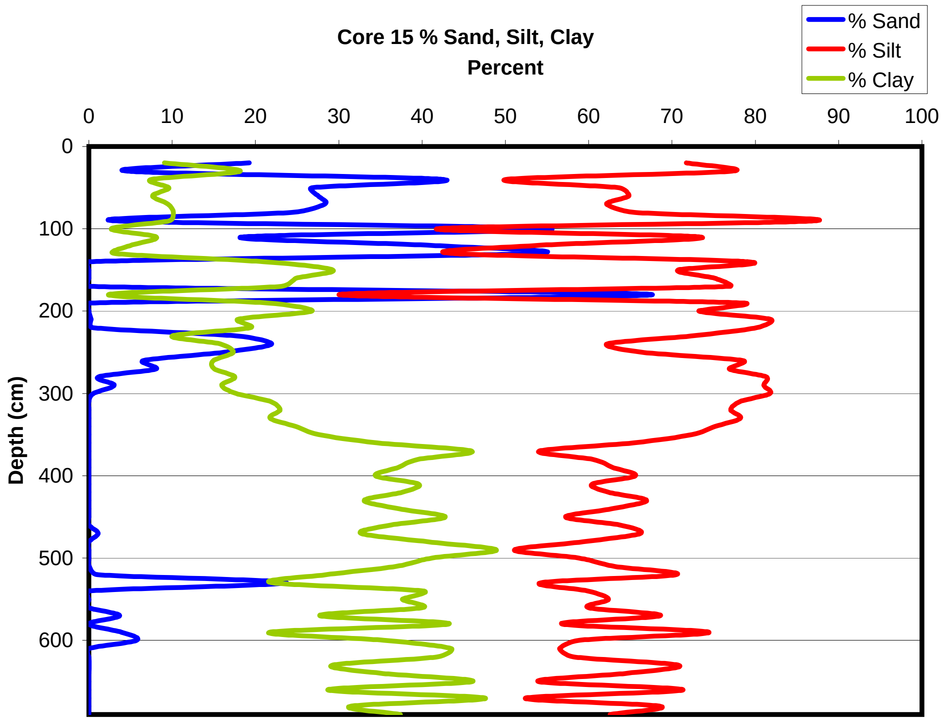


Figure 13B. Percent Sand, Silt and Clay.



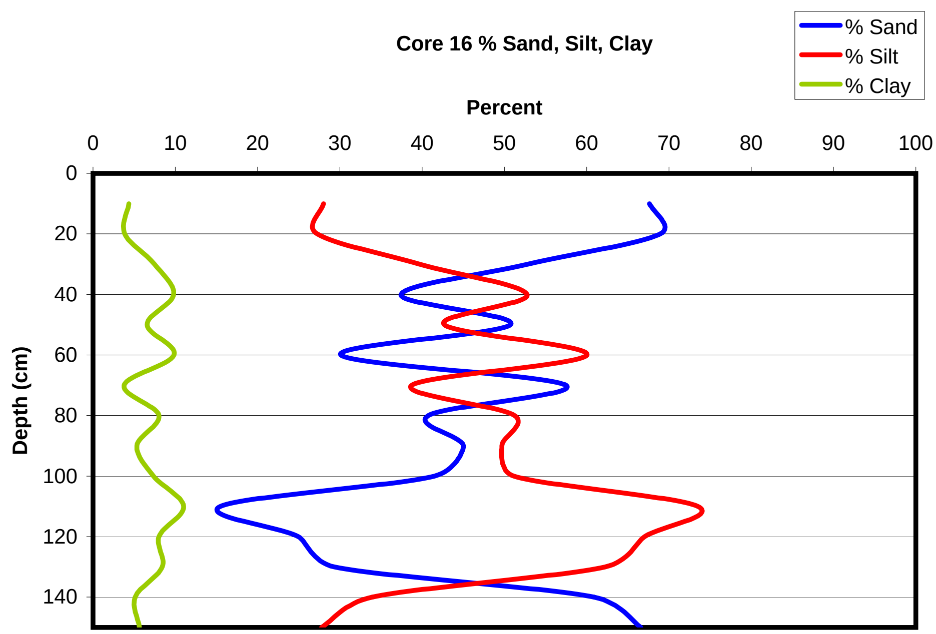
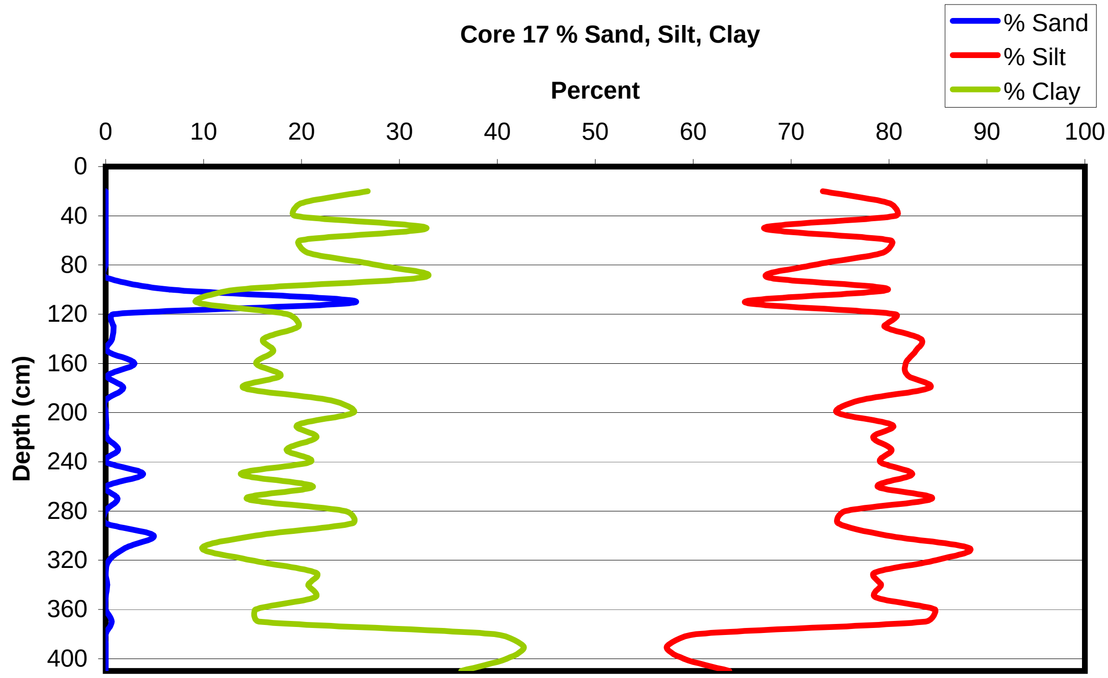


Figure 13C. Percent sand, silt and clay.



SUMMARY OF RATES OF LACUSTRINE SEDIMENTATION

Sedimentation rates in ice-contact, proglacial lakes suggest that lake morphology is not a controlling factor. Ablation rates coupled with variations in suspended sediment loads, substrate and glacier hydrology are important factors controlling sedimentation. Crystalline bedrock is significantly more resistant to abrasion and scour than sedimentary rocks or unconsolidated substrates. This is an important consideration when using any alpine glacier as an analog to the Laurentide Ice Sheet. Scale and morphology of the glacier, such as an alpine valley, piedmont or marine glacier, and their scale compared to ice sheet lobes are additional considerations when calculating sedimentation rates. Ice flow and ablation rates directly contribute to discharge and high suspended sediment loads. Lacustrine deposits in the Tonawanda Basin, New York, suggest that the Laurentide Ice Sheet was capable of reworking and depositing large quantities of sediment. While it is impossible to say how much was the result of scour and abrasion, it is likely that many sediments were simply reworked, transported and deposited in the Tonawanda Basin. A more detailed chronology is required to constrain sedimentation rates in the Tonawanda Basin for accurate comparison to the Bering Glacier (Table 1).

METHODS

FIELD SAMPLES

Cores from the Tonawanda Basin were collected using 1 and 4 cm gauge corers from the State University of New York, College at Brockport (courtesy of W. J. Autin). At each core site a GPS location, date, and elevation were recorded. Cores were 0.5-1 m in length and were collected in 10 cm intervals, and placed in plastic Ziplock bags. Samples were kept frozen to preserve their freshness. Field data for each 10 cm interval consisted of texture (peat, mucky peat, silty clay, clay, etc.), presence of carbonate (using dilute HCl), color (using the Munsell Soil Color Charts, 1994 edition), mottled pattern (presence and color), consistency (stickiness, plasticity), fibers present (type, quantity), soil structures, contacts, and the presence of any wood or shells. These data are located in Appendix F. Figures 4 and 5 show locations of cores collected during this study.

LABORATORY PROCEDURES

LOSS-ON-IGNITION

The ignition of organic and carbonate materials was performed using a modified procedure of Dean (1974) and Heiri et al. (2001). Air-dried samples were sieved to < 2 mm, air dried and weighed. Samples were placed in crucibles with lids off and heated in a furnace for two hours at 550ºC; stirring once during heating to ensure homogeneity of ignition. Samples are removed from the furnace, allowed to cool, and weighed. The difference in weight was organic matter lost. Samples were returned to an 850ºC furnace for two hours; again stirring during heating. Samples were then re-weighed to determine the amount of carbonate evolved. Following both heatings, samples were weighed as quickly as possible to prevent rehydration via uptake of humidity (Heiri et al., 2001). Several trial methods were used to develop this procedure (Natel, 2002; Natel and Autin, 2001). Lab facilities at the State University of New York, College at Oneonta, the State University of New York, College at Brockport and Arkansas State University were used.

Initially, the procedure of Dean (1974) was used for samples from Core 17 (Method A; Figure 14B). Samples were heated to 100ºC for one hour and weighed after cooling. Samples were then heated for one hour at 550ºC and weighed after cooling. This procedure failed to remove all organic matter from samples. Removal of all organic matter is necessary if treated sediment will be used for particle size analysis, as the presence of organics or ash will skew particle size results.

A second procedure (method B) involved heating samples for one hour at 100ºC and sixteen hours 350ºC (Nelson and Sommers, 1996). Nelson and Sommers (1996) recommend their procedure for soil samples containing > 1% organic matter. The length of time in the furnace, 16 hours in this case, had no consistent effect on organic removal. Cores 5, 6, 7, 8 and 15 were processed in this manner (Figure 14B). In all cases, data showed organic-rich sediments from surface horizons to depth, which was inconsistent with field observations. Where samples were organic-rich (> 20% organics), this procedure failed to remove all organics (Buol et al., 1997). This outcome resulted in a modification of the procedures outlined by Dean (1974) and Heiri (2001) for use with organic-rich soils typically found in mucklands.

Method C involved modifying procedures of Heiri (2001) and Dean (1974) by stirring samples mid-way during each heating. This allowed a homogeneous removal of organic carbon (at 500ºC) and carbonate (850ºC). Cores 1, 8, 16 and 17 were treated in this manner (Figure 14A). Total organic matter and carbonate were added together and subtracted from original sample weight to produce the total inorganic matter content in each core (see Appendix B). Data Tables are located in Appendix B.

Prior to the development and use of loss-on-ignition method C, carbonate was removed by digestion. Samples were digested in dilute glacial acetic acid (10-20%) to remove carbonate. Cores 1 and 15 were treated in this manner. Data tables are located in Appendix B (Figure 15). This procedure proved to be very time

Figure 14. Weight percent Organic carbon.

A. Graph of loss-on-ignition results for cores 1, 8, 15, 16, and 17. Weight percent organic carbon versus depth for each core, and based on the modified procedures of Dean (1974) and Heiri (2001). All cores show a general trend of less than 10% organic carbon. Increases in organic carbon content are consistent with stratigraphic intervals of peat.

B. Shows graphs of loss-on-ignition data for cores 1, 5, 6, 7, 8, 16, and 17 from LOI Trials A and B. These data were deemed inaccurate because high organic carbon content did not coincide with field observations of stratigraphy.

FIGURE 14A. Wt% ORGANIC CARBON.

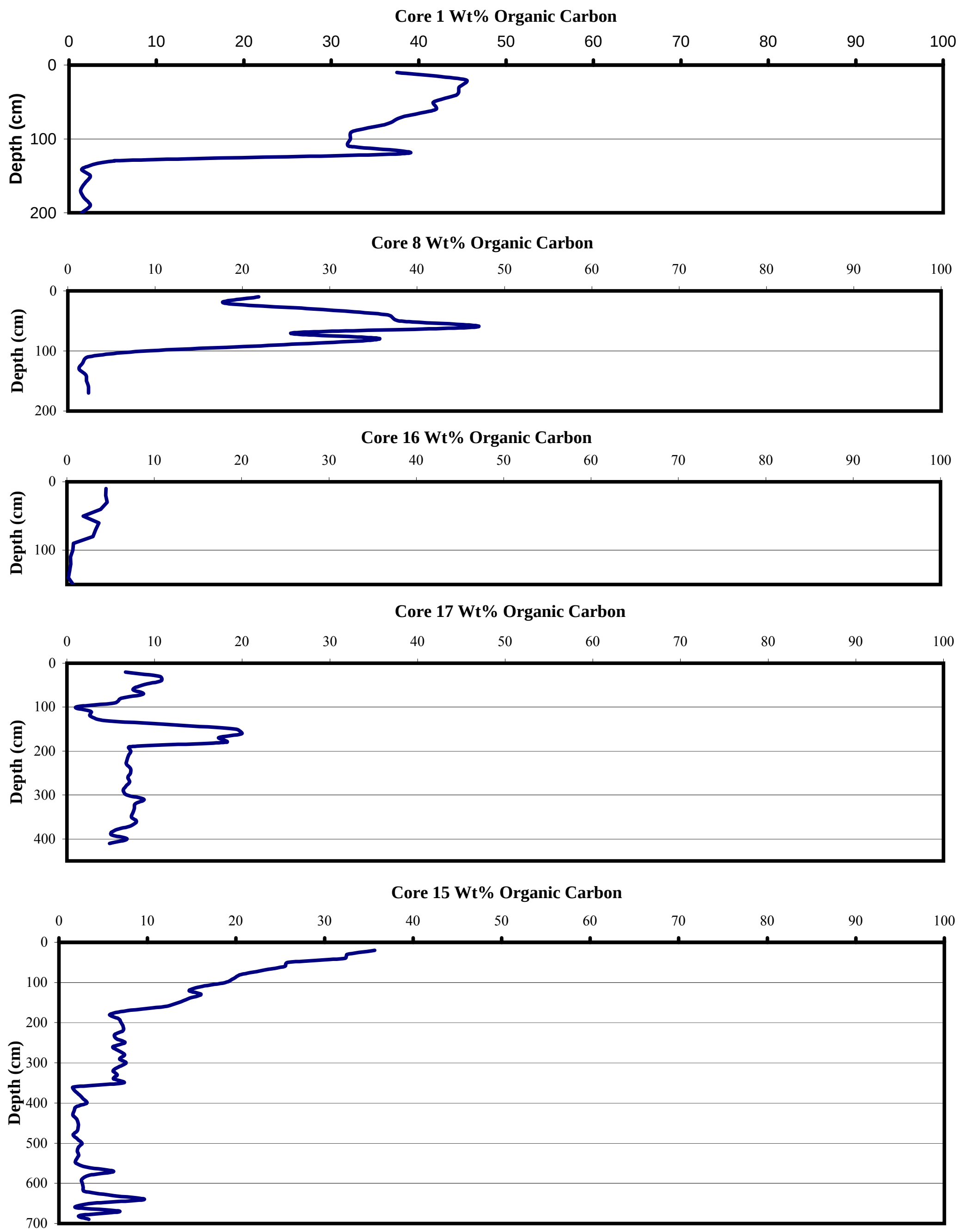


FIGURE 14B. Wt% ORGANIC CARBON.

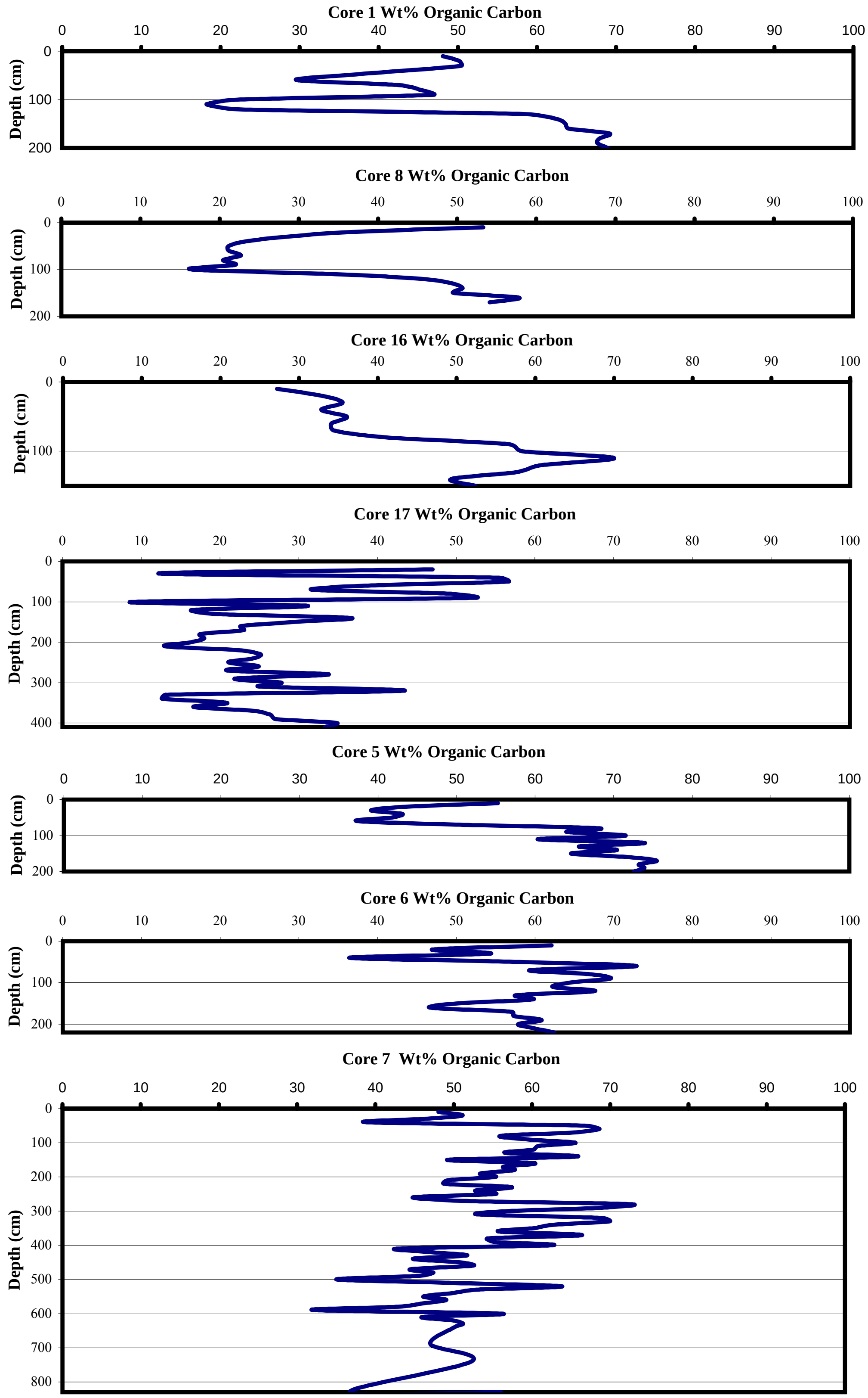
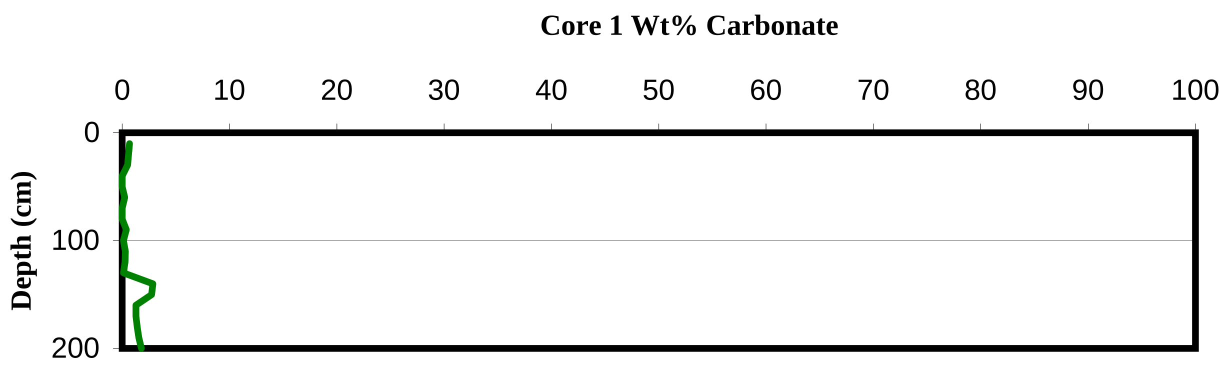


Figure 15. Weight percent Carbonate.

A. Graphs show results of carbonate removal from cores 1 and 15 by digestion in dilute CH_3COOH (glacial acetic acid).

B. Graph shows results of carbonate removal from cores 1, 8, 16 and 17 using loss-on-ignition procedures (ignition at 850ºC). This procedure was performed on samples following heating of samples for two hours at 550ºC. Core 1 was analyzed using both digestion and ignition methods for comparison purposes. See Appendix B for comparison graph of all carbonate data.

Figure 15A. Wt% Carbonate (Method A: Digestion).



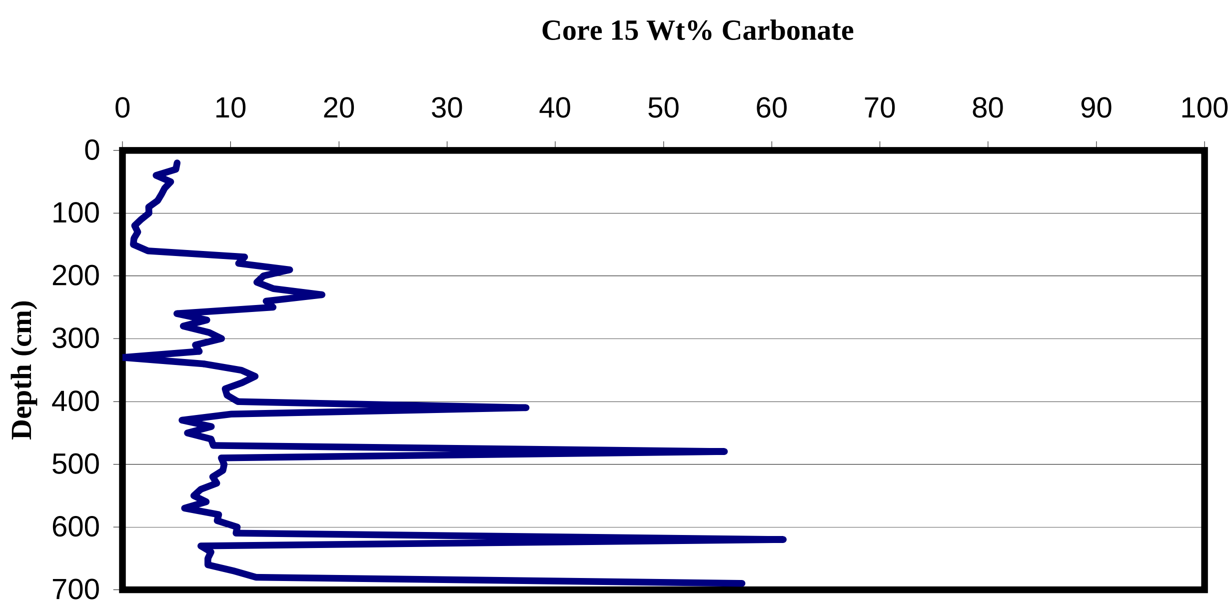
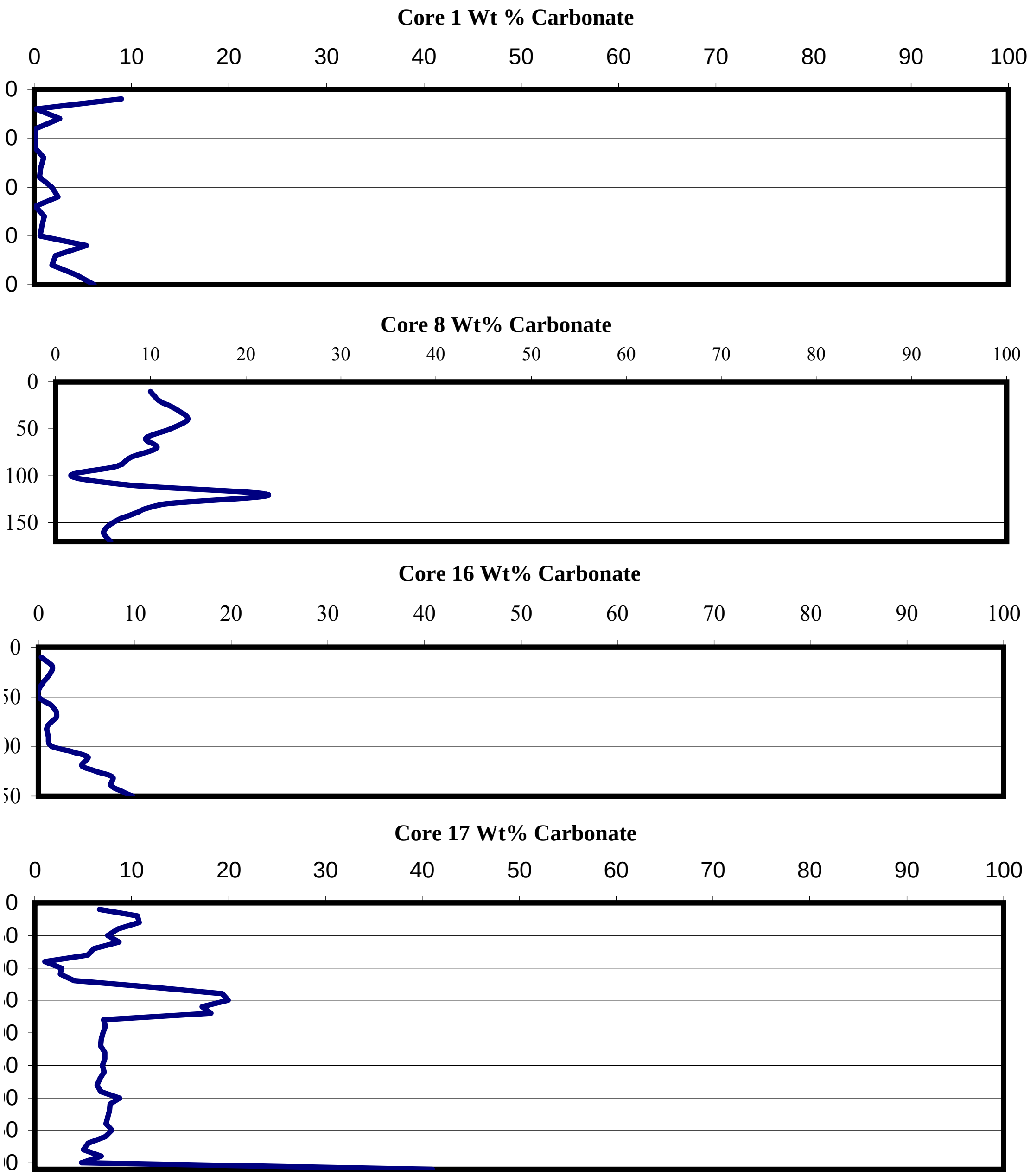


Figure 15B. Wt% Carbonate (Method B: Ignition).



consuming and costly, using tremendous amounts of glacial acetic acid. Weight percent carbonate data is presented in Figure 15 (Appendix B).

PARTICLE SIZE ANALYSIS

Samples for particle size analysis were treated by removing organic matter and carbonate content (see loss-on-ignition and carbonate removal procedures; Figures 14 and 15). Particle size analysis was performed using a Coulter-Beckman LS230 Particle Counter in the Earth Sciences Department at the SUNY College at Brockport (courtesy of W. J. Autin). The LS230 analyzer measures particle sizes ranging from 0.04 to 2000 µm. A solution of 65 g/L of Calgon in de-ionized water was added to each sample during loading (40 mL per run) to prevent flocculation of particles. Prior to each analysis, samples were disaggregated using a mortar and pestel to disaggregate particles. Samples were sonicated during loading and during the 90 second run to further reduce flocculation. Cores 1, 8, 15, 16, and 17 were analyzed for particle size (data tables are located in Appendix C). Each data point represents a 10 cm interval. Percent sand, silt and clay were calculated using the LS230 program (LS.exe) and graphed (Figure 13). Original data from the particle size analysis can be found on the CD in pocket (Binder Graph ~.pdf). Graphs of mean grain size for each core and a bivariate diagram showing the degree of sorting were plotted (Figures 16 and 17).

MAGNETIC SUSCEPTIBILITY

Mass and frequency dependent magnetic susceptibility were measured on weighed, air dried samples that were disaggregated and sieved using a < 2 mm sieve. Magnetic susceptibility is a convenient way to measure the relative ferrimagnetic minerals in a bulk sample (Lecoanet et al., 1999). A Bartington MS2B probe was used at

Figure 16. Average (arithmetic mean) grain-size.

Figure 16 shows mean grain size (µm) distribution from cores 1, 8, 15, 16 and 17. Core 16 has an anomalously coarse-grained basal interval above the till layer.

Figure 16. Mean Grain Size (µm)

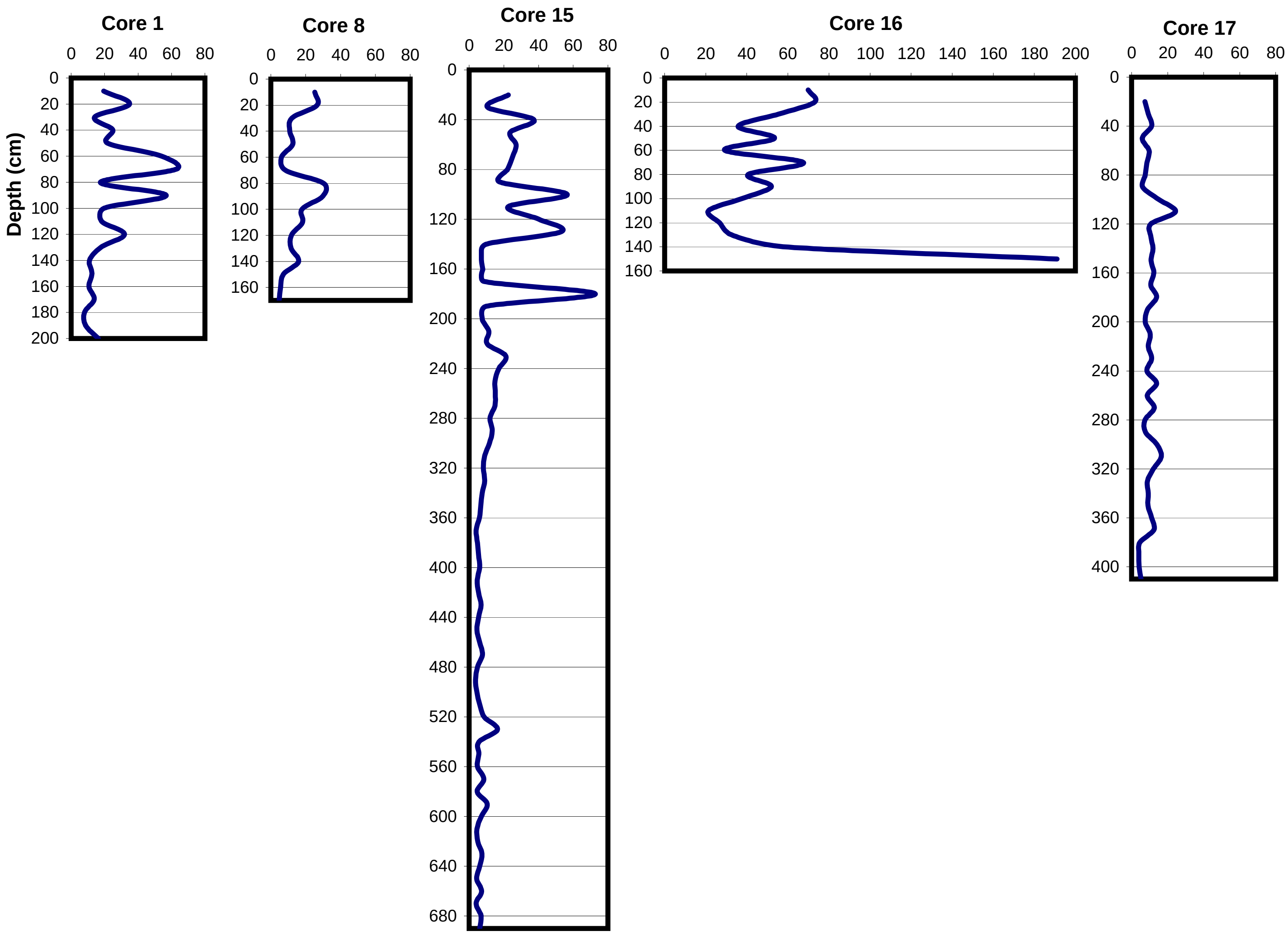
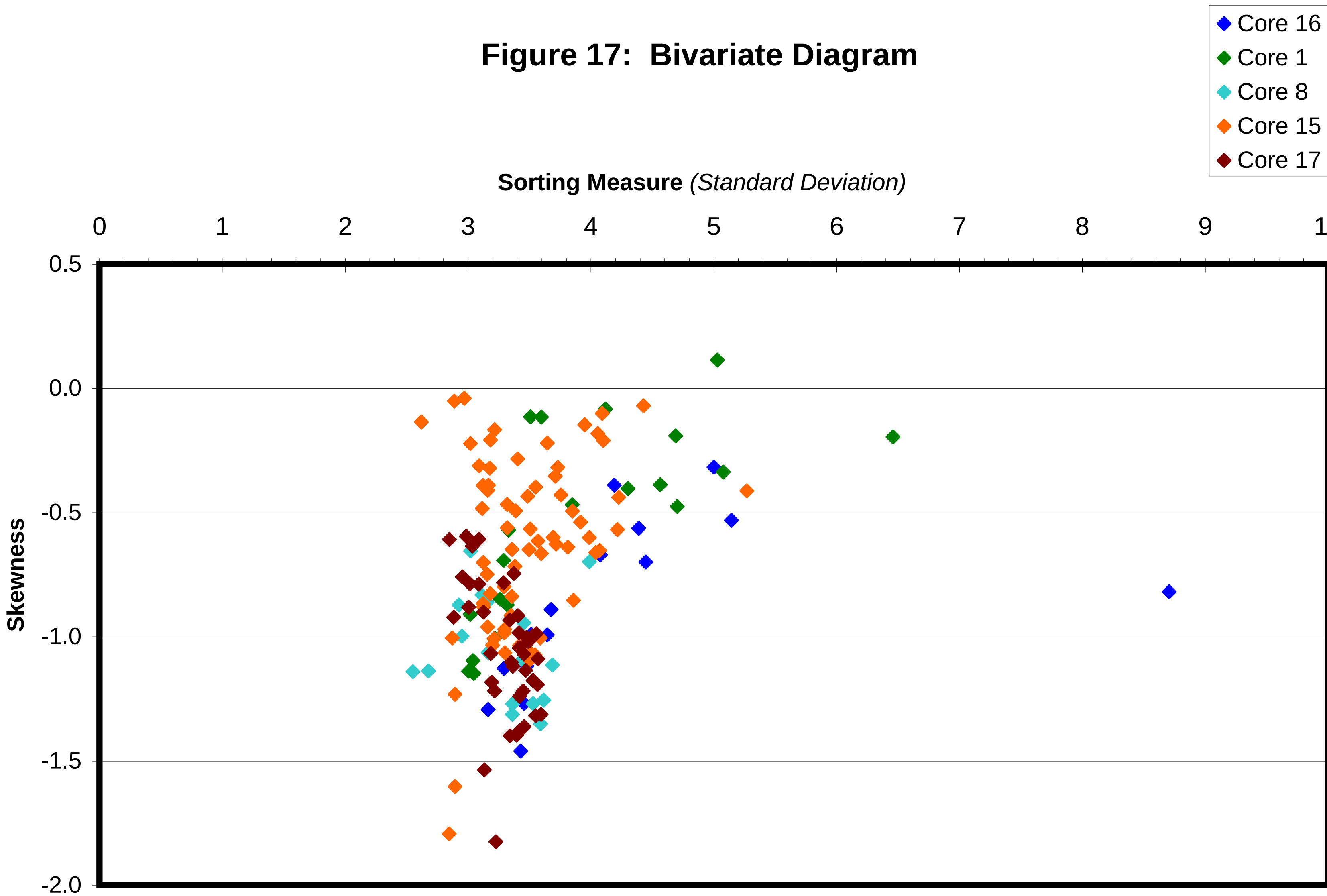


Figure 17. Bivariate diagram.

Figure 17 shows the standard deviation of each 10 cm depth interval with the skewness of each grain size distribution to produce a measure of sorting. Most samples can be characterized as left skewed (mostly fine sediments) and this is shown by the clustering of data between 3 and 4 on the Bivariate diagram. Cores 1, 8, 15, 16 and 17 are included in this graph. See Appendix C and the attached data CD for particle size data.

Figure 17: Bivariate Diagram



the SUNY College at Brockport (courtesy of W. J. Autin). Twenty gram samples were air dried and placed in tarred clean and clear film canisters. Samples were analyzed for high frequency (HF) and low frequency (LF) susceptibility using SI units. Output from the Bartington program is located in Appendix D for cores 1, 5, 7, 16, and 17. This procedure allows for mapping of relative concentrations of magnetic minerals within a profile (Figure 18).

Variations within a profile can be used to infer relative age of landscape as a function of texture and composition derived from depositional process and provenance. Increases in magnetic susceptibility upward through a soil/sediment profile are thought to be the result of pedogenic processes associated with *insitu* formation of fine-grained ferromagnetic minerals (Grimley, 1995). Low frequency magnetic susceptibility values were plotted independently versus depth for each core (Figure 18).

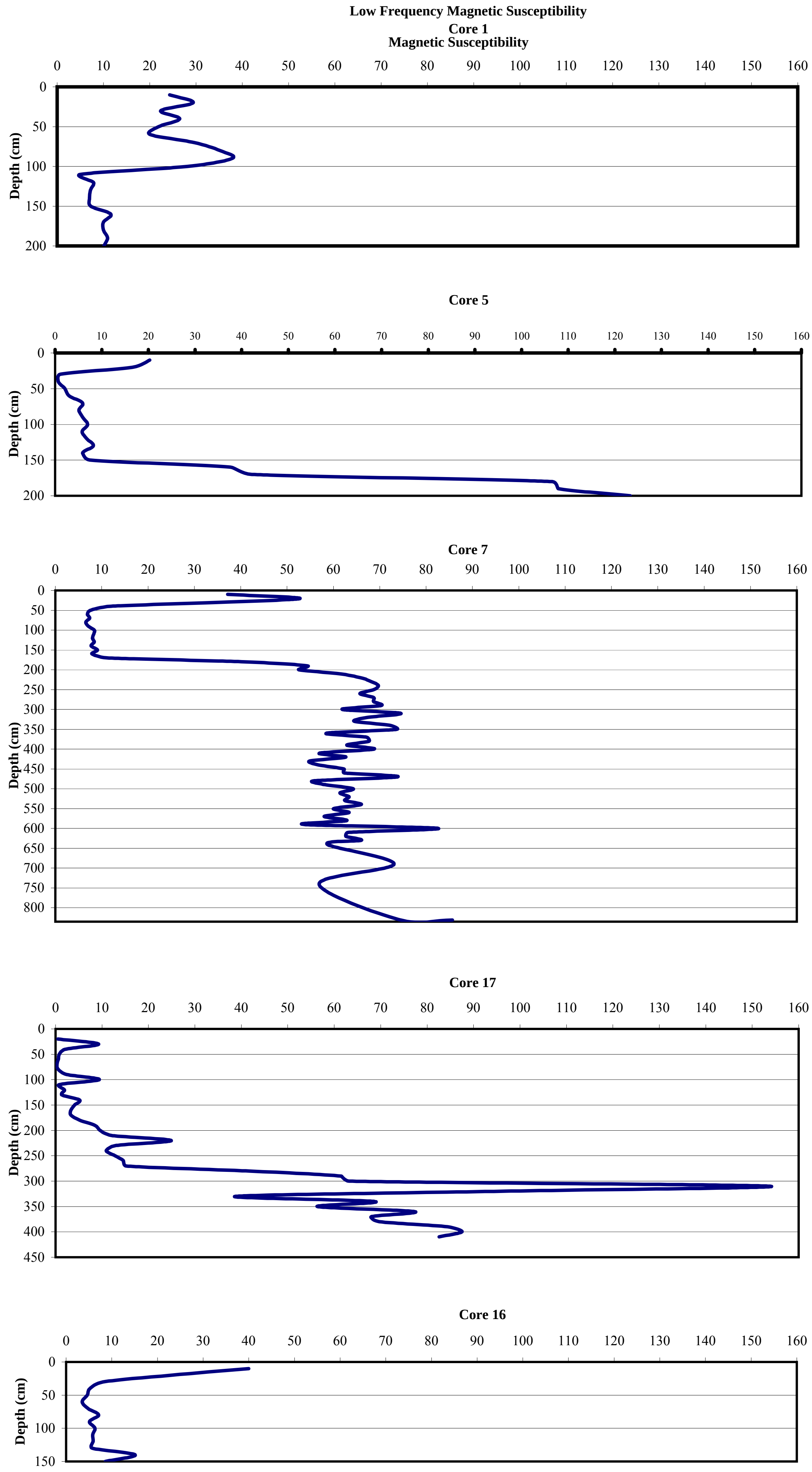
MACROFOSSIL ANALYSIS

The procedure involved filling a graduated cylinder with 100 mL of water and adding sample to obtain 50 cm^3 of sample by means of displacement (Yansa and Basinger, 1999). Samples were then poured over a series of three sieves (1 mm or 500 µm, 250 or 212 µm and 63 µm) in a sink. Different sized sieves were used to sort materials to facilitate collection of macrofossils. Water was allowed to wash over samples until fines were removed. Contents of each sieve were collected using de-ionized water and placed in specimen bottles. Using a 20x microscope, petri dishes, insect forceps, very fine paint brushes, and additional de-ionized water, mollusks and seeds were collected. Once all three portions of a particular interval were picked,

Figure 18. Magnetic susceptibility.

Figure 18 shows low frequency magnetic susceptibility values versus depth (cm) for cores 1, 5, 7, 16 and 17. These trend lines show effects of weathering, with more heavily weathered intervals having a low magnetic susceptibility. Magnetic susceptibility measurements are in SI units and are weight based (see Appendix D for raw data).

Figure 18. Magnetic Susceptibility.



mollusca were sorted by species, where possible, to genus level. Species were identified based on morphology descriptions in literature sources (see Appendix A and Plates 10 and 11 for images used for identification purposes). Representative specimens of different species were photographed using a 35 mm camera mounted on a microscope (Appendix A, Plates 1-7). Additional specimens were photographed using a scanning electron microscope in the Biology Department, Arkansas State University (courtesy of S. E. Trauth; Appendix A, Plates 8 and 9). A complete taxonomy of mollusk species from the Holley Embayment of the Tonawanda Basin was constructed (Figure 19). Mollusks were collected in modern glacial environments (Bering and Matanuska Glaciers) to compare species with those identified in Tonawanda sediments and to verify habitats.

Western New York

Core 15 was analyzed for macrofossils. Results of this analysis can be found in Appendix A. Once species were identified, species were sorted by niche (riverine, lacustrine, pond and near-shore species; Figure 20). The total number of mollusks by depth interval was plotted as well as the number of each species by depth (Figure 21). Determination of the appearance and disappearance of specific species, depending on their particular niche, is useful for water depth interpretations in the Tonawanda Basin (see results and discussion herein).

Figure 19. Mollusk taxonomy.

Figure 19 shows the taxonomy of mollusks (gastropods and bivalves) identified in core samples from Glacial Lake Tonawanda. Species identification was necessary to determine ecological niche (e.g., lacustrine (swimmers); pond (attach to branches and debris at waters edge); riverine (living on branches and sediment in rivers and creeks); and terrestrial (forests and creek or river banks)).

Figure 19. Mollusk Taxonomy.

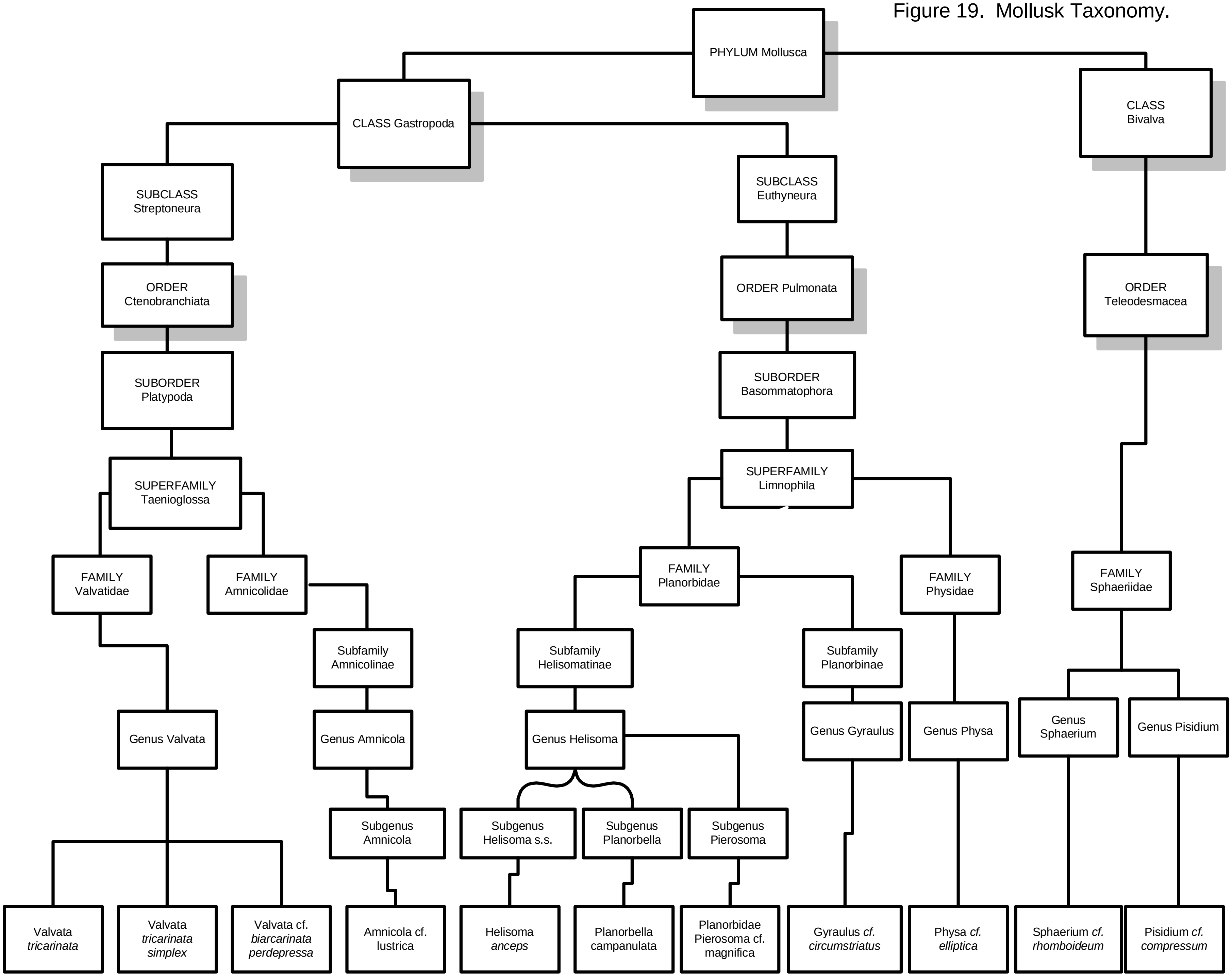


Figure 20. Mollusk ecology, Tonawanda Basin, western New York.

Figure 20 is a graph of mollusk species from core 15. Species are divided into three groups: Riverine, pond and near shore, and lacustrine species. Note that all species disappear at 150 cm of depth, indicating a shift to unfavorable water conditions. Riverine and pond species do not appear in the 690 cm core until 400 cm of depth. The presence of riverine species indicates their deposition as detrital debris. Lacustrine species are the first to appear in the core; bivalves were found in the basal part of the core at 690 cm depth.

Figure 20: Mollusk Ecology of Western New York, Tonawanda Basin.

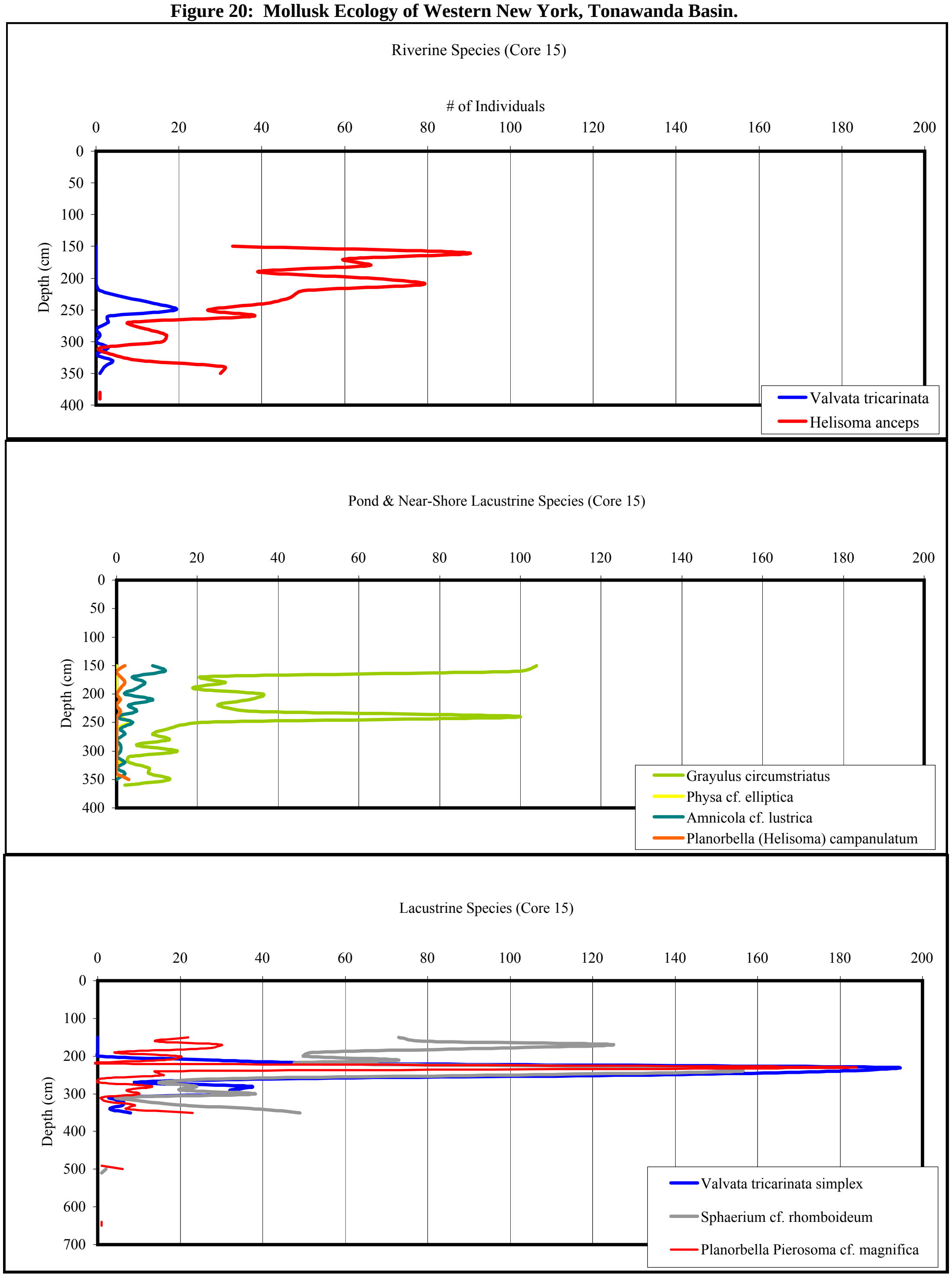


Figure 21. Total mollusks.

Figure 21 shows the total number of mollusks collected in core 15, and the total number of each species identified in core 15 by depth.

FIGURE 21: TOTAL MOLLUSKS BY DEPTH (CORE 15), TONAWANDA BASIN.

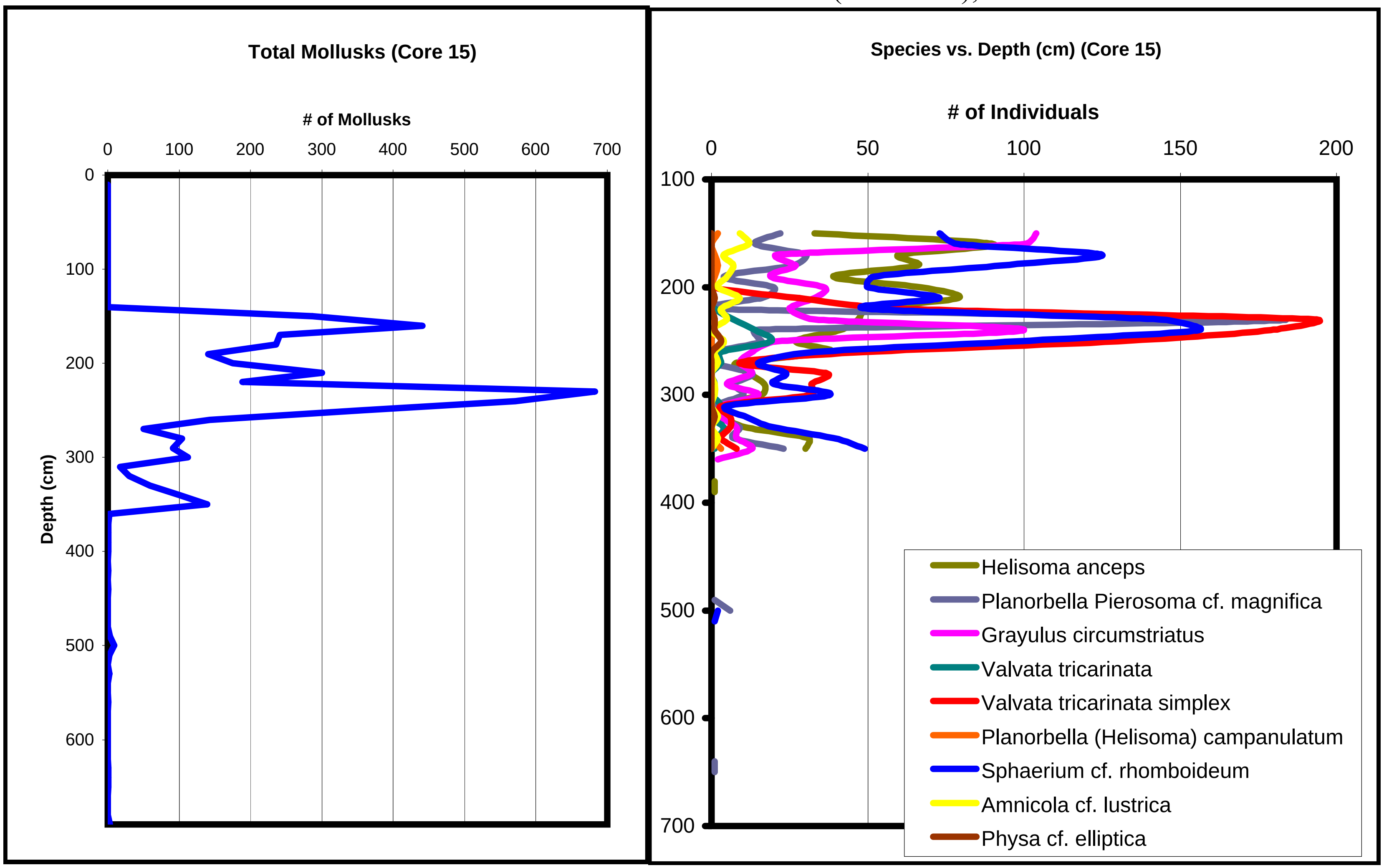


TABLE 2. MOLLUSK SPECIES FROM THE PROGLACIAL REGIONS OF THE BERING AND MATANUSKA GLACIERS, ALASKA.

Bering Glacier	**Species**	**Counts**	**Habitat**
Umlaufland	*Fossiara parva* *Succinea cf. grosvenori*	10 4	Lives in marshy places on debris. Terrestrial; able to withstand severe cold and short summer seasons. Found in shallow water on muddy bottom.
Temp. (ºF)/ pH	49 / 6.71 (Upper pond) 48 / 6.68 (Lower pond)		
Matanuska	**Glacier**		
Ice-contact, proglacial Lake	*Fossiara obrussa decampi* *Succinea cf. grosvenori* *Valvata sincera*	4 51 18	Mud flat species. See description above. Lacustrine; deep cold water.
Ice-contact drainage (debris band)	*Succinea cf. grosvenori*	4	See description above.

Table 2. Mollusca from the Bering Glacier and the Matanuska Glacier, Summer 2003. Mollusk specimens at the Bering Glacier were living at time of collection in a shallow pool (see Figures 7 and 9). The ponds have poor drainage and are fed by meteoric water. Matanuska Glacier: a) UTM coordinates are 0458485, 6850803, elevation @ 475 m asl, dried ice-contact, proglacial lake basin; specimen shells only; b) UTM coordinates are 0461241, 6850536; specimens were living at time of collection in a shallow stream draining from an ice-cored debris band (see Figure 11).

<u>Bering Glacier, Alaska</u>

During the 2003 field season at the Bering Glacier, Weeping Peat Island, Alaska, living mollusks were collected in a shallow, ephemeral pond (Figures 10 and 11; 2003 sampling site labeled with green arrow on Figure 7; data are presented in Table 2).

<u>Matanuska Glacier, Alaska</u>

During the 2003 field season at the Matanuska Glacier, Alaska, mollusca were collected from a dried ice-contact, proglacial lake basin and from drainage from ice-cored terrain (Figure 10). Living specimens were collected from a stream in ice-cored terrain. The ice-contact, proglacial lake has been drained for several years (S. Goetz, personal communication, 2003). Figure 11 shows the dry ice-contact, proglacial lake basin and the ice-contact drainage way sites. Species identified are reported in Table 2.

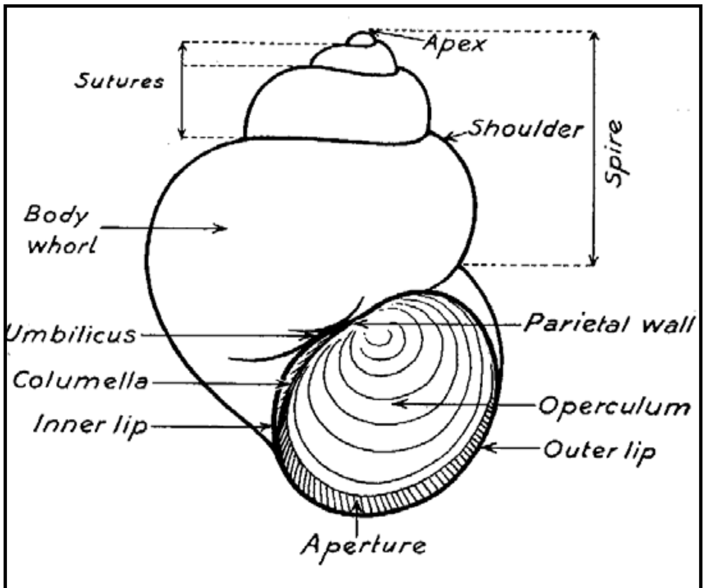
Mollusk Criteria for Identification

Shells of gastropods and bivalves were identified based on their morphology (Figure 22). Key features used in identification are:

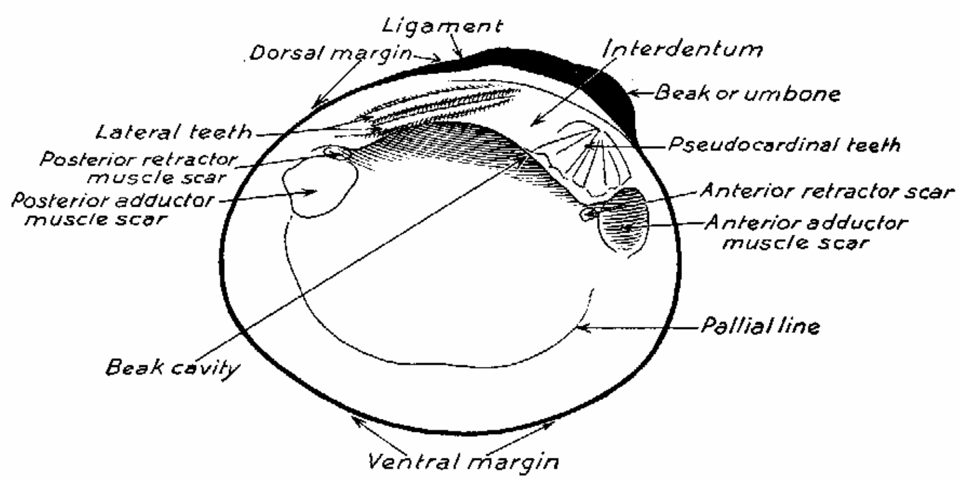
- General shape
  - Simple conical
  - Flat and whorls elevated into spire
  - Presence or absence of keels (ridges on whorls)
- Shape and orientation of aperture
  - Dextral (right-sided)
  - Sinistral (left-sided)
  - Varicose (thickening of outer lip)
- Number of whorls
  - Count the number of complete whorls starting with the apex
- Symmetry
  - Whether or not the mollusk shell is symmetrical top/bottom or side/side.
- Shape of umbilicus
  - Deep
  - Shallow
- Sutures

Figure 22. Mollusk Morphology.

From Burch (1962): A, Gastropod morphology; B, Bivalve morphology (diagrams are not to scale). Used for examples of terminology and in shell identification.



Parts of gastropod shell



Parts of pelecypod shell

    - Prominent
    - Less prominent (smooth surface)
- Shells photographed and compared with photographs, drawings and descriptions in literature sources (Appendix A, Plates 1-11).

Specific species identification is necessary for determination of ecological niche. Modern species of mollusca are typically identified based on their soft-parts. For mollusca represented by only shells in sediment cores, care must be given to allow for natural variation within a species. The total numbers of mollusks in sediment cores is typically used as an indicator of lake level change (Mullins, 1998; Karrow and Mackie, 2001). Species numbers, their disappearance and appearance can signify changes in paleohydrology, biodiversity, water shallowing and subsequent warming of shallower water bodies. It is necessary to separate out species that are transported into a lacustrine environment, such as riverine and other terrestrial species, so that mollusk numbers are not inflated.

X-RAY DIFFRACTION (XRD) ANALYSIS

X-ray diffraction of representative mollusk shells was performed to determine the chemical composition of shell materials. The XRD in the Earth Sciences Department, SUNY College at Oneonta, was used (courtesy of D. Wohlford). Shell materials of a single species were crushed using an agate mortar and pestel in acetone. Powder was placed on a slide and placed in the XRD. Resulting graphs were analyzed for peaks corresponding to those of specific minerals (e.g., calcite, aragonite, etc.).

ELEMENTAL GEOCHEMICAL ANALYSIS BY ICP-MS

Sediment samples from cores 1 and 16 were analyzed using an Inductively Coupled Plasma Mass Spectrometer (ICP-MS) in the Chemistry and Physics Department, Arkansas State University, Water-Rock-Life Laboratory (courtesy of R. E. Hannigan). The results of these analyses are found in Appendix E.

Approximately 0.05 g of each sample, representing a 10 cm interval, was treated to remove everything except silicate particulates using ASTM procedure 6357-00a and that of Hannigan (2001). Using prepared single element standards (1000 ppm), samples were run using a Multi-Collector ICP-MS (ASTM procedure 6357-00a; Hannigan, 2001). Results of ICP-MS analyses from Cores 1 and 16 are found in Appendix E. Data were normalized to upper continental crust (sample/UCC) using values posted on the GERM website (www.earthref.org/GERM). Rare earth element data can be found in Appendix E and Figure 28 graphs La/Sm to determine the source rock of the sediment. REE data were normalized to upper continental crust values (UCC) for the purposes of comparison (Taylor and McClennen, 1985; Boyle, 2001; Eusterhues et al., 2002). La/Sm ratios greater than '1' indicate a granitic source (enriched in LREE) rock and less than '1' a carbonate source rock (depleted in LREE; R.E. Hannigan, personal communication, 2004).

DATA AND RESULTS

SEDIMENTARY ANALYSES

Sediments in Glacial Lake Tonawanda are primarily silt sized particles (50-90%; Figure 13). Clay content in cores 1, 8 and 15 begins high (30-50%) and diminishes towards the surface as organic content increases (Figures 13 and 14A). Sand content in all cores appears to be more episodic in nature, as pulses of sand deposition are absent in places, and are subsequently followed by increased input of silt and clay into the Holley Embayment (Figure 13). This suggests pulses of sediment input are due to drainage from an adjacent ice front along the Barre Moraine. More specifically, it may also reflect seasonal melting or warm summer periods of increased melting. Esker preservation along the Barre moraine may suggest regional ice stagnation between the Lockport Escarpment, the Clarendon-Linden fault scarp and the Barre Moraine (Figure 1; Brennand, 2000). Stagnant ice in this region is supported by the presence of eskers, which would have been otherwise reworked if the ice were still actively retreating. This suggests that the ice on the escarpment became dynamically detached from the ice sheet as the ice sheet retreated. As ice thinned on top of the Lockport Escarpment, ice in the Ontario basin was actively retreating (Figures 1A and 13).

Most cores, except for core 16, show rather consistent thickness of silt deposition, despite the differing core depths (depths range from 2 m in Core 1 to 6.9 m in core 15; Figures 12 and 18). Sand accumulation reaches a maximum in core 16, during peat formation in the Holley Embayment. Increased sand deposition implies a more energetic depositional environment, compared to that necessary for settling of finer suspended sediment loads, and may be in part due to the close proximity to the Batavia Moraine and the esker. Sand deposition may simply reflect runoff associated with meteoric water. However, other possibilities include renewed drainage of meltwater into the basin from adjacent esker sources. Another possibility is that the Niagara River re-occupied the Tonawanda Basin, but this is unlikely as the sand is consistent with sand from esker deposits adjacent to the Barre Moraine (Figures 1 and 12; Kindle and Taylor, 1913).

Organic content varied from less than 10% to greater than 70%. Organic content below a weight percent of 10% signified lacustrine sedimentation and is associated with silt, clay and marl deposits (Figures 12 and 14A). This can be seen in Cores 1, 8, 15 and 17 (Figure 14A), as increases in organics signify peat deposition and a muckland environment. Data suggest an overall trend of lacustrine sedimentation that shifts abruptly to bog formation associated with deposition of organic matter (10-70% by weight).

Carbonate content typically ranged from 0-10%, with marls and shell-rich horizons containing greater than 20% carbonate by weight. Core 15 is very fossiliferous, and consequently there are several zones where carbonate content increases to above 60%. Although some carbonate content represents groundwater deposition of calcite and the paleohydrologic conditions. Biological precipitation of carbonate from flora and fauna are significant contributors of carbonate (Yansa and Basinger, 1999). *Chara* was found in core samples and is a freshwater alga (Yansa and Basinger, 1999). *Chara*, or Stonewort algae, is 0.9-1.1 mm long and 1.0-1.3 mm in diameter (Yansa and Basinger, 1999). As *Chara* grows it removes CO_2 from the water precipitating an encrusting soft $CaCO_3$ shell, which eventually causes the algae to sink to the bottom where they decompose leaving their carbonate shells on the lake bed (Hasser, 1954). Their presence in lake sediments suggests that an active biological community existed in the photic zone during later pond stages of Glacial Lake Tonawanda (Mullins, 1998; Yansa and Basinger, 1999). The presence of *Chara* and their consumptive use of CO_2 in the presence of carbonate-rich water, allows for the precipitation of marl which is indicative of quiet water deposition of carbonate mud with little to no terrigenous input, suggesting that the ice front had retreated north of the Lockport Escarpment, where its drainage remained in the Ontario Basin (Karrow and Mackie, 2001).

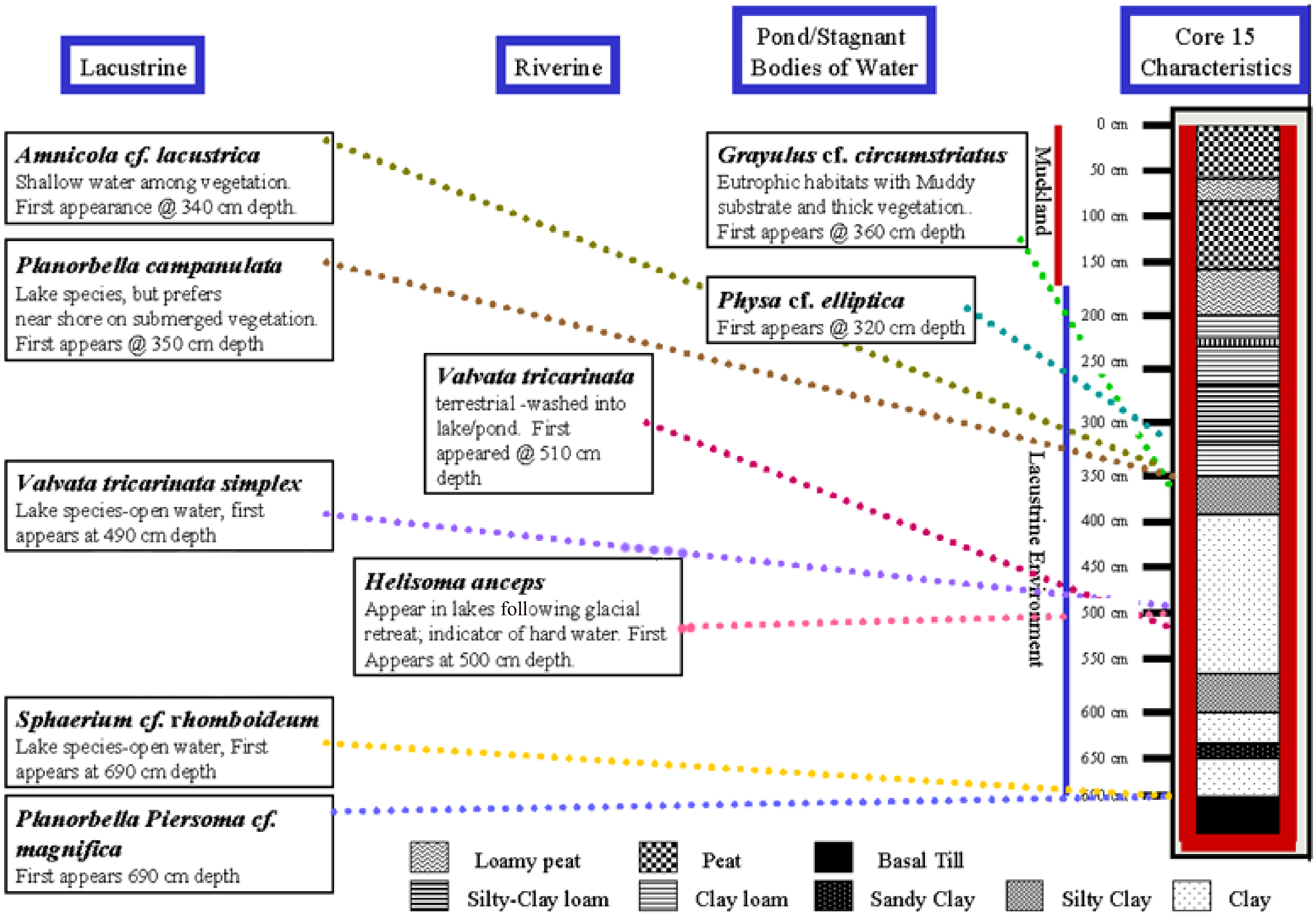
Mollusk shells provide an additional source of carbonate (aragonite) in sediment samples based on XRD analyses. Aragonite is a less stable form of carbonate under ambient pressure conditions (1 atm), making it more soluble in surface waters (Drever, 1997). Shells were readily visible at the coring site of core 15 (Figure 3). Elsewhere in the Holley Embayment, shells were much less common, both at the surface and at depth. Mollusk shells, such as *Helisoma anceps*, are known indicators of carbonate-rich paleohydrologic conditions. *Helisoma anceps* is represented in core 15 from 150-400 cm, indicating a carbonate-rich water environment and associated with marl stratigraphy (Figures 12, 20, 21 and 23; LaRocque, 1968). LaRocque (1968) suggests that *Helisoma anceps* is a pioneer species preferring carbonate-rich, cold, freshwater lakes. However, since mollusks in core 15 do not appear until 400 cm depth, this suggests that lake waters were not suitable to support mollusks prior to deposition of these sediments due to low carbonate content (< 10%, Figures 15 and 16), a high suspended sediment load, and turbid water associated with lacustrine sedimentation in an ice-contact, proglacial environment (Figure 13; Ashley, 1987).

Magnetic susceptibility (MS) allows reconstruction of past environmental conditions (Thompson and Oldfield, 1986). Magnetic susceptibility of sediment cores from the Holley Embayment, suggests neoformation of ferromagnetics in the sediments. Spikes in low frequency (X_{lf}) magnetic susceptibility indicate the accumulation of

Figure 23. Ecological significance of mollusk species.

Figure 23 shows the stratigraphy of core 15 representing two environmental deposition associated with a muckland and lacustrine sedimentation. The upper portions of the lacustrine deposition are characterized by marl deposits (200-350 cm). The muckland deposits are primarily organic-rich peat deposits (0-150 cm). Shown are species classified by habitat (lacustrine, riverine, pond species) and their first appearance in the stratigraphic column.

FIGURE 23. ECOLOGICAL SIGNIFICANCE OF MOLLUSK SPECIES, TONAWANDA BASIN.



magnetic minerals (Figure 18). The linear relationship between X_{lf} and X_{hf} suggests that X_{fd} is insignificant, as X_{fd} is the ratio between X_{lf} and X_{hf}, and represents variations in magnetic susceptibility as a function of depositional process (Autin, personal communication, 2004; Figure 24 and Appendix D). This is consistent with well sorted, predominantly silt-sized particles, entering imponded water (Figure 19).

Grimley (1995) and Thompson and Oldfield (1986) suggest that spikes in magnetic susceptibility represent neoformation of ferromagnetics (< 12,000 years old). MS is sensitive to pedogenic ferromagnetic minerals in the 0.018-0.020 mm range and indicates relative age based on oxidation of parent material (Grimley, 1995; Autin, personal communication, 2004). For a particle size distribution that can be represented by a bell-shaped curve, ferromagnetic minerals are typically represented by silt-sized particles, assuming the mode grain size is medium sand.

Particle size distributions for cores from Glacial Lake Tonawanda are predominantly left skewed, with silt-sized particles the coarsest fraction (Figure 12; Appendix C). This suggests that ferromagnetic minerals are not the result of *insitu* formation of ferromagnetics, but rather they are from depositional processes (energy of transport). Similarly, when X_{lf} and X_{hf} are equal, not only is X_{fd} insignificant, but any magnetic signal is most likely the result of variation in the depositional processes controlling sediment sorting (Autin, personal communication, 2004).

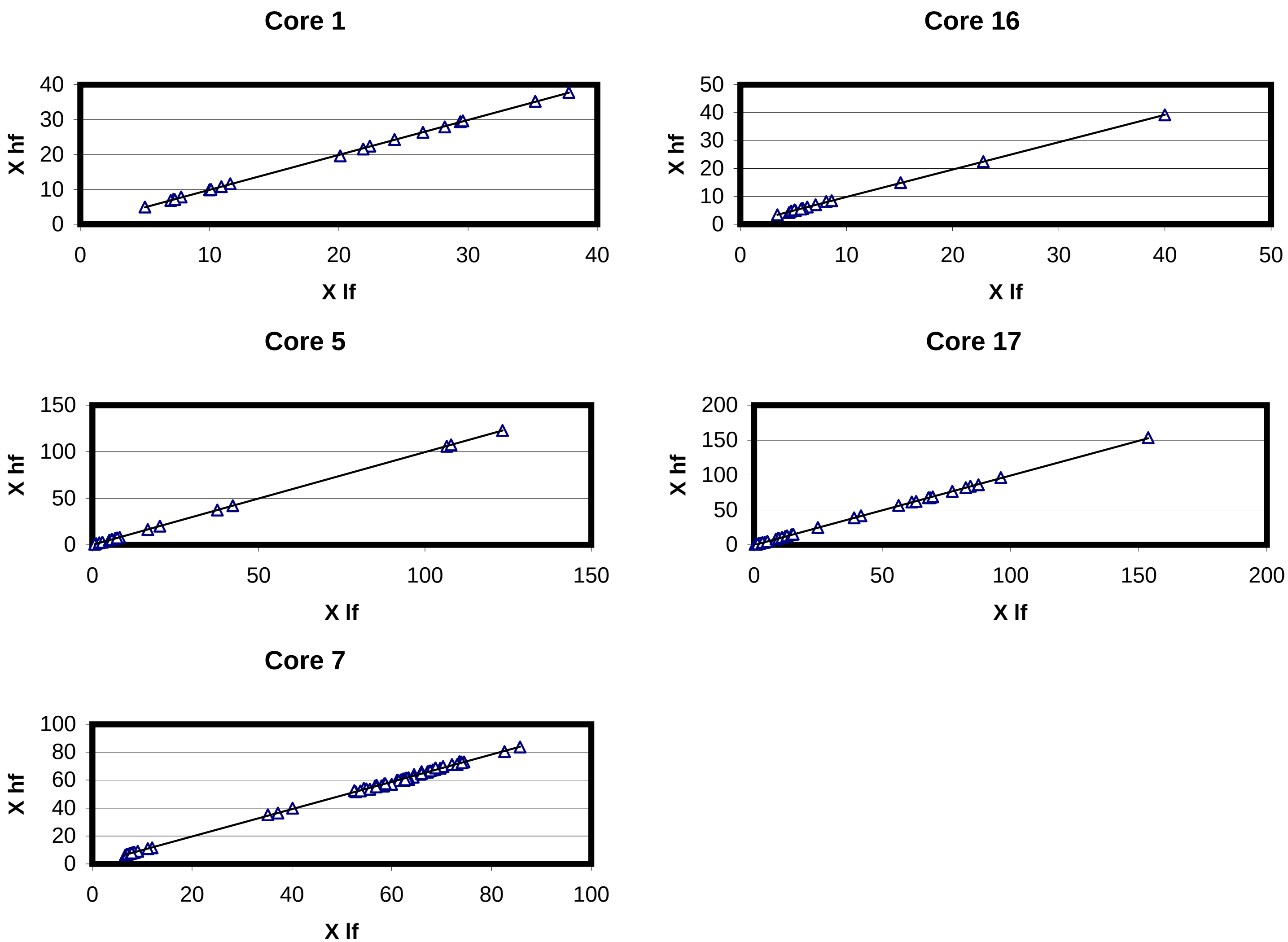
PROVENANCE OF LACUSTRINE SEDIMENTS

Trace element geochemistry of Cores 1 and 16 suggests that a carbonate rocks (consistent with the Lockport Escarpment) was the source for lacustrine sediments in the Tonawanda basin (see Appendix E for trace element data). Both cores are rich in Fe, Al,

Figure 24. Comparison of low frequency and high frequency magnetic susceptibility Data.

Figure 24 shows low frequency and high frequency magnetic susceptibility data. Linear relationships indicate a lack of significance with respect to soil forming processes.

Figure 24. Comparison of low frequency and high frequency magnetic susceptibility data.



Ca, and Mg which represent mineral composition consistent with local dolostone deposits of the Lockport Escarpment (Appendix E). The data also suggest enrichment in aluminum (Core 1), that may suggest Al-silicate rich clays in lacustrine sediments of the Holley Embayment. Benson et al. (2003) suggests that this relationship may indicate the source rocks that were glacially abraded (Figure 25).

Sm/La ratios > 1 (normalized to the upper continental crust) indicate enrichment in La associated with a granitic source rock (Figure 26). Where the ratio is <1, a carbonate source rock enriched in Sm is indicated (Hannigan, personal communication, 2004). The sediments of the Holley Embayment, represented by cores 1 and 16, are consistent with a carbonate source for sediment (dolomite of the Lockport Escarpment). Core 16 shows a slight mixing of carbonate and granitic sources between 0-60 cm (Lockport Escarpment (Silurian) and Canadian Shield (Grenville) rocks). The granitic signature is consistent with REE concentrations of muscovite (Taylor and McClennen, 1985; Appendix E; Figure 26).

PALEOECOLOGY

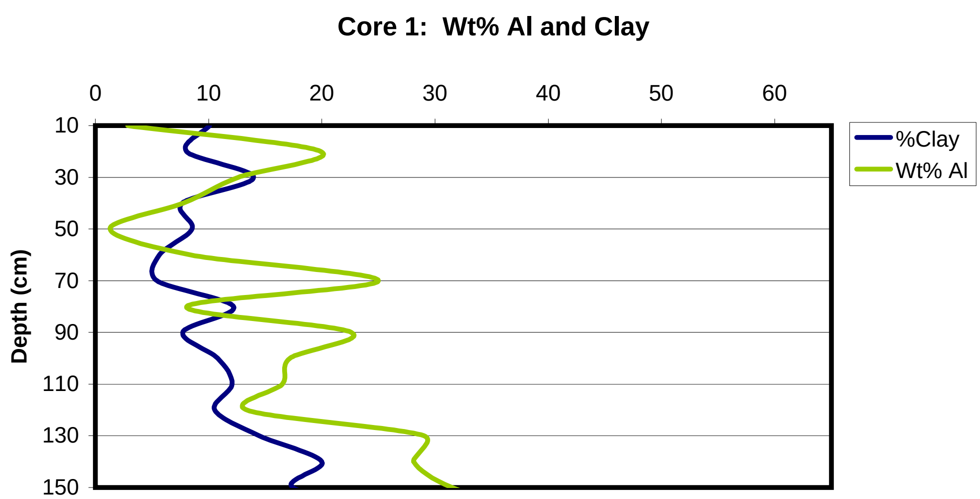
Mollusk species from core 15 include: a) riverine species (terrestrial mollusca that live along sandy-bedded creeks or streams on branches or other debris in the running water); b) fresh-water pond and near-shore species (mollusca that lived in shallow, warm water bodies with adequate vegetation along shorelines); and c) lacustrine species (mollusca that could swim; living on floating objects or vegetation in fresh water lakes of various depths) (Figure 21).

LaRocque (1966) provides a detailed analysis of Pleistocene mollusca in Ohio, stating that tens to hundreds of thousands of mollusk shells can be found in one quart of

Figure 25. Weight percent Al and clay (glacial flour).

Figure 25 shows the weight percent of Al concentrations and clay size particles in cores 1 and 16. Clearly all clay sized particles are not aluminum-rich, but a majority appears to be, as they share similar trends. This is consistent with a slight granitic signature in sediments of the Holley Embayment, and suggests that both the Lockport Escarpment and granitic rocks (possibly the Canadian Shield) were abraded to produce the glacial flour (clay-sized particles; Cannon, 1955).

FIGURE 25. Wt% Al and CLAY DEPOSITION (GLACIAL FLOUR).



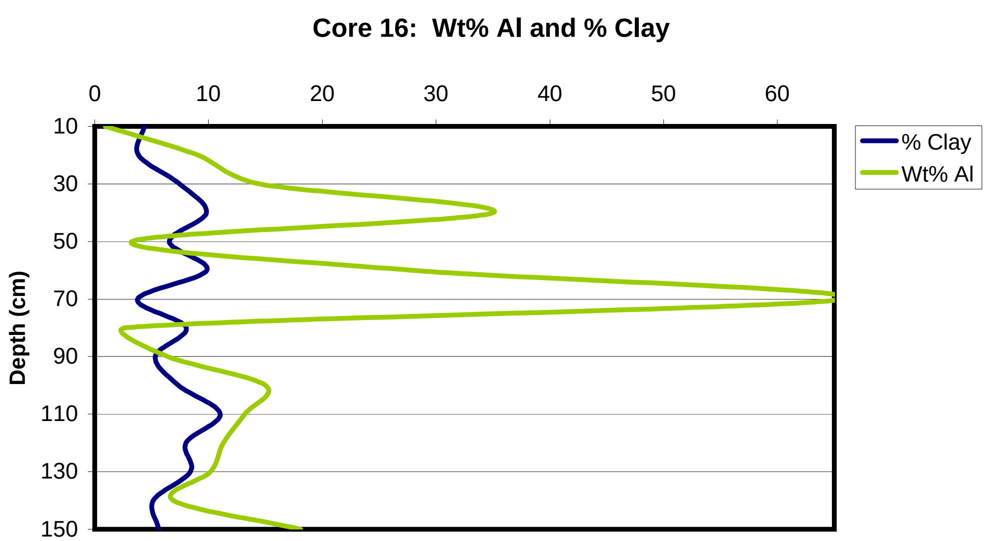


Figure 26. La/Sm Ratio.

Figure 26 shows the ratio of two REE (La/Sm) for the purposes of determining the parent material producing the REEs. Values greater than 1 indicate a granitic source enriched in light REE (LREE) and values less than 1 indicate a carbonate source (depleted in LREE). The ratios all indicate a granitic source consistent with the Canadian Shield rocks of Grenville age.

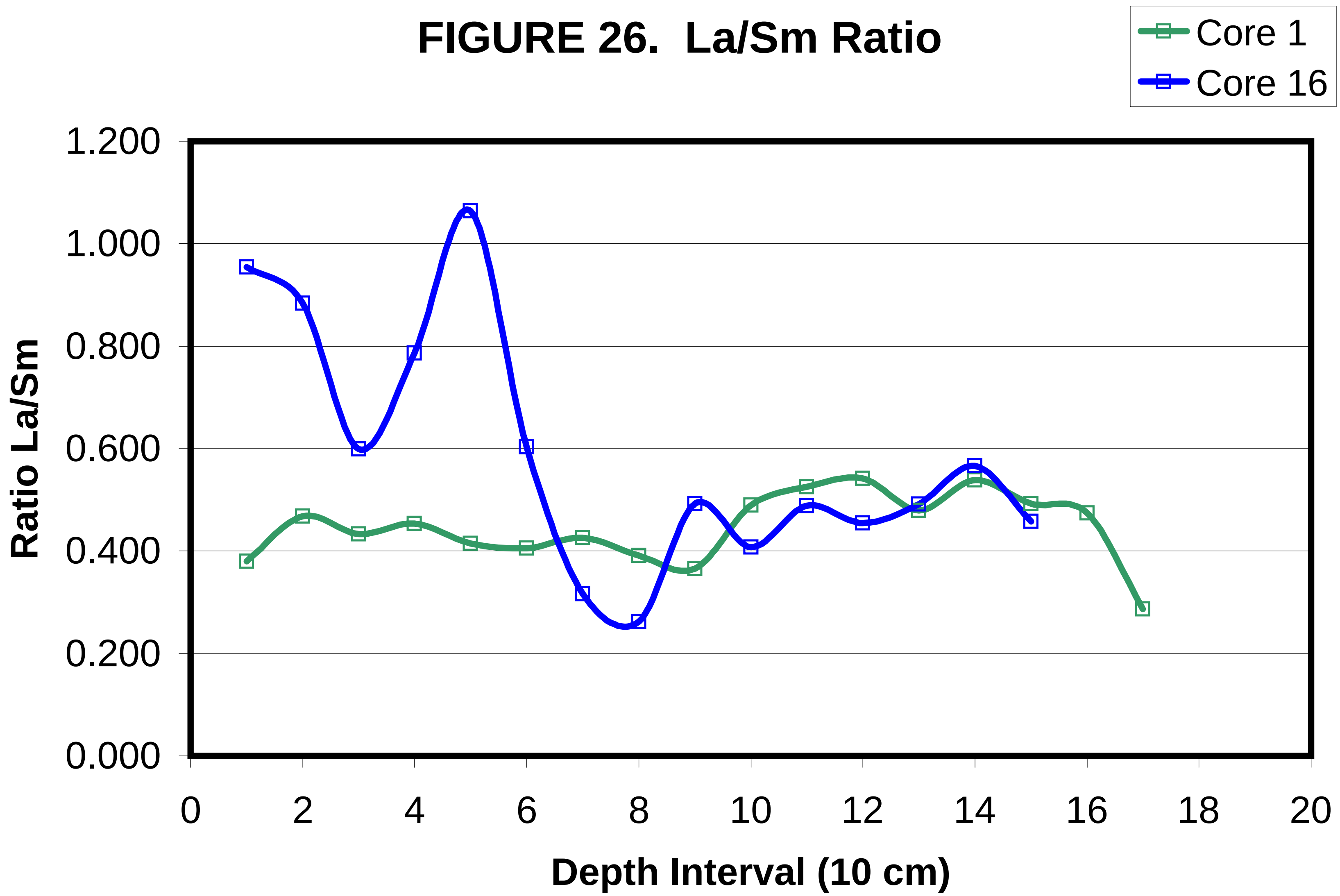


FIGURE 26. La/Sm Ratio
Core 1
Core 16
Ratio La/Sm
1.200
1.000
0.800
0.600
0.400
0.200
0.000
0
2
4
6
8
10
12
14
16
18
20
Depth Interval (10 cm)

many marl deposits of Ohio, and that shell deposits may represents tens to hundreds of generations of a single species. Lacustrine sedimentation typically progresses through three stages affecting molluscan assemblages (LaRocque, 1966). The first stage is sediment deposition with no vegetative growth. Stage two is represented by marl accumulations from either chemical or biological processes (e.g., *Chara*'s consumptive use of CO_2 to produce $CaCO_3$ precipitates; heavy build ups cause them to sink to the lake bottom and decompose). The final stage of lake development affecting lacustrine mollusca assemblages is when the lake bed fills up with plants. This causes water levels to become shallower as more and more plants grow, forming peat deposits. Core 15 represents all three phases of lacustrine development, associated with eutrophication based on its stratigraphy, though this is not consistent across the Holley Embayment (Figures 12, 20 and 23). According to Yu (2000) the presence of a diverse mollusk fauna indicates high biological activities associated with ponds. This is consistent with drainage of Glacial Lake Tonawanda to lower pond levels with the onset of marl deposition (Figures 12, 20 and 23). Draining of Glacial Lake Tonawanda to pond levels would have coincided with the opening of eastward drainage in the Ontario basin, following detachment from stagnant ice at the Barre Moraine with that of the actively retreating Laurentide Ice Sheet (Figure 27). Pond level would have been associated with water levels below that of the Barre moraine and beaded esker deposits near Holley, New York, and associated with shore line interpretations of Cadwell et al. (1988), Muller (1977), and D'Agostino (1958).

Figure 27. Landscape morphology of retreating ice sheet.

This figure illustrates ice retreat from the Batavia moraine to the position of the Barre moraine in western New York. Elevations of features are summarized in Table 5. This illustration is not to scale.

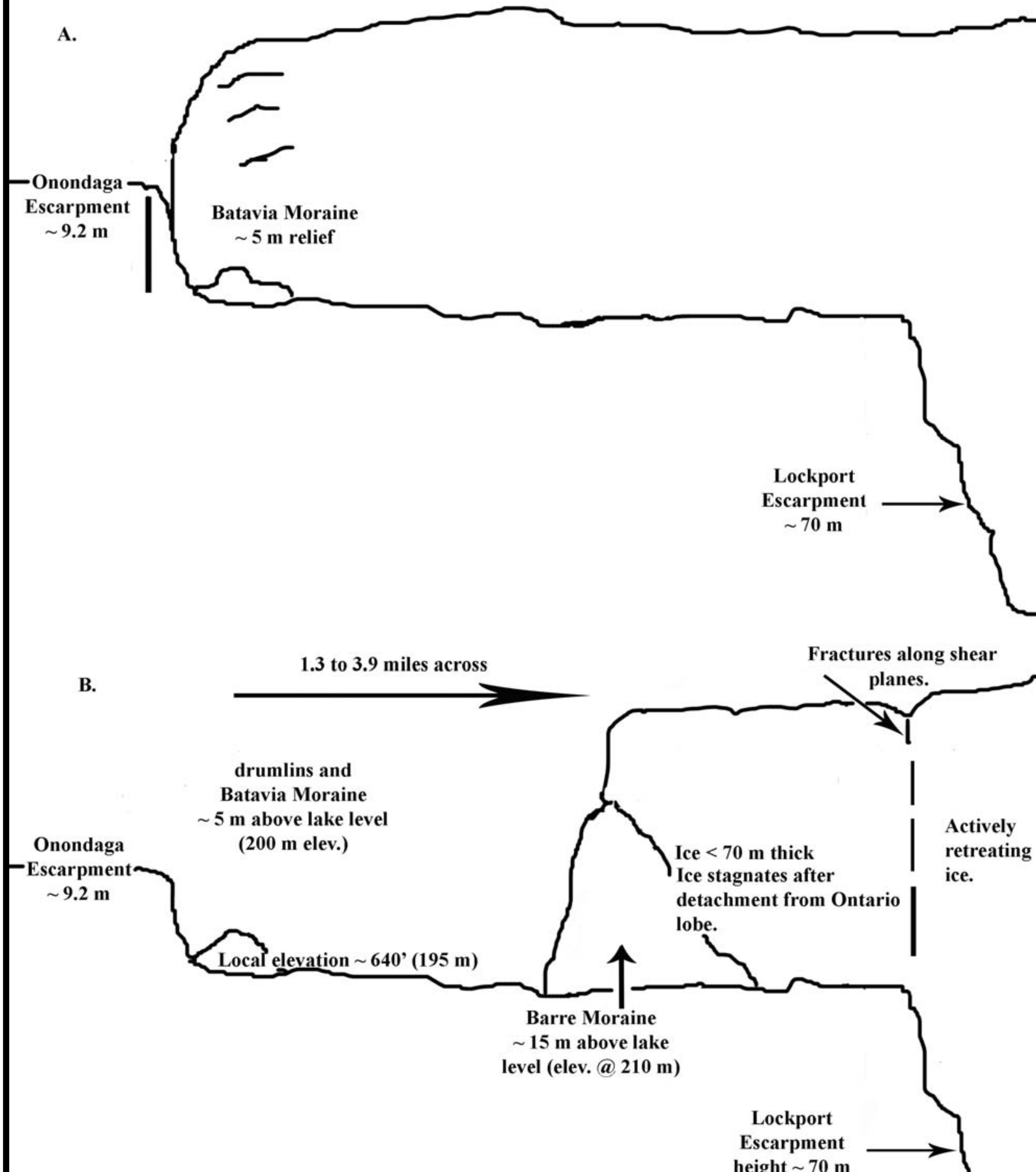


Figure 27. LANDSCAPE MORPHOLOGY OF RETREATING ICE SHEET. NOT TO SCALE.

DISCUSSION AND INTERPRETATIONS

LANDSCAPE MORPHOLOGY OF THE TONAWANDA BASIN

The Batavia moraine represents the last major Wisconsinan readvance into western New York. The Barre, Albion and Carlton Moraines represent minor readvances. The assemblage of landforms consisting of end moraines, eskers and kames and a lacustrine plain indicates an ice contact, proglacial environment for the Tonawanda Basin (Figure 1). The Barre moraine is composed of stratified drift that interfingers with lacustrine sediments in the Tonawanda plain (Brennand, 2000). Associated with this end moraine, are esker deposits with kame and kettle topography, one of which ends in a classic "Gilbert-style" delta that interfingers with sediments on the lacustrine plain forming rhythmic bedding (Muller, 1977b; 1983; Benn and Evans, 1998; Figure 12). Stratified drift of the Barre moraine and esker deposits represent previously deposited, re-worked sediments of local origins (E. Muller, personal communication, 2002). The presence of rhythmically bedded sediments in the lacustrine record of the Tonawanda Basin supports the overall interpretation of an ice-contact, proglacial lake environment during deposition of the Barre moraine and adjacent eskers (see Core 1, Figure 12). Brennand's (2000) morphological classification includes: a) long, dendritic eskers; b) short, subparallel eskers; c) short deranged eskers, with eskers terminating subaqueously or subaerially. Esker deposits associated with the Barre moraine are subparallel to one another and near Holley, New York, and have large scale dunes or megaripples across their top surfaces (Figures 3 and 5). At first glance, the presence of the megaripples at suggests a subaqueous depositional environment. However, the Tonawanda strand line mapped by Muller (1977a) and Cadwell et al. (1988) is at approximately 650 feet a.s.l.

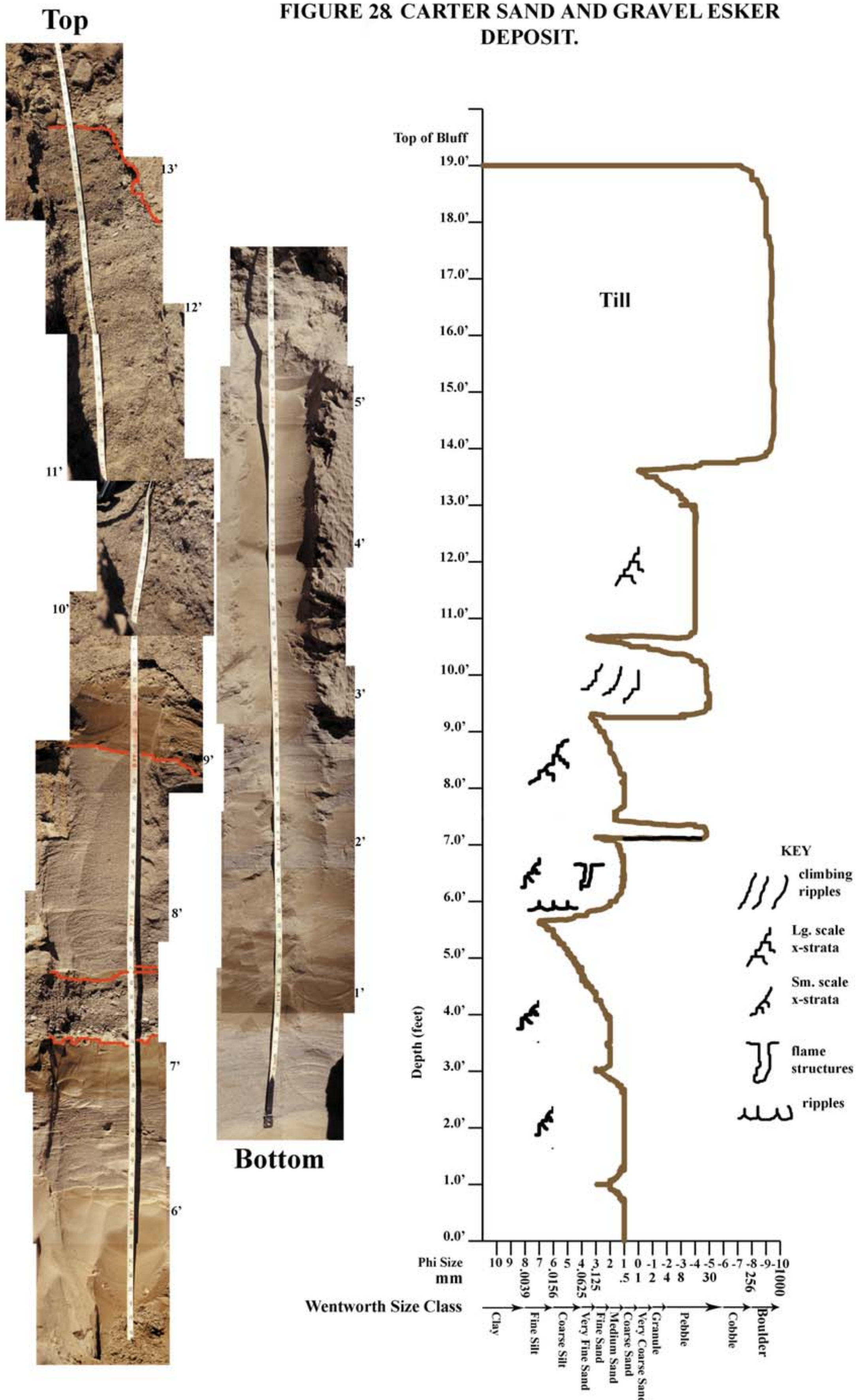
There is no strand line identified that corresponds with an elevation of 689 feet a.s.l., corresponding to the elevation at the base of the megaripples on the top surface of the delta (Figures 3 and 27; U.S.G.S. Holley Quadrangle topographic map, 7.5", 1983). A strandline at this elevation is also unlikely because it would place water over the Onondaga Escarpment and the Niagara Falls moraine, well after water associated with glacial Lake Warren III had begun to lower within the Erie basin. Glacial Lake Tonawanda's depth during the ice-contact phase of its existence would have been a discontinuous ~ 75+ feet (23 m) to basal till. This depth is based on the location of rythemites and field observations (several cores were dug to 18 m without reaching basal till or refusal by auger (Figures 3, 4, 12). The presence of dunes or megaripples on the esker near Holley, New York, would then be subaerial, indicating excess sediment supply, which is also supported by the presence of coarser grained particles (gravels to pebbles) in the esker stratigraphy (Figure 28; Beyrle, 2002). The shoreline mapped by Muller (1977a) and Cadwell et al. (1988) most likely represents the pond stage of Glacial Lake Tonawanda, which persisted for a much longer period of time than did the ice-contact, proglacial lake phase, though it did briefly occur.

Eskers represent accumulated sediment within subglacial conduits through which melt water flowed towards the ice front (Brennand, 2000; Ashley and Warren, 1997; Warren and Ashley, 1994; Syverson et al., 1994). As the subglacial conduit discharged water into the Tonawanda Basin at the ice front, ice motion would have slowed, paused and subsequently ended, thus initiating down wasting. This scenario has been documented at the Bering Glacier following a subglacial outburst in 1994/1995 that

Figure 28. Stratigraphy of an esker deposit, Albion, New York.

This deposit is located in the Carter Sand and Gravel Quarry, Eagle Harbor Road, Albion, New York. This esker deposit is primarily composed of stratified sands and gravels. This stratigraphy represents part of the esker located along of the western edge of the quarry. Along the eastern portions of the quarry, sediment sizes are much finer and are predominantly represented by silt and sand deposits. Throughout the quarry, large scale channel fill deposits can be seen (see Figure 30).

FIGURE 28. CARTER SAND AND GRAVEL ESKER DEPOSIT.



deposited sediments infilling part of the Tsivat Lake basin (Fleisher et al., 2003; Muller and Fleisher, 1995).

The sediments of Glacial Lake Tonawanda are primarily silt-sized particles with sand interbeds derived locally during deposition of sediments discharged subglacially into the basin (Figures 4, 5, and 12). Sediments entering the lake via the subglacial conduit formed a "Gilbert-style" delta near Clarendon, New York (a beaded esker; Benn and Evans, 1998; Brennand, 2000). Esker deposits predominantly consist of sand-sized particles interbedded with gravels and fines (Beyrle, 2002). The red color of esker sands is consistent with interbedded sand found in lacustrine sediments at Core 1 and elsewhere (Figure 12; Beyrle, 2002).

Stratigraphy in the esker located along Eagle Harbor Road, Barre, New York, suggests that it is predominantly sands with beds of gravel and fines (Figure 28). Eskers are represented by sand and gravel facies that fine upward towards the ice margin (Brennand, 2000; Ashley and Warren, 1997; Warren and Ashley, 1994; Syverson et al., 1994). Both esker deposits trend to the S-SW, and are represented by a series of cut and fill channels (Figure 29). Downwasting of the ice sheet associated with stagnant ice conditions is necessary for preservation of the of the esker deposits (Brennand, 2000). Backwasting associated with active retreat, as suggested by Brennand (2000) would have existed only briefly before downwasting began. Benn and Evans (1998) suggest that melt-out from large buried masses of ice can produce ridges and mounds (moraines and kames) that might suggest remnants of former channels. This provides a possible explanation for formation of the Barre moraine and associated kame and esker deposits. This is supported by the Barre moraine which is draped over the terminus of the

Figure 29. Carter sand and gravel pit, Albion, New York.

Figure 29 shows the stratigraphy at the bottom of the esker deposit, along the western limits of the quarry. There is an additional 6 feet of colluvium below this level. These photographs show the stratified sands with thin lenses of silt in ripple troughs. There is a zone above the tape measure that is carbonate-rich, representing a paleo- water table.

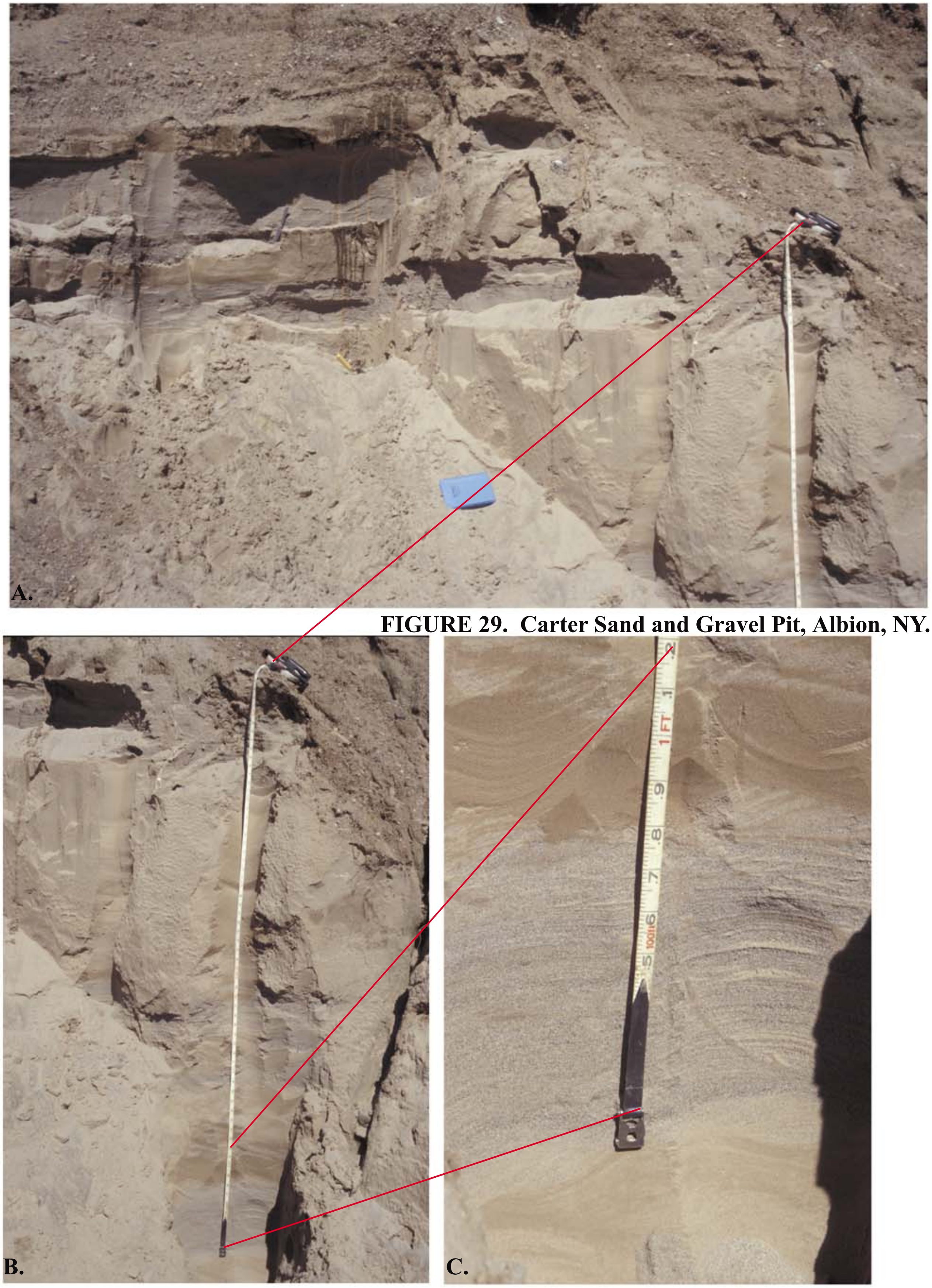


FIGURE 29. Carter Sand and Gravel Pit, Albion, NY.

Barre esker on Eagle Harbor Road, in Albion, New York. Muller (1977a) mapped the esker as draped over the Barre moraine, but this was later changed to reflect the moraine superimposed on the esker (Cadwell, 1988). The moraine superimposed on the esker suggests that moraine development occurred following a pause in actively by the retreating ice sheet lobe but prior to stagnation. Melt out is further supported by the presence of dewatering structures in the quarry sediments that most likely occurred during deposition of the overlying till surface (Figure 30; Menzies, 2002).

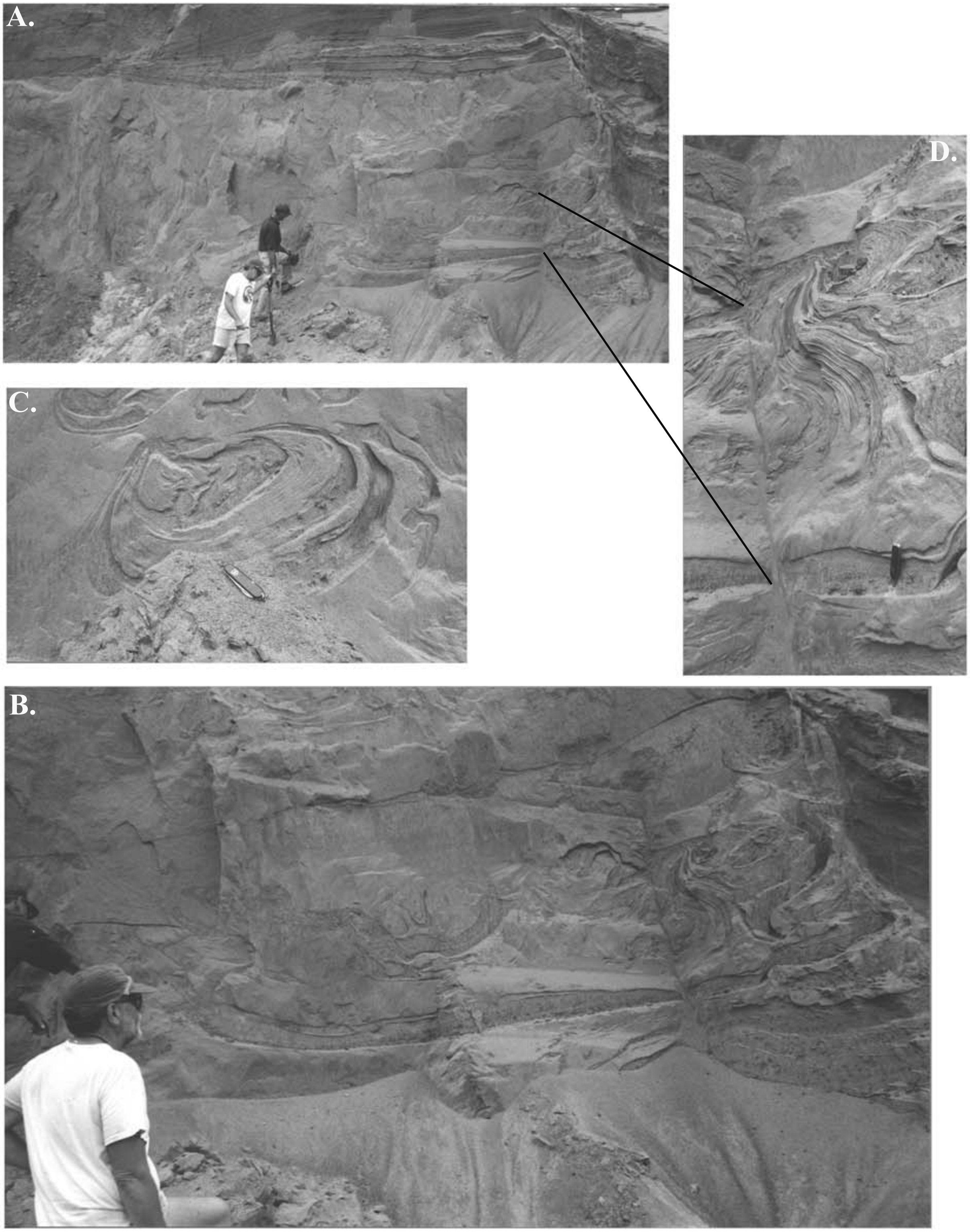
LACUSTRINE SEDIMENTATION AND STRATIGRAPHY

Lacustrine sedimentation in the Tonawanda Basin can be summarized to have included 50-90% silt particles with 50-30% clay-sized particles (Figure 12). Lacustrine sedimentation is associated with < 10% organic carbon and > 10% carbonate content (Figures 14a, 15, and 16). In intervals that contain < 10% carbonate, mollusk habitation is not present (Figures 15, 16, and 25). This is particularly true of early lacustrine sedimentation associated with rythmites and clay, due primarily to a high suspended sediment load and turbidity. Where organic carbon content is > 10%, sediments are associated with muckland formation and correspond to no mollusk habitation (Figures 14a and 25). Glacial Lake Tonawanda sediments are rhythmically bedded silts and sands that grade into silt and silty marls before infilling of the lake permitted peat development due to water warming (Figure 13). Marl deposition suggests that ice-contact discharge had ceased and that lake waters were calm, warm and shallow permitting carbonate precipitation and an actively biological community.

Figure 30. Carter sand and gravel, dewatering structures.

This series of photographs taken during August, 2003, show dewatering structures associated with the esker in the Town of Barre (Figures 4, 5, and 28). The dewatering structures suggest that the ice above was discharging melt waters.

FIGURE 30. CARTER SAND & GRAVEL, DEWATERING STRUCTURES.



<u>CORRELATION WITH WESTERN PORTIONS OF THE TONAWANDA BASIN</u>

D'Agostino (1958) mapped the Tonawanda lacustrine sediments as fine sand associated with beach sand from a test pit just south of Clarence Center that is located south of the Niagara Falls moraine, which is typically associated with Glacial Lake Warren (Figure 1, Table 3). These test pits are outside the physiographic region of the Tonawanda Basin as mapped by Muller (1977a) and Cadwell et al. (1988), and may actually represent beach sediments associated with Glacial Lake Warren (stage III) or Glacial Lake Lundy (Muller and Prest, 1985).

D'Agostino (1958) characterized the lacustrine sediments of the Tonawanda basin as a blue-black, hard pan overlain by peat, marl and additional peat, with a repetitious sequence of blue-black clay and marl lenses along the southwest portions of the basin. Smith and Calkin (1990) and Smith (1990) attribute a purplish-gray, water lain silt till as having been associated with Lake Iroquois deposits in Niagara County. This is consistent with stratigraphic findings in the Holley Embayment that consist of grey clay, rhythmically bedded grey clays/red sands, marls and peat deposits (Figure 12). D'Agostino (1958) attributes the blue-black clay and marl rythmites to increasing water depths resulting from isostatic uplift, but states that they are not seen in the eastern portions of the basin. D'Agostino suggested that differential uplift in the western portions of the basin resulted in lowering of drainage outlets in the west end of the basin and uplift in the Holley Embayment resulted in uplift of drainage outlets causing ponding of the eastern portions of the basin. I would suggest that ponding of waters in the Holley Embayment might simply reflect elevated topography surrounding the basin (drumlins

TABLE 3. SUMMARY OF SEDIMENTARY ANALYSIS BY D'AGOSTINO (1958).

	Location/ Description	Depth of Sample	Classification	Medium-grained sand	Fine-grained sand	Finer sized sediment
Sample 1	.1 miles south of Buffalo-Millersport Rd, on Transit Rd Clarence Center Quadrangle	2 to 3.5 feet deep observation hole (2 feet wide)	Silty sand; 70-75% fine sand; 25-30% silt; 20 gr sample	None	0.60 gr	19.40 gr
Sample 2	Black Creek, Walcott Rd, Clarence Center Quad.	4 feet from surface	Silty sand; 20 gr sample	None	3.10 gr	16.90 gr
Sample 3	0.05 mi north of Miland Rd in Wolcottsville Quad	5 feet from surface	Fine to very fine sand (sand bar); 20 gr sample	1.50 gr	7.30 gr	11.20 gr
Sample 4	0.53 mi northeast of Forest Rd in Tonawanda East Quad.	1 foot within sand dune on Heim Road	Sand dune; fine sand	4.05 gr	10.30 gr	5.65 gr
Sample 5	0.35 mi northeast of Forest Rd in Tonawanda East Quad.	2 to 3 feet deep in observation hole on Heim Rd.	Beach sand; fine sand; 20 gr sample	3.80 gr	10.10 gr	6.10 gr
Sample 6	0.30 mi north of Forest Rd. on Buffalo-Millersport Rd., Clarence Center Quad.	1.5 ft hole in beach sand	Fine sand (beach sand); 20 gr sample	4.60 gr	10.50 gr	4.90 gr
Sample 7	0.75 mi north of Rapids Rd in Wolcottsville Quad on Greenbush Rd	Delta area of Murder Creek; 1 to 2 ft observation hole	Fine to very fine sand; 20 gr sample	2.60 gr	7.40 gr	10.00 gr

Table 3. Summary of sedimentary analysis and sample descriptions from D'Agostino (1958). D'Agostino used sieves to determine percentages of grain sizes from grab samples at various depths in different observation holes ranging from 2-3 feet deep in western portions of the Tonawanda Basin(?).

and moraines). Certainly isostatic uplift played a role in the deglacial history of western New York, but the effect remains to be determined.

TONAWANDA CHRONOLOGY AND THE DEGLACIAL HISTORY OF WESTERN NEW YORK.

Calkin and Fenestra (1985) summarize the deglacial history of western New York and constrain the time between the ice front position at the Huron moraine and the Barre moraine as occurring between 13,000 and 12,000 years BP (Table 4). Furthermore, they suggest a date of 12,000 to 12,100 years BP as corresponding to post Glacial Lake Warren. Muller (1977b) dated wood overlying gravel and water-bedded lake clays found near a breach in the Barre moraine to 10,920 ±160 years (site located west of Lockport, New York; Figure 2). This suggests that Tonawanda basin drained prior to 10,920 years BP, after which shallow water levels would have promoted plant growth and eutrophication of the lake leading to the muckland environment in the Holley Embayment characterized by peat deposition.

While it is possible for debris covered ice to remain for up to 1,000 years, it is likely that ice at this location existed for years to decades, based on observations at the Bering Glacier by BERG, in which an ice berg (the size of a three story barn) stranded in a sandur deposit melted within 7-8 years (P. J. Fleisher, personnel communication, 2002). Glacial Lake Tonawanda may have existed between 12,000 to 10,920 years BP (Calkin and Fenestra, 1985; Muller, 1977b). However, it is likely that ice contact, proglacial lacustrine deposition represents a very short period of time coinciding with esker and beaded esker deposits near Barre and Holley, New York. Esker sedimentation indicates that their formation occurred from several events, due to the presence of rip-up

TABLE 4. GLACIAL CHRONOLOGY OF WESTERN NEW YORK.

Reference	Age (years BP)	Lake	Lake Stage	Water Elevation
Hansen (1989)		g. L. Whittlesey		738 ft asl.
Hansen (1989)		g. L. Warren		685 ft asl
Hansen (1989)				670-675 ft asl
Hansen (1989)				660 ft asl
Hansen (1989)		g. L. Lundy	L. Grassmere	640 ft asl
Hansen (1989)			L. Dana (eastern Ontario Basin)	590 ft asl
Hansen (1989)			L. Elkton	620 ft asl
Carney (1916)		g. L. Warren		680 ft asl
Calkin & Fenestra (1985)	13,000 BP (from Goldthwaite, 1958; Dreimanis, 1966, Calkin, 1970; Barnett, 1978; Fullerton, 1980; Totten, 1982)	g. L. Whittlesey	Ice margin at Port Huron Moraine (drainage to the west via Ubly, Michigan)	740 ft asl (226 m)
Calkin & Fenestra (1985)		g. L. Warren	I (initial barrier, Marilla Moraine)	685 ft asl (209 m)
Calkin & Fenestra (1985)	13,050±50 BP (from Totten, 1982)		II (briefly lowered to L. Wayne phase with drainage to the East)	675 ft asl (206 m) (L. Wayne phase @ 660 ft)
Calkin & Fenestra (1985)		g. L. Warren	III, Ice margin at Batavia Moraine	670 ft asl (204 m)
Calkin & Fenestra (1985)	Post Warren ages of 12,000 to 12,100 BP (from Dreimanis, 1966; Calkin, 1970 and Fullerton, 1980)	g. L. Lundy	g. L. Grassmere (ice margin at Barre Moraine	640 ft asl (195 m)
Calkin & Fenestra (1985) (Spencer, 1894)		g. L. Lundy	L. Elkton (ice margin at Albion Moraine)	620 ft asl. (189m)
Calkin & Fenestra (1985) (Fairchild, 1907, Calkin 1970)			L. Dana (in Erie Basin) (considered inappropriate)	590 ft asl (180 m)
Calkin & Fenestra (1985) (Kindle & Taylor, 1913)	10,920 ±160 BP		L. Tonawanda (shift from drainage over scarp at Lockport to Niagara River)	585 ft asl (178.3 m)
Calkin & Fenestra (1985) (Calkin & Brett, 1978)	10,450 ±400 BP (from Muller, 1977)		L. Tonawanda (eastern drainage near Holley, New York)	560 ft asl (170 m)
D'Agostino (1958)			L. Tonawanda	580 feet asl (referred to as middle shore line)
Muller & Prest (1985)		g. L. Warren	I	850 ft (259 m) (ref: Blackmon, 1956; Calkin, 1970)
Muller & Prest (1985)		g. L. Warren	III	670 ft asl (204.2 m)
Muller & Prest (1985)		g. L. Grassmere		640 ft asl (195 m)
Muller & Prest (1985)	12,600±400 and 12,080±300 (from Muller, 1965)	g. L. Lundy		620 ft asl (189 m)
Muller & Prest (1985)	12,500 to 12,400 BP (from Prest, 1970); 12,600 - 12,100 yrs BP; Muller & Calkin, 1993)	g. L. Iroquois	Maximum	335 ft asl (102.1 m)
Muller, 1977b	10,920±160 years BP		Glacial Lake Tonawanda	

Table 4: Summary of ice contact, proglacial water depths in Erie-Ontario basins during Late Wisconsinan ice retreat from western New York. Where a particular paper presented data from another paper, the reference is given in parentheses.

clasts embedded within the esker sediment in the Barre esker (Figure 31). Marl sedimentation in the lake basin may have represented a longer period of time, and peat accumulation associated with the swamp environment exists to the present day, except where the swamps have been drained and are presently used for agriculture (muckland). Additional radiocarbon ages are required to constrain lake stages in the Tonawanda Basin.

Carbonate content in Tonawanda sediments ranges from 5-10% by weight, which is consistent with carbonate content in other cores, such as core 17 and 1, where mollusks were noted but not sampled (Figure 15, Appendix F). A likely cause for slow inhabitation of Glacial Lake Tonawanda waters by mollusk species is high suspended sediment load and turbidity associated with ice-contact, proglacial depositional environment.

D'Agostino's (1958) study of the history of the Tonawanda Basin suggests that the lake existed as a remnant of Lake Lundy and Lake Dana (co-existing lakes in the Lake Erie and Ontario Basins following glacial retreat into Canada). Calkin and Fenestra (1985) suggest that Lake Dana (or Glacial Lake Dana) existed along the ice front in eastern portions of the Ontario Basin and drained eastward. It is likely that water in the Tonawanda Basin initially spilled in from Glacial Lake Warren (stage III) following ice retreat from the Batavia moraine and readvance to the Barre moraine. It is unlikely that waters from the Ontario basin would have been able to spill into the Tonawanda basin, because of locally high relief associated with the Barre moraine and drumlins that surround the basin (Muller, 1977b).

Figure 31. Guinea Road quarry, Clarendon, New York.

A. A panoramic view of the Guinea Road Quarry during the summer of 2002. Large scale channel fills can be seen above truck in picture. This region is within the esker that begins near Holley, New York, and extends into the Holley Embayment.

B, C. Close up of rip-up clasts that can be seen in Figure C with Whitney Autin for scale. Clasts are approximately 5 feet in length and represented renewed flow within the subglacial conduit, suggesting that the esker represents sedimentation from more than one flow event.

FIGURE 31. GUINEA RD QUARRY, CLARENDON, NY.

CONCLUSIONS

Based on the evidence presented herein, it is suggested that Lake Tonawanda should be referred to as Glacial Lake Tonawanda, primarily because of interfingering of sediments associated with eskers at the ice front (Barre moraine position) and lacustrine silts and clays in the Tonawanda Basin (Figures 12, 25). Glacial Lake Tonawanda existed as a short-lived ice-contact, proglacial lake, followed by a shallow, ice-contact pond, as ice stagnated between the Barre moraine and the Lockport Escarpment, later evolving into a swamp or muckland. The stratigraphic record suggests a dynamic environment associated with the actively retreating ice front of the Laurentide Ice Sheet 13,000 to 12,000 years ago BP, in western New York (Calkin and Fenestra, 1985; Muller, 1977b).

The ice-contact, proglacial lake phase is characterized by the deposition of silt-sized particles with zones of carbonate mud and ferromagnetics (Figures 12, 13, 15, 18 and 25). Lacustrine sedimentation during the ice-contact, proglacial phase has a carbonate source-rock signature, based on trace element geochemistry, which is consistent with the Lockport Escarpment (Figure 26; Appendix E). Turbidity and high sediment loads initially prevented mollusk survival in the lake, but once waters became calm and clear a diverse biological community developed (Appendix A).

The pond phase is associated with calm, warm waters that permitted marl deposition and mollusk habitation of the pond (Figures 12 and 20). Mollusk habitation requires carbonate-rich water for the production of aragonite shells (based on XRD analysis). During the pond phase there was a decreasing trend in ferromagnetic and carbonate mud accumulations, as melt water discharge into the basin had diminished over time (Figures 15 and 18). The swamp phase is associated with high organic content (> 50%) with less than 10% carbonate, and lacking mollusks (Figures 12, 15 and 20). The tilled surface in most cores shows a slight enrichment in ferromagnetic minerals likely formed by pedogenic processes (Figure 18).

The beaded esker forming the delta entering into the eastern end of Glacial Lake Tonawanda was most likely deposited subaqueously in an ice-contact, proglacial lake whose elevation was 689-715 feet a.s.l., based on GPS elevations and field observations. However, a lack of strand lines suggest that the ice-contact, proglacial lake was short-lived, before active ice-retreat north of the Lockport Escarpment allowed water in the Tonawanda Basin to drain over the escarpment, lowering water in the Tonawanda Basin to the pond stage (640-650 feet a.s.l). The dissected nature of the delta (Figure 4) is most likely the result of stranded ice bergs eventually forming kettles. Buried ice in the delta of the esker may have existed for decades or longer, suggesting that mollusk habitation in the lake basin may also have existed for a prolonged period of time (Core 15, Figures 3, 4 and 20). Additional radiometric dating is required to constrain the period of mollusk habitation, and possibly the duration of buried ice in the delta.

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