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DEGLACIATION OF NOVA SCOTIA: STRATIGRAPHY AND CHRONOLOGY OF LAKE SEDIMENT CORES AND BURIED ORGANIC SECTIONS

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ABSTRACT The deglaciation of Nova Scotia is reconstructed using the AMS-dated chronology of lake sediments and buried organic sections exposed in the basins of former glacial lakes. Ice cleared out of the Bay of Fundy around 13.5 ka, punctuated by a brief readvance *ca*. 13-12.5 ka (Ice Flow Phase 4). Glacial Lake Shubenacadie (1) formed in central Nova Scotia, impounded by a lobe of ice covering the northern Bay of Fundy outlet. Drainage was re-routed to the Atlantic Ocean until the Fundy outlet became ice free after 12 ka. When this lake drained, bogs and fens formed on the lake plain during climatic warming. Organic sediment (gyttja) began to accumulate in lake basins throughout Nova Scotia. Glacierization during the Younger Dryas period (*ca*. 10.8 ka) resulted in the inundation of lakes and lake plains with mineral sediment. The nature and intensity of this mineral sediment flux or "oscillation" varies from south to northern regions. Southern lakes simply record changes in total organic content whereas northern lakes, where most buried peat sections are found, feature a thick inorganic sediment layer. Glacial ice or permanent snow cover and seasonal melting are essential in the formation of this mineral sediment layer; both to provide the water source for erosion, and to prevent plant recolonization and landscape stabilization. Some northern lakes do not appear to record the Younger Dryas event, with organic accumulation starting around 10 ka. During the Younger Dryas, fine and coarse-grained deposits were deposited in Glacial Lake Shubenacadie (2) and other lowland areas at elevations similar to former (12 ka) lake levels, impounded by re-invigorated residual ice caps and permanent snow/aufeis.

RÉSUMÉ *La déglaciation en Nouvelle-Écosse : stratigraphie et chronologie à partir de carottes sédimentaires lacustres et de coupes de matériaux organiques enfouies.* Il s'agit d'une reconstitution à partir de la chronologie par spectométrie de masse (SMA) de sédiments lacustres et de certaines coupes de matériaux organiques enfouis dans les bassins d'anciens lacs glaciaires. La baie de Fundy, libre de glace vers 13,5 ka, a connu une récurrence vers 13-12,5 ka. Le Lac glaciaire Shubenacadie (1), qui s'est formé au centre de la Nouvelle-Écosse, a été endigué par un lobe glaciaire qui recouvrait la sortie de la baie de Fundy au nord. L'écoulement a alors été détourné vers l'océan Atlantique jusqu'au retrait de la glace vers 12 ka. Une fois le lac asséché, tourbières et marécages se formèrent sur l'ancien fond lacustre au cours du réchauffement climatique. Ensuite, la gyttja s'accumula dans les bassins lacustres à travers la Nouvelle-Écosse. L'englacement au cours du Dryas récent (vers 10,8 ka) a entraîné le comblement des lacs et des plaines lacustres par des sédiments minéraux. La nature et l'intensité de ce flux de sédiments ou « oscillation » varient du sud au nord. Dans les lacs du sud, la teneur organique totale a simplement varié, tandis que dans les lacs du nord, une importante couche de sédiment inorganique a été déposée. Certains lacs du nord ne semblent pas avoir enregistré l'épisode du Dryas récent, l'accumulation de matière organique ayant commencé vers 10 ka. Au Dryas récent, des dépôts à grains fins et grossiers ont été mis en place dans le Lac Shubenacadie (2) et autres basses terres à des altitudes comparables aux anciens niveaux lacustres (12 ka), endigués par des calottes glaciaires résiduelles ravivées et de neige permanente ou de lentilles de glace.

ZUSAMMENFASSUNG *Die Enteisung von Nova Scotia : Stratigraphie und Chronologie von See-Sedimentkernen und Torf-Profilen.* Man hat die Enteisung von Nova Scotia rekonstruiert, indem man mittels Massenspektroskopie die Chronologie von See-Sedimenten und vergrabenen organischen Profilen, die in den Becken ehemaliger glazialer Seen ausgesetzt sind, bestimmt hat. Etwa um 13,5 ka verschwand das Eis aus der Fundy-Bai, unterbrochen durch einen kurzen Rückvorstoß um etwa 13-12,5 ka (Eisflussphase 4). Der glaziale Shubenacadie-See (1), der sich im Zentrum von Nova Scotia bildete, wurde durch eine Eislobe eingedämmt, welche den Nordausläufer der Fundy-Bai bedeckte. Die Dränierung wurde zum Atlantischen Ozean umgeleitet bis der Fundy-Ausläufer nach 12 ka eisfrei wurde. Als dieser See austrocknete, bildeten sich Torfmoore und Fehne auf der See-Ebene während einer Klimaerwärmung. Organisches Sediment (Gyttja) begann sich in den See-Becken durch ganz Nova Scotia anzusammeln. Die Vereisung in der jüngeren Dryas-Zeit (etwa 10,8 ka) führte zur Anschwemmung der Seen und See-Ebenen mit Mineral-Sedimenten. Die Art und Intensität dieses Mineral-Sedimentflusses oder seiner Oszillation ist von Süden nach Norden unterschiedlich. Die südlichen Seen belegen einfach Veränderungen im gesamten organischen Gehalt, wohingegen die nördlichen Seen, in denen die meisten vergrabenen Torfprofile gefunden werden, eine dicke anorganische Sedimentschicht aufweisen. Glaziales Eis oder eine permanente Schneedecke und jahreszeitbedingtes Schmelzen sind bei der Bildung dieser Mineral-Sedimentschicht entscheidend, sowohl, um die Wasserquelle für die Erosion bereitzustellen, als auch um die Wiederansiedlung von Pflanzen und die Stabilisierung der Landschaft zu verhindern. Einige nördliche Seen scheinen das jüngere Dryas-Ereignis nicht zu belegen, denn ihre organische Ablagerung begann um etwa 10 ka. Während des jüngeren Dryas wurden feine und grobkörnige Ablagerungen im glazialen Shubenacadie-See (2) und anderen Gebieten des Tieflands in Höhen, die den früheren See-Ebenen (12 ka) entsprechen, abgelagert, eingedämmt durch wiederbelebte Alluvialeiskappen und permanenten Schnee/Aufeis.

INTRODUCTION

For the past 30 years Nova Scotia and New Brunswick lakes have been cored and studied by R. J. Mott with the purpose of defining the late-glacial paleoenvironments and documenting glacier fluctuations during this period. The focus of that research was to establish the vegetation history of the late-glacial to the present, and from that deduce the climate history (*e.g.* Mott, 1975; Jetté and Mott, 1989; Mott, 1991). In most studies of cored lake basins pollen and macrofossil stratigraphy are emphasized (*e.g.* Mott, 1991; Mayle *et al.*, 1993; Marcoux and Richard, 1994). In this report we concentrate on the late-glacial sedimentary record of lake cores and buried organic deposits (Stea and Mott, 1989; Mott and Stea, 1994). The purposes of this report are threefold:

1 . Document the stratigraphy, and chronology of the lake sediments and late-glacial buried peat sections, specifically the period of deglaciation between 14 and 10 ka.

2. Develop an age-depth regression model based on 50 AMS (Accelerator Mass Spectrometry) and beta decay radiocarbon ages from basal lake sediments, in order to extrapolate the age of deglaciation (*cf.* Ogden, 1987) and the onset of the Younger Dryas interval (*cf.* Mayle *et al.*, 1993).

3. Produce a comprehensive paleogeographical reconstruction of the deglaciation of Maritime Canada after 15 ka, based on this new data.

This paper presents a new chronology of deglaciation of Maritime Canada based on AMS dating of plant macrofossils from lake sediment cores in Nova Scotia and New Brunswick updating previous work on lake basin dating and deglaciation (Stea and Mott, 1989; Mayle *et al.*, 1993; Mott and Stea, 1994; Mott, 1994). Forty lake basins are examined, 35 from Nova Scotia and 5 from New Brunswick (Fig. 1). A deglaciation chronology has already been established using beta decay radiocarbon ages from buried peats and organic lake sediments (Stea and Mott, 1989). . The validity of a chronol-

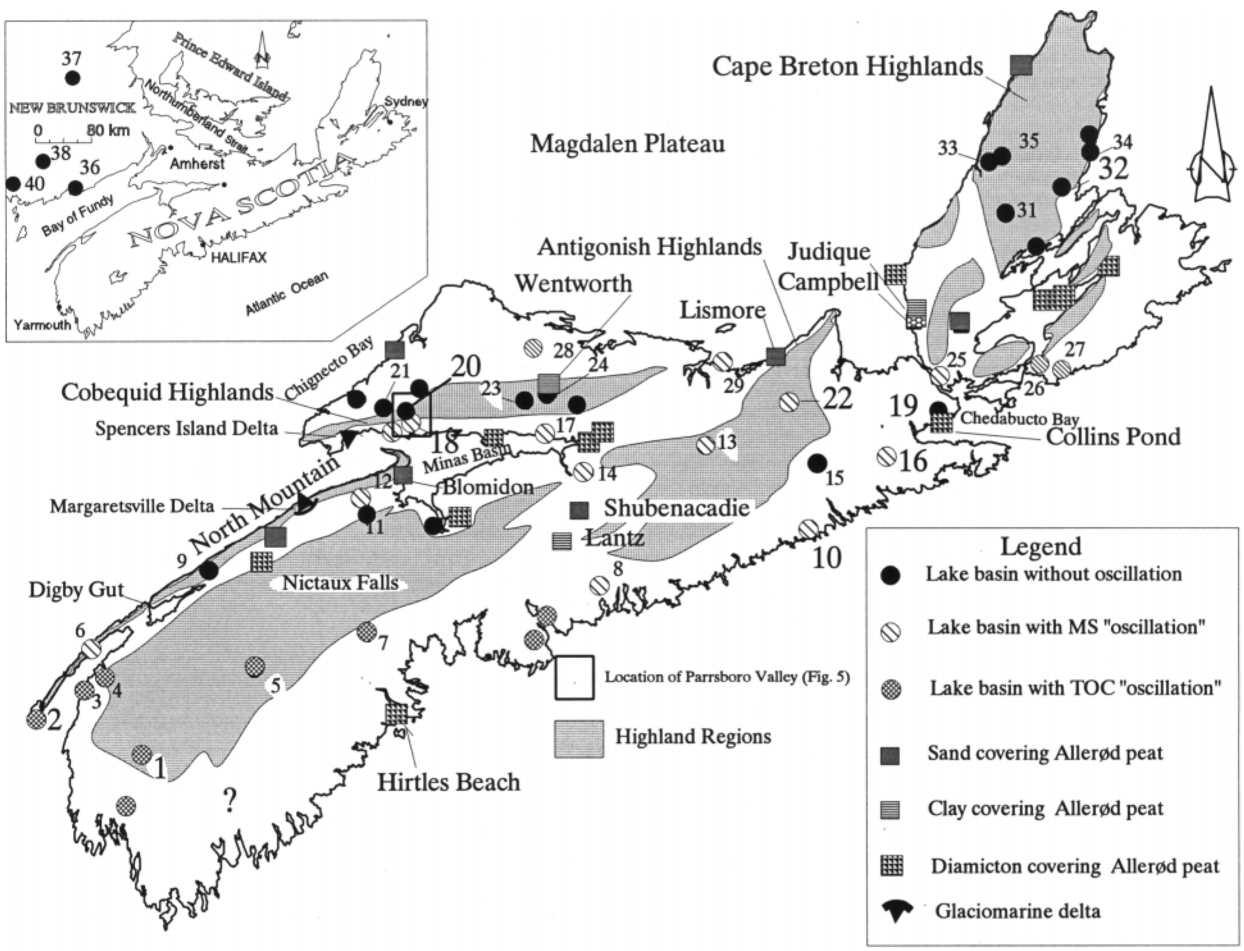

FIGURE 1. Location map of buried peat and lake sites in Nova Scotia classified according to mineral sediment (MS) or total organic carbon (TOC) "oscillations". Lake basins are numbered according to Table I. Some unlabelled lake basins were classified from stratigraphic information in Ogden (1987). Large numbered lakes are in the stratigraphic transect of Figure 2. The Cape Breton Highlands attain elevations of 500 m. The highest point on mainland Nova Scotia (Cobequid Highlands) is 300 m.

Localisation de la tourbe enfouie et des sites lacustres en Nouvelle-Écosse classés selon les « oscillations » de sédiment minéral (MS) ou de carbone organique (TOC). La numéroration des bassins lacustres suit celle du tableau I. Les bassins lacustres sans désignation sont classés selon les données stratigraphiques de Ogden (1987). Les grands chiffres identifient les lacs dont la stratigraphie apparaît en figure 2. Les hautes terres du Cap Breton atteignent 500 m d'altitude. Les hautes terres de Cobequid sont la partie la plus élevée de la Nouvelle-Écosse continentale (330 m).

ogy based on organic lake sediments has been questioned (Sutherland, 1980; Walker and Lowe, 1990) because of the "hard water effect" and redeposition of older organic materials. In this study, we utilized 24 AMS radiocarbon ages from below a distinctive Younger Dryas marker horizon to establish an age for initiation of sedimentation in a lake basin or a "minimum" age for deglaciation.

METHODS

Lake basins were selected for this study based on 1) Evidence from nearby stratigraphic sections buried by glacial or periglacial sediments related to Younger Dryas ice cover (Stea and Mott, 1989; Mott and Stea, 1994; King, 1994), 2) Geomorphological evidence of former ice margins, 3) Uniform geographic distribution.

Lake basins were surveyed, prior to coring, using an echosounder and the deepest part of the lake basin was chosen for coring. The basal sediments were cored using a modified wire-line Livingstone corer (Wright *et. al.,* 1965). Total organic carbon (TOC) was assessed by LECO carbon analysis, taking the difference between total carbon and carbon after digestion in 10 % HCL and drying. Beta decay radiocarbon ages were mostly determined by the radiocarbon laboratory of the Geological Survey of Canada (Table I). Lake sediment samples were pre-treated with hot acid and base washes and rinsed with distilled water. Wood samples were dated using Accelerator Mass Spectrometer (AMS) dating techniques from BETA Analytic (Florida) and Isotrace laboratories (Toronto). Ages provided through the radiocarbon dating laboratory of the Geological Survey of Canada were corrected for $\delta^{13}C$ = -25.0 % PDB. For more detail on the radiocarbon methodologies as well as lake and buried peat site descriptions and palynology, see McNeely and Atkinson (1996).

RESULTS

LAKE BASIN MORPHOLOGY AND STRATIGRAPHY

Four major lithostratigraphic "units" can be discerned from lake cores in Nova Scotia (Fig. 2). In Figure 2, these units are

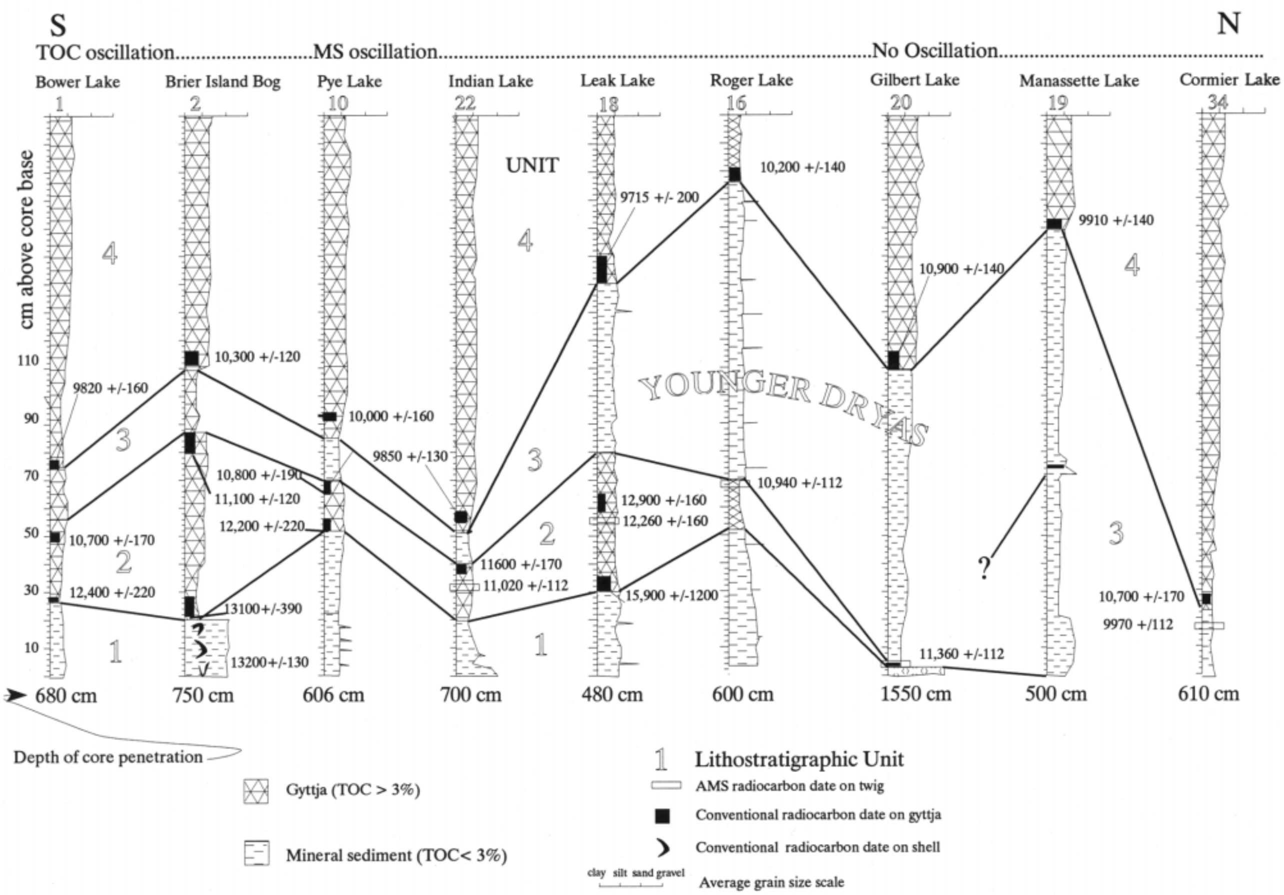

FIGURE 2. Stratigraphic correlation diagram and chronology of the basal 2 m of selected lake basins across Nova Scotia. The total core penetration is shown in centimetres at the bottom of the sections. The vertical scale is in centimetres above the core base. The core grain size variations are represented by median clay on the far left to median gravel on the far right. Beta decay dated intervals are shown as a solid block. AMS dated intervals are shown as open blocks. Units 1-3 can be defined either by oscillation in total organic carbon values (TOC) or a mineral sediment (MS-Unit 3) sandwiched between gyttja units (Units 2/4).

Diagramme de corrélation stratigraphique et chronologie des deux premiers mètres de certains bassins lacustres choisis à travers la Nouvelle-Écosse. La profondeur du forage apparaît à la base des coupes. L'échelle verticale au-dessus de la base de la carotte est en centimètres. La granulométrie varie des argiles moyennes à l'extrême gauche au gravier moyen à l'extrême droite. Les unités 1 à 3 sont définies par une oscillation soit des valeurs de carbone organique total (TOC), soit du sédiment minéral (unité 3) contenu entre deux unités de gyttja (unités 2 et 4).

TABLE I

Radiocarbon ages, core depths and basal model age of lakes in Nova Scotia and New Brunswick

No.	Location	Latitude N/ Longitude W	Laboratory Age/Date/Type	Depth Below YD (cm)	Date	2s	Total Thick. (Unit 2)	Basedate (Age Model)
1.	Bower Lake	4405.24' 65046.56'	C(GSC-5023)	8.0	10700	170	28.0	12253
1.	Bower Lake		C(GSC-5024)	23.0	12400	220	28.0	12253
2.	Brier Island Bog	44012.84' 66022.03'	C(GSC-4556)	5.0	11100	120	75.0	13200*
2.	Brier Island Bog		C(GSC-4443)	65.0	13100	190	75.0	13200*
2.	Brier Island Bog		C(GSC-4431)	75.0	13200	130	75.0	13200*
3.	Lac a Magie	44015.83' 66004.66'	A(TO-1431)	27.0	12140	130	33.0	12533
4.	Church Point	44019.23' 66008.38'	C(GSC-4476)	3.0	12300	130	40.0	12925
4.	Church Point		C(GSC-4440)	38.0	14300	150	40.0	12925
5.	Pat Kempton Lake	44026.08' 66010.15'	C(GSC-5031)	No OSC	11900	190	BASE	11610
5.	Pat Kempton Lake		A(TO-3972)	No OSC	11610	130	BASE	11610
6.	Sandy Cove Lake	44028.97' 66005.20'	A(TO-3343)	96.0	12320	130	105.0	12820#
6.	Sandy Cove Lake		C(GSC-4460)	96.0	13300	130	105.0	12820#
7.	Canoran Lake	44035.00' 64033.0'	C(GSC-4462)	0.0	11200	150	25.0	12085
8.	Porters Lake	44048.22' 63022.75'	C(GX-17565)	4.0	11700	450	21.0	11861
8.	Porters Lake		A(BETA-53125)	32.0	11750	120	21.0	11861
9.	Youngs Lake	44049.14' 65026.44'	C(GSC-5048)	No OSC	12500	220	BASE	11693
10.	Pye Lake	44058.50' 62005.45'	C(GSC-5242)	1.0	10800	190	16.0	11581
10.	Pye Lake		C(GSC-5543)	18.0	12200	200	16.0	11581
11.	Tupper Lake	4501.0' 64035.3'	C(GSC-5049)	No OSC	12800	200	BASE	11993
12.	Silver Lake	4506.81' 64035.67'	A(BETA-16400)	10.0	11350	85	14.0	11469
12.	Silver Lake		C(GSC-5053)	0.0	12600	280	14.0	11469
13.	Piper Lake	45020.83' 62039.63'	A(TO-3973)	0.0	10890	113	15.0	11525
13.	Piper Lake		C(GSC-5252)	13.0	12300	200	15.0	11525
14.	Brookfield	45014.42' 63020.63'	A(TO-3970)	No OSC	10870	112	BASE	10870
15.	Cumminger Lake	45016.11' 6202.63'	C(GSC-5250)	No OSC	-		28.0	12253
16.	Roger Lake	45021.23' 61040.60'	A(BETA-62944)	2.0	10940	112	19.0	11749
17.	Little Dyke Lake	45023.08' 63033.63'	A(TO-3576)	23.0	10930	130	52.5	12400#
17.	Little Dyke Lake		A(BETA-54888)	35.0	11560	130	52.5	12400#
18.	Leak Lake	45026.0' 64020.0'	A(TO-3971)	48.0	12260	130	65.0	13204#
18.	Leak Lake		C(GSC-2728	48.0	12900	160	65.0	13204#
19.	Manassette Lake	45026.96' 61020.63	C(GSC-4964)	No OSC	9910	110	BASE	9910
20.	Gilbert Lake	45028.0' 64021.0'	A(TO-807)	0.0	11360	112	BASE	11360
21.	Welton Lake	45029.0' 64029.0'	C(GSC-4474)	No OSC	9720	110	BASE	9720
22.	Indian Lake	45030.83' 62011.90'	A(BETA-62945)	11.0	11020	112	24.0	12135
22.	Indian Lake		C(GSC-5943)	2.0	11500	170	24.0	12135
23.	Sutherland Lake	45031.11' 63040.70'	C(GSC-5571)	No OSC	10100	110	BASE	9293

TABLE I (Continued)

Radiocarbon ages, core depths and basal model age of lakes in Nova Scotia and New Brunswick

No.	Location	Latitude N/ Longitude W	Laboratory Age/Date/Type	Depth Below YD (cm)	Date	2s	Total Thick. (Unit 2)	Basedate (Age Model)
24.	Folly Lake	45033.0' 63033.0'	C(P-951)	No OSC	10764	101	BASE	9957
25.	Hector Lake	45039.13' 61021.75'	A(TO-3975)	20.0	11910	130	25.0	12085
25.	Hector Lake		C(GSC-5283)	25.0	13400	170	25.0	12085
26.	Gillis Lake	45039.0' 60046.0'	C(GSC-4230)	17.0	12000	130	170.0	13204#
26.	Gillis Lake		C(GSC-4246)	168.0	13900	190	170.0	13204#
27.	Chase Lake	45039.08' 60040.50'	A(TO-2327)	3.0	11260	130	47.0	13317
27.	Chase Lake		A(TO-2326)	10.0	11410	85	47.0	13317
27.	Chase Lake		A(TO-2325)	25.0	12570	130	47.0	13317
27.	ChaseLake		A(TO-2324)	50.0	14010	140	47.0	13317
28.	Wigmore Lake	45044.0' 63038.14'	C(GSC-5567)	10.0	11400	140	20.0	11805
29.	Chance Harbour Lake	45046.40' 62033.95'	C(GSC-4267)	13.0	12100	120	20.0	11805
29.	Chance Harbour Lake		C(GSC-4328)	13.0	13400	160	20.0	11805
30.	Baddeck Bog	4606.75' 60046.33'	C(GSC-4865)	No OSC	9100	100	BASE	9100
31.	Third O'Law Lake		C(GSC-4414)	No OSC	12400	130	BASE	11593
32.	Timber Lake	46022.77' 60039.92'	C(GSC-5259)	No OSC	11200	200	BASE	10393
33.	Cormier Lake	46029.0' 6104.01'	A(BETA-61401)	No OSC	9970	112	BASE	9970
33.	Cormier Lake		C(GSC-5275)	No OSC	10700	170	BASE	9970
34.	MacInnes Lake	46029.0' 60026.81'	C(GSC-4656)	No OSC	10900	110	BASE	10003
35.	Pembroke Lake	46029.77' 60059.75'	C(GSC-5185)	No OSC	10700	190	BASE	9893
				New Brunswick Lakes used in this analysis				
36.	Mayflower Lake	45018.17' 66004.25'	A(TO-1948)	0.0	10880	150	21.0	11861
36.	Mayflower Lake		A(TO-2351)	7.0	10560	140	21.0	11861
36.	Mayflower Lake		A(TO-1947)	16.0	12030	150	21.0	11861
37.	Joe Lake	46045.42' 66040.08'	A(TO-2302)	7.0	10800	99	32.0	12477
37.	Joe Lake		A(TO-2303)	19.0	11110	120	32.0	12477
38.	Fredericton Bog	45056.0' 66041.0'	C(GSC-4806)	26.0	11400	120	72.0	11770#
38.	Fredericton Bog		C(GSC-4778)	67.0	12300	180	72.0	11770#
39.	Long Lake	45018.7' 6603.6'	C(GSC-4544)	15.0	12200	150	45.0	13205
40.	Splan Lake	45015.33' 67019.83'	C(GSC-1645)	1.0	11300	180	59.0	13695
40.	Splan Lake		C(GSC-1067)	22.0	12600	270	59.0	13695
40.	Splan Lake		A(TO-1928)	22.0	11640	120	59.0	13695
40.	Splan Lake		A(TO-1930)	0.0	10690	112	59.0	13695

Radiocarbon age date types: A-AMS C-Beta Decay. Brier Island basal age date (asterisk) determined from marine shells. Basal ages in "kettle" lakes (# symbol) determined by linear extrapolation in Unit 2 using Little Dyke Lake sedimentation rate (0.018 cm/year). Depth below YD refers to the depth below the Younger Dryas inorganic marker (Unit 2/3 boundary) where the age date was obtained. No Osc: TOC or MS oscillation was not noted. Total "thick" refers to thickness of gyttja (Unit 2) underlying YD marker. "Base" in YD column denotes age dates from the base of lakes without an TOC or MS "oscillation". The basal age of these lakes dated by Beta decay, were calculated by substracting 807 years (difference between regression intercepts of beta decay vs. AMS radiocarbon ages). Note: Several radiocarbon ages above the YD marker (shown on Fig. 2) are not summarized on this table.

AMS ages for Lakes 3, 27, 36, 40 from Mayle *et al.* (1993).

AMS ages for Little Dyke Lake from Frappier (1996).

AMS age for Porters Lake (D. B. Scott, pers. comm., 1995).

Beta decay age for Folly Lake from Ogden (1987).

correlated from southern to northern Nova Scotia in selected lake cores. The base of most lake sediment cores feature a distinctive, visible, sedimentological change or marked fluctuation in organic carbon percentages (Mott, 1975; Mott *et al.*, 1986; Mayle *et al.*, 1993). The basal unit (Unit 1) represents the maximum depth of core penetration, usually into clastic sediment, and consists of several lithofacies including coarse and fine-grained mineral sediment. The coarse-grained lithofacies is either sand or gravelly-sand (Fig. 2). The fine-grained lithofacies is generally a laminated clayey-mud, often reddish in hue, with massive mud layers interbedded with thin (.1-1 cm) laminae of silt and sand (Fig. 2). Unit 1 often grades into, or is sharply overlain, by brown or black gyttja, a mixture of inorganic sediment and organic (algal) components (Unit 2). Unit 3 is either a silty gyttja or a mineral sediment which marks the onset of the Younger Dryas (Mott, 1975). In southern Nova Scotia lakes, the onset of the Younger Dryas interval is associated with a marked decrease in TOC values (Fig. 3). In northern Nova Scotia lakes, Unit 3 consists of either a massive mud or a mud with silty interbeds. It can also contain beds or lenses of fine sand (Unit 3- Fig. 2). Unit 3 is overlain by gyttja, (Unit 4) which continues uninterrupted by visible mineral sediment to the top of the core.

High-resolution sampling of Maritime Canada lake cores has revealed organic carbon fluctuations within Unit 2, termed the Killarney Oscillation (Levesque *et al.*, 1993). Sampling frequency in this study was insufficient to resolve higher frequency "events" within Unit 2.

Significant latitudinal and elevational variations are apparent in the stratigraphic style of the basal part of the lake sediment cores. Lakes in Nova Scotia can be classified on the basis of presence or absence of a mineral sediment (MS) or TOC "oscillation" between mineral-rich Units 1 and 3 and organic-rich Units 2 and 4 (Stea *et al.*, 1992; Figs. 1, 2, 3). The basal oscillation is present in most lowland lake cores, but absent in many lakes in highland regions of northern Nova Scotia (Fig. 1).

The south to north lake basin stratigraphic transect and data from the rest of the lakes studied (Figs. 1, 2; Table I) reveal regional stratigraphic variations including:

1 . A south to north decrease in the thickness and organic content of the pre-Younger Dryas gyttja sequence (Unit 2).

2. A concomitant south to north increase in the thickness and mineral content of the Younger Dryas marker horizon (Unit 3). This unit either increases in thickness such that in northern Nova Scotia it is too thick to be penetrated by conventional coring devices, or is absent due to non-deposition.

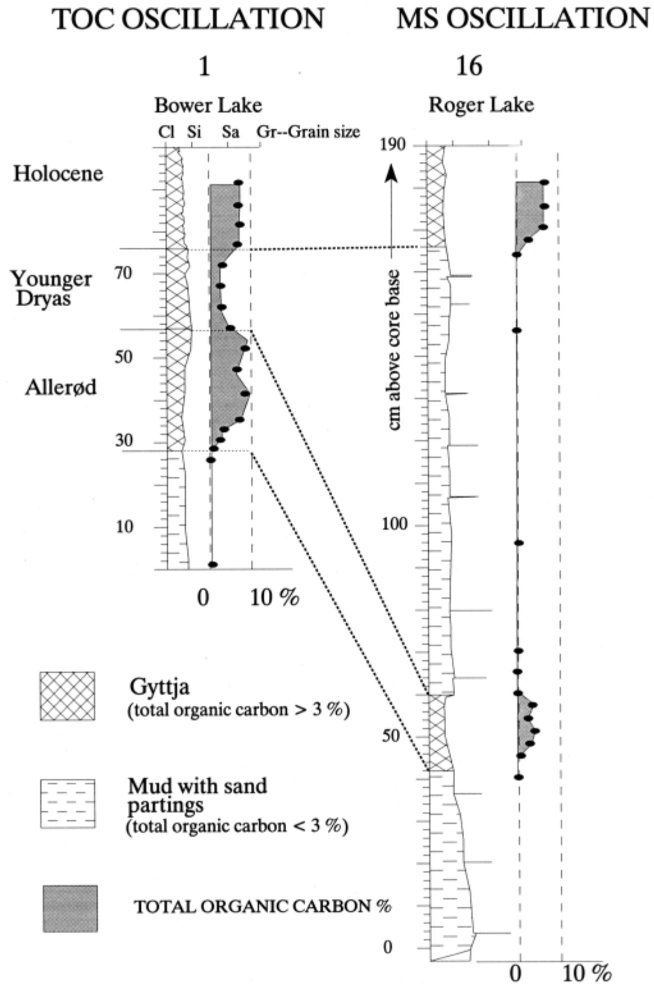

FIGURE 3. Plot of total organic carbon (TOC) values versus depth and visual stratigraphy for two lake cores in southern and northern mainland Nova Scotia. The Younger Dryas interval is discernable by a noticeable drop in TOC values in many lakes in southern Nova Scotia, whereas, a distinct inorganic (mineral sediment: MS) interval is present in some northern Nova Scotia lakes. The oscillation characterized by a drop in TOC values is known as a TOC oscillation, the mineral sediment layer as "MS" oscillation.

Courbes des valeurs de carbone organique total (TOC) sur la profondeur et stratigraphie de deux carottes lacustres en Nouvelle-Écosse continentale septentrionale et méridionale. L'intervalle du Dryas récent se réflète par une baisse des valeurs de carbone dans de nombreux lacs du sud de la Nouvelle-Écosse, tandis qu'un intervalle inorganique (sédiment minéral) apparaît dans quelques lacs du nord. L'oscillation caractérisée par une baisse des valeurs de carbone est appelée « TOC oscillation *» et la couche de sédiment minéral est appelée «* MS oscillation *».*

MINIMUM DEGLACIATION AGE OF LAKE BASINS

Fifty beta decay and AMS radiocarbon ages were obtained from Unit 2 beneath the Younger Dryas marker horizon (Unit 3; Table I; Fig. 1). Beta decay gyttja ages were obtained from the lowest organic interval exceeding 10% organic content. AMS ages were obtained wherever a twig or plant fragment was available from various depth intervals above the inorganic base of the lake sediment core. To obtain valid "minimum" ages for deglaciation the age data must be extrapolated to the base of Unit 2 using sedimentation rates. The consistency of sedimentation rates between lake basins has to be assessed before age estimates can be made..

Sedimentation rates are affected by lake basin morphology. A histogram of Unit 2 (gyttja) thicknesses beneath the Younger Dryas "marker" horizon demonstrates this effect.

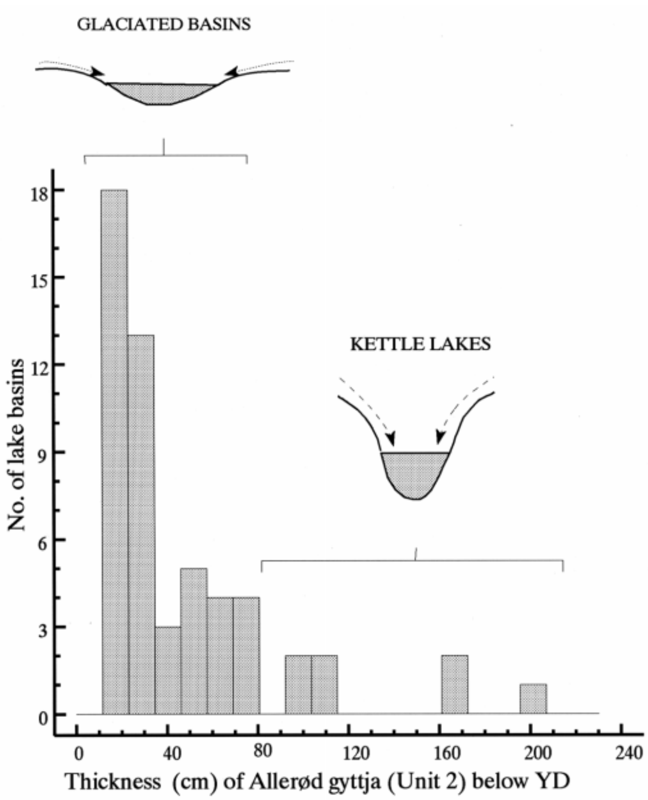

FIGURE 4. Histogram of total gyttja thickness in Unit 2 beneath the Younger Dryas marker horizon (Unit 3). Two lake types can be differentiated based on sedimentation rates: glaciated basins with Unit 2 gyttja thicknesses less than 80 cm and "Kettle" lakes with late-glacial gyttja thicknesses > 80 cm.

Histogramme de la profondeur totale de gyttja dans l'unité 2, sous la limite de l'horizon du Dryas récent (unité 3). On distingue deux types de lacs fondés sur le taux de sédimentation : les bassins englacés avec des profondeurs de gyttja (unité 2) de moins de 80 cm et les lacs de type kettle avec des profondeurs de gyttja tardiglaciaire supérieures à 80 cm.

Two discrete populations (Table I; Fig. 4) are present reflecting variations in the thickness of Unit 2 (the basal organic sequence) associated with lake basin shape and topographic setting. Most lakes represented by the 10-80cm thickness interval are in muted topographic settings, with gently-sloping margins and shallow basins that discourage sediment influx. These lake basins are defined as glaciated basins. Lakes where Unit 2 is >80 cm thick generally lack a distinct break between Units 1 and 2 and contain a thick sequence of slightly organic (1-3% TOC) sediment, diluted with mineral sediment or carbonates. Examples are Sandy Cove Lake, Little Dyke Lake, Leak Lake, and Gillis Lake (sites 6, 18, 20, 26; Fig. 1), which with the exception of Gillis Lake are steep sided, kettle lakes, surrounded by outwash or ice contact stratified drift (Fig. 5). It is therefore assumed that organic sedimentation rates are similar within these broad lake classifications, and Unit 2 thickness is an indication of duration of deposition.

DEGLACIATION AGE MODEL

Ogden (1987) used a core depth-age regression technique to determine the age of first accumulation of organic lake sediments in Nova Scotia and hence a "minimum" age of deglaciation. He used beta decay basal radiocarbon ages on gyttja from lakes throughout Nova Scotia and accommodated varying core depths by normalizing core depth to percentages. Using this regression method, and assuming regional synchroneity of ice retreat, he obtained a general age for deglaciation of 12 ka. Age variation resulting from varying sedimentation rates, and dating error were mitigated by this regression technique, but ice retreat rates could not be established. In our study, absolute (not relative) core depth or gyttja thickness is used. Regional variation in timing of deglaciation can be established using varying gyttja thicknesses, assuming that gyttja thickness is related to time elapsed since ice removal, sedimentation rates between cores are similar, and that the Younger Dryas marker is roughly synchronous throughout the region.

Figure 6 shows the results of the linear regression analysis using both beta decay and AMS ages against gyttja thickness or core depth below the Younger Dryas marker horizon (Unit 3). The deglaciation age model for AMS ages developed by this analysis is (Fig. 6):

1. Age= 10685+(D X 55.99)

D = Depth below Younger Dryas boundary (Units 2/3)

There are strong linear correlatons between sample depth below the Younger Dryas boundary (base of Unit 3) and radiocarbon ages for both AMS analysis (R^2=0.89, n=17, Line 2-Fig. 6) and beta decay ages (R^2 =0.75, n=17; Line 1-Fig. 6). The YD marker horizon intercept value of 10,656 yr BP is well within uncertainty limits of the YD inception times of 10.8 ka quoted by Mott and Stea (1994) based on radiocarbon ages on 26 buried wood and peat sites. The intercept value of 11,492 yr BP is 806 radiocarbon years older than the AMS intercept. The two regression lines (1,2 Fig. 6) are nearly parallel, indicating a consistent error that is not solely related to decreasing core depth and organic carbon content. This depth-independent result indicates that "hard water" effects are prevalent, related perhaps to groundwater flux through the substrate (Fritz, 1983) rather than early deglacial meltwater effects (Sutherland, 1980). The sedimentation rate for glaciated basins using the linear regression model is 0.014 cm/yr.

Lakes without oscillations are not included in the age model (Table I). Unit 4 (Fig. 2) in these lakes has higher organic content and can be sampled quite close to the actual base of the core. Minimum deglaciation ages were estimated from these basal ages, because sedimentation rates for Unit 4 are not available. Eight hundred and six years (the systematic age discrepancy between pre-YD gyttja beta decay and AMS radiocarbon ages-Fig. 6) were subtracted from beta decay radiocarbon ages older than 10,000 years to obtain the basal age (Table I).

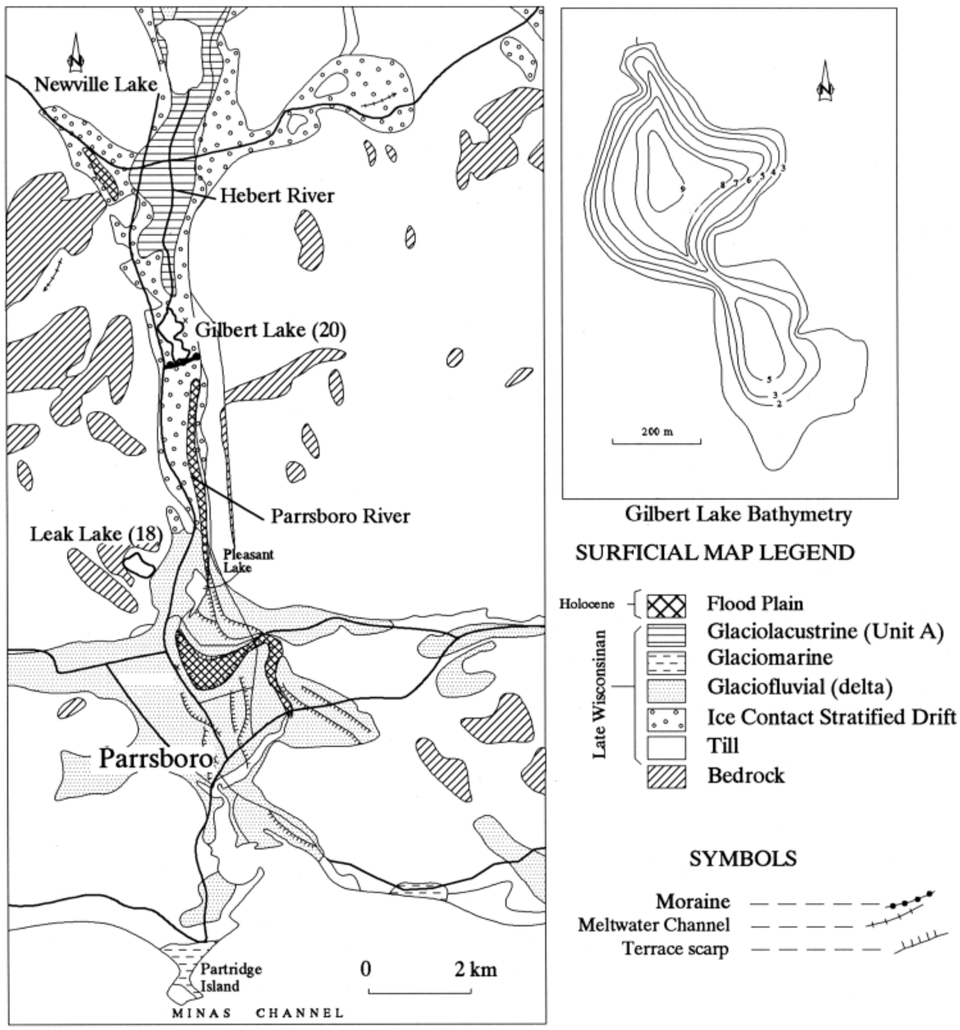

FIGURE 5. Surficial geology of the Parrsboro Valley, Nova Scotia (Fig. 1). Ice margins are located north of Leak Lake (ice contact head) and at the southern end of Gilbert Lake (moraine). (After Wightman, 1980; Stea *et al.*, 1986.) Bathymetry of Gilbert Lake (metres) from Wightman (1980).

Géologie superficielle de la vallée de Parrsboro. Les marges glaciaires sont situées au nord de Leak Lake (tête du contact glaciaire) et à la partie inférieure du Gilbert Lake (moraine) (d'après Wightman, 1980; Stea et al.*, 1986). La bathymétrie du Gilbert Lake (en mètres) est de Wightman (1980).*

Two AMS ages were obtained from Little Dyke Lake (Frappier, 1996; 17; Fig. 1) at 23 and 35 cm below the Younger Dryas (Unit 2/3) boundary. The basal age of this kettle lake was rather crudely established by simple linear extrapolation to the base of Unit 2 from two age-depth points (Fig. 6). By this method, the sedimentation rate for Little Dyke Lake is 0.018 cm/year. Other kettle lakes contained twig fragments within Unit 2, and were sampled at or close to the base of this unit. The basal ages of these lakes were extrapolated using the Little Dyke Lake sedimentation rate. These deglaciation estimates are in all likelihood, older than the actual time of ice removal because of the nonlinearity of sedimentation rates towards the base of the cores. This age overestimation is mitigated somewhat by the time lag between deglaciation and lake organic sedimentation.

A core in Brier Island Bog in southern Nova Scotia (Figs. 1, 2) bottomed in sand containing marine shells that were radiocarbon dated at 13,200 yr BP (Table I). This date is used as the minimum deglaciation age for the local area. A standard marine reservoir effect correction of 400 radiocarbon years (Mangerud, 1972) would suggest deglaciation before 13.6 ka.

SECTIONS WITH BURIED ORGANIC DEPOSITS: ANOTHER DEGLACIATION RECORD

In previous publications (Stea and Mott, 1989; Mott and Stea, 1994) 23 sites have been documented in Nova Scotia that feature peat and wood buried by sediment formed by mass wasting and glacigenic processes. In central and northern Nova Scotia (Figs. 7, 8) several crucial sections record the retreat of ice and the formation of glacial lakes, during initial deglaciation and the subsequent Younger Dryas cold stage. These sections can be used as an independent check on the AMS lake basin chronology of ice retreat. Six informal lithostratigraphic units (Units A-F) can be defined within the late-glacial sequence.

At the base of late-glacial sediment sections in the Shubenacadie valley of central Nova Scotia is a massive silty clay or rhythmically-bedded clay-silt (Unit A). In the southern part of the valley, Unit A (Sections 1 and 2, Figs. 7,

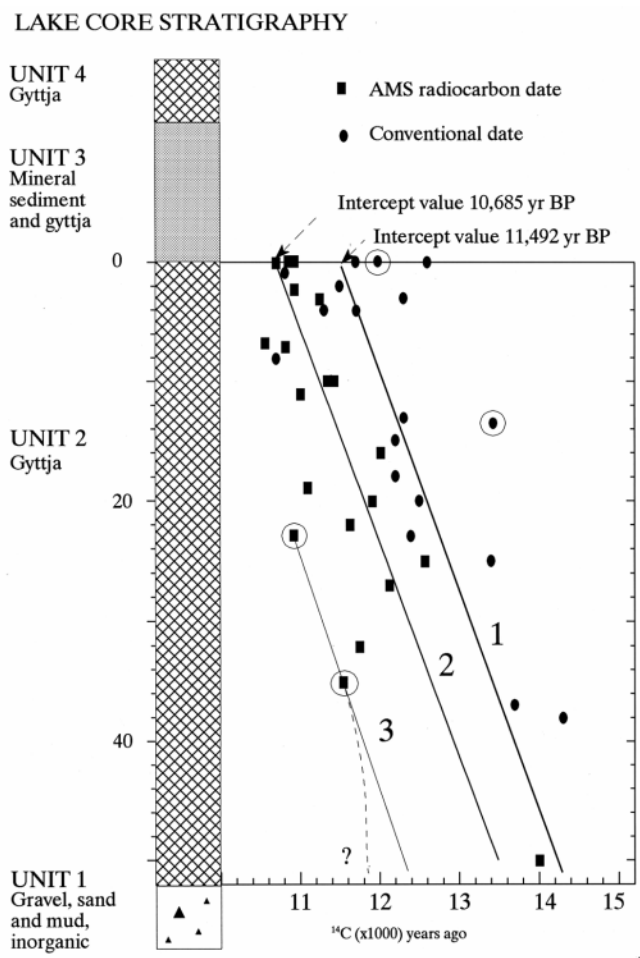

FIGURE 6. Linear regression analysis of AMS and beta decay radiocarbon ages and total depth below the base of the Younger Dryas marker horizon (Unit 2/3 contact). Dark ovoid dots are beta decay ages. The circled dots denote data from a marl lake (Chance Harbour Lake, Jetté and Mott, 1989) excluded from the analysis. The squares are AMS ages. Circled squares were obtained from Little Dyke Lake, for a crude estimate of sedimentation rates for a kettle lake. Line 1=beta decay regression model. Line 2=AMS regression model. Line 3=Little Dyke Lake extrapolation.

Analyse en régression linéaire des dates au radiocarbone conventionnelles et par spectométrie de masse et profondeur totale sous la base de l'horizon du Dryas récent (contact des unités 2 et 3). Les points ovales représentent les dates conventionnelles. Les points encerclés représentent des données en provenance d'un lac à dépôts marneux (Chance Harbour Lake, Jetté et Mott, 1989) exclu de l'analyse. Les carrés encerclés proviennent de Little Dyke Lake d'où on a obtenu une estimation approximative du taux de sédimentation dans un lac de type kettle. Ligne 1 : modèle de régression pour date conventionnelle ; ligne 2 : modèle de régression pour date obtenue par spectométrie de masse; ligne 3 : extrapolation à partir du Little Dyke Lake.

8) is brown, fine-grained and massive, becoming distinctly laminated to the north, with sand lenses and occasional boulder-sized dropstones. (Section 14; Figs. 7, 8). Hughes (1957) mapped these clay deposits sporadically at elevations below 30m in the Shubenacadie valley (Fig. 9). The senior author, in recent mapping, has found that glaciolacustrine facies are found in virtually all regions below 30 m in the Shubenacadie Valley.

Unit B is a massive, coarse to medium-grained, moderately-sorted sand which overlies Unit A at most lowland sections (Fig. 7). It is overlain by Unit C comprising peat, or organic-bearing layers, that are dated between 11.9 and 10.5 ka. The basal age of peat accumulation decreases northward from 11.8 to 11.4 between Lantz and Shubenacadie (Fig. 7). At Brookfield (14, Fig. 1) an organic silty clay, was found containing twigs (*Salix*), dated 10.8 ka by AMS (Table I). Lismore, although further north, records peat deposition between 11.9 ka and 10.5 ka. The Judique section, in Cape Breton Island, (Fig. 1; 6-Fig. 7) is located in a fluvial terrace (Grant, 1994) at about 10 m elevation. Here, only a thin, organic bed with a few twigs was exposed in a thick massive to finely laminated silty-clay sequence (Fig. 7).

The Wentworth Valley section is located in a glaciated valley that bisects the Cobequid Highlands (Fig. 1; 5-Fig. 7; Fig. 8). At the base of the section (Unit B, Fig. 7) is a gravelly-sand overlain by Unit C, an organic sand unit with wood fragments (*Picea, Salix*). One of the *Salix* fragments was dated at 10,700 +-100 yr BP (GSC-4929). Above this organic bed is 1-3 m of of sand and clay rhythmites (Unit D; Figs. 7, 8). The upper part of Unit D is deformed into irregular, small-scale folds and is capped by an undated 1 cm thick organic seam. Unit D grades laterally into a clay diamicton containing striated and facetted clasts and is overlain by coarse sediments consisting of pebbles and boulders in a coarse sand matrix (Unit F). Peat and modern fill cap the sequence (Unit G).

DISCUSSION

DEPOSITIONAL ENVIRONMENTS INFERRED FROM LAKE SEDIMENTS AND ORGANIC SECTIONS

Basal inorganic sediments of the late-glacial sections and lakes (Units A and 1 Figs. 2, 7) are interpreted as glacial lake deposits with facies of subaqueous deposition including clay-sand couplets. These laminae may represent coarse, summer underflow layers and finer, winter, suspension rainout layers (Ashley, 1975). Gravelly-sand and sand deposits (Unit B) are fluvial and probably represent the re-establishment of graded streams after lake outlets were freed of ice. The transition from inorganic Unit 1 to organic Unit 2 in lake cores represents the diversion of meltwater into alternative routes as divide headwaters were freed of ice (Rogers and Ogden, 1991).

Sections containing buried Allerød peats (Unit C) are mostly found in the vicinity of lake basins with a MS oscillation (Fig. 1). Southern Nova Scotia, whose lakes are characterized by TOC oscillations appears to be devoid of these buried peat sections in spite of mapping scrutiny (Grant, 1980). This region was deglaciated earliest. Organic beds (fens, bogs; Unit C) formed in low lying areas during a period of climatic warming after drainage of ice peripheral glacial

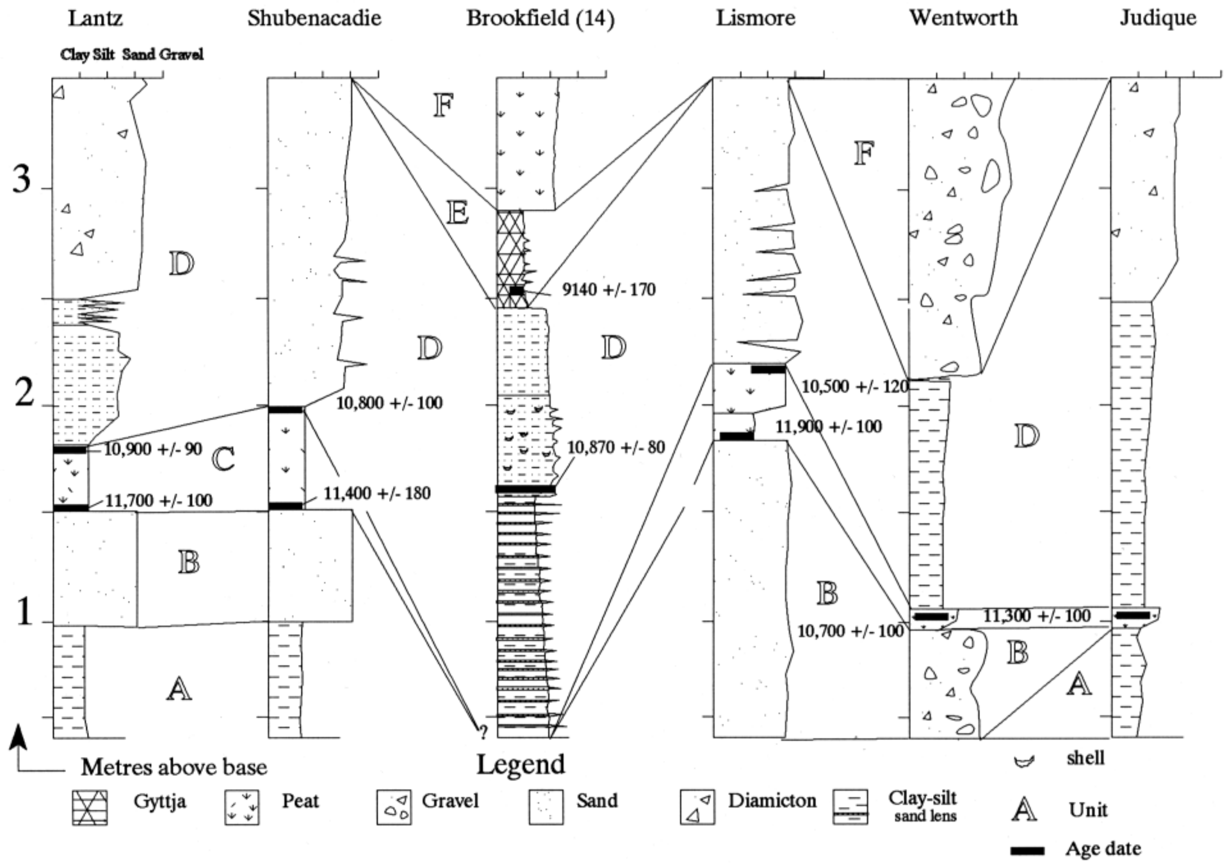

FIGURE 7. Lithofacies diagram indicating stratigraphic correlations and chronology of selected late-glacial buried peat and organic sections in Nova Scotia.

Diagramme des lithofaciès montrant les corrélations stratigraphiques et la chronologie de certaines coupes organiques et de tourbe tardiglaciaire enfouie, en Nouvelle-Écosse.

lakes. These bogs probably received inorganic and organic material from runoff from formerly vegetated slopes (Stea and Mott, 1989).

The organic sediments in lakes and bogs (Units 2 and C) were covered by inorganic sediment (Units 3 and D). The variety of sedimentary lithofacies and architectures and the patchy occurrence of these sites implies complexity of processes (Figs. 7, 8). Massive and laminated silty clays and clay-rich diamicton lithofacies of Unit D are found at Lantz, Brookfield, Judique and Wentworth (Fig. 7). The lithofacies association is consistent with a glaciolacustrine or glaciomarine origin (Eyles and Eyles, 1992). At Shubenacadie, coarse, horizontally-laminated, moderately well-sorted sand lithofacies of Unit D was originally interpreted as outwash (Hughes, 1957; Stea and Mott, 1989). Regional mapping indicates that Unit D sand deposits are found at or below 30 m (Fig. 9) flanking the Shubenacadie Valley and lack gravel, crossbedding and cut and fill structures, and these deposits grade into a fine-grained lithofacies at lower elevations such as at Lantz (Figs. 1, 7; Stea and Mott, 1989). A lacustrine origin is implied, and the sand deposits probably formed at the shoreface of a former glacial lake. Marine submergence during the Younger Dryas is unlikely as a sea-

FIGURE 8. Photographs of late-glacial sections. (A) Basal lake sediment core (base to left) (Pye Lake 10-Figs. 1, 2) showing mineral sediment "oscillation". Units 1-4, (Fig. 2) marked on photo, Units 1 and 3 (mineral sediment) marked by blue dots, and Units 2/4 (organic sediment) marked by red dots. (B) Intensely deformed, compacted peat beds overlain by diamicton, Collins Pond, Nova Scotia (Fig. 1; Stea and Mott, 1989; Mott and Stea, 1994). (C) Blomidon section (Figs. 1, 7). Late-glacial peat overlain by sand interpreted as lacustrine deposit. (D) Wentworth Section (Figs. 1, 7) 1 metre of glaciolacustrine, laminated clay (Unit D) overlying late-glacial wood and peat. (E) Lismore section (Figs 1. 7). Laminated silt and sand overlying late-glacial peat. (F) Lantz section (Figs 1, 7). Massive glaciolacustrine clay (Unit D) overlying peat.

Photographies de coupes tardiglaciaires. A) Carottes lacustres basales (base à gauche) (Pye Lake 10, fig. 1 et 2) montrant l'« oscillation » du sédiment minéral. Les unités 1 à 4 sont identifiées, les unités 1 et 3 (sédiment minéral) sont identifiées par les points bleus et les unités 2 et 4 (sédiment organique) sont identifiées par les points rouges. B) Couches de tourbe très déformées recouvertes par un diamicton, au Collins Pond (fig. 1; Stea et Mott, 1989; Mott et Stea, 1994). C) Coupe de Blomindon (fig. 1 et 7) : tourbe tardiglaciaire recouverte d'un sable considéré comme un dépôt lacustre. D) Coupe de Wentworth (fig. 1 et 7) : 1 m d'argile glaciolacustre laminée (unité d) recouvrant du bois et de la tourbe tardiglaciaire. E) Coupe de Lismore (fig. 1 et 7) : limon et sable laminés recouvrant une tourbe tardiglaciaire. F) Coupe de Lantz (fig. 1 et 7) : argile glaciolacustre massive (unité D) recouvrant de la tourbe.


A
4
3
2
1
CORE 6
PYE LAKE, N.S.
91-MS-05
CORE 7
B
C
D
E
F

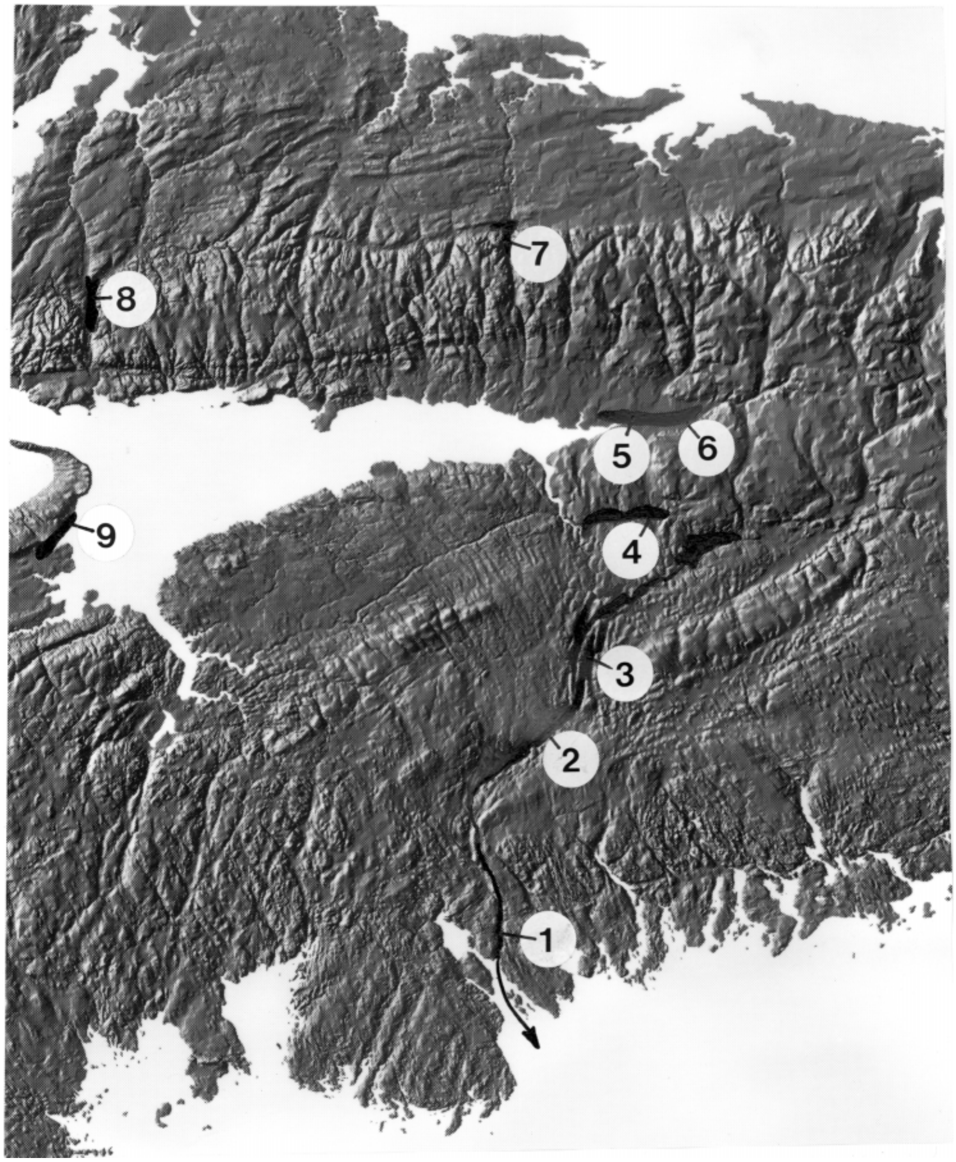

FIGURE 9. Late-glacial lakes in central Nova Scotia (in black). (Digital terrain model of Nova Scotia courtesy of Tim Webster, College of Geographic Sciences, Lawrencetown Nova Scotia). 1. Glacial Lake Shubenacadie outlet, Lake William-Lake Charles system. Lake drainage to the Atlantic Ocean through Halifax Harbour. Sill at 30 m elevation. 2. Location of Lantz site (Fig. 7). 3. Shubenacadie site (Fig. 7). 4. Brookfield site (Fig. 7). 5. Truro diamicton site (Stea and Mott, 1989). Retreating ice blocked the mouth of the Shubenacadie River (east of 5), and formed Glacial Lake Shubenacadie between 12.5 and 11.7 ka. A lake of comparable size reformed after Younger Dryas glaciers reoccupied the Minas Basin ca 10.8 ka. 6. The Brookside site indicates an ice marginal lake in the Salmon River Valley west of Truro during the Younger Dryas (Mott and Stea, 1994). 7. The Wentworth valley site in the Cobequid Highlands (Fig. 7). 8. Parrsboro Valley (Fig. 5). 9. Blomidon site (Figs. 7, 8.) Lake problematic.

Lacs tardiglaciaires du centre de la Nouvelle-Écosse (en noir) (modèle numérique de la Nouvelle-Écosse gracieusement offert par Tim Webster du College of Geographic Sciences, Lawrencetown, Nouvelle-Écosse). 1. Exutoire du Lac glaciaire Shubenacadie, réseaux des lacs William et Charles. Écoulement vers l'Atlantique par le port de Halifax (seuil à 30 m d'altitude). 2. Localisation du site de Lantz (fig. 7). 3. Site de Shubenacadie (fig. 7). 4) Site de Brookfield (fig. 7). 5) Site du diamicton de Truro (Stea et Mott, 1989). La glace en retrait a obstrué l'embouchure de la Shubenacadie River (à l'est de 5) et a formé le Lac glaciaire Shubenacadie entre 12,5 et 11,7 ka. Un lac de taille semblable s'est reformé après que les glaces du Dryas récent ait occupé le bassin de Minas vers 10,8 ka. 6) Le site de Brookside présente un lac de marge glaciaire, au cours du Dryas récent dans la vallée de la Salmon River à l'ouest de Truro (Mott et Stea, 1994). 7) Le site de la vallée de Wentworth dans les hautes terres de Cobequid (fig. 7). 8) La vallée de Parrsboro (fig. 5). 9) Le site de Blomidon (fig. 7et 8.) : un lac problématique.

level lowstand (-60m below sea level) is documented in the Gulf of Maine at around 10.5 ka (Belknap *et al.,* 1987; Stea *et al.,* 1987; Barnhardt *et al.,* 1995).

How were these lakes impounded?. We propose that the buildup of permanent snow and/or glacial ice created and enlarged already existing glacial lakes by damming regional drainage pathways. In southern Nova Scotia thin silty-sand and gravel lithofacies of Unit D may have been transported by surface meltwater released from snowfields. Similar deposits in southwest England have been termed "Niveo-Fluvial" (Kerney *et al*., 1964). The Nictaux Falls site, found on the north-facing slope of the Annapolis Valley is overlain by a till-like diamicton and thick mound of gravelly-sand, suggesting a nearby source of ice (Nielsen, 1976; Mott and Stea, 1994; Fig. 1). In northern Nova Scotia, snow accumulation was sufficient to form larger bodies of permanent snow or to reinvigorate glaciers. Bodies of aufeis may have formed and persisted in lowlands, fed by melting snow and groundwater seepage (Veillette and Thomas, 1979). Larger glacial lakes were created by ice blocking former outlets such as the mouth of Shubenacadie River (Fig. 9). Fine-grained sediments were deposited (Unit D) in these glacial lakes.

The development of a lake in the Wentworth Valley (Fig. 9) is only possible if free drainage to the Northumberland Strait had been blocked by ice covering the lowlands north of the Wentworth Valley (Fig. 9). Clay lithofacies of Unit D outcrop sporadically along the north flank of the Cobequid Highlands suggesting an extensive ice-dammed lake system, perhaps draining westward and northward out of the River Hebert system into Chignecto Bay (Figs. 5, 9). The southernmost site in Nova Scotia (Hirtles Beach; Fig. 1) features an unconsolidated organic and clay-rich diamicton, probably a solifluction deposit, overlying peat in a hollow between drumlins (Miller, 1995). Mass wasting processes were prevalent in the transitional areas between permanent snowfields to the north and seasonally frozen ground to the south.

Direct evidence for glacier activity during the Younger Dryas has been found in northern regions where till and glaciotectonically-deformed peat overlies Allerød peat (Collins Pond; Figs. 1, 8; Stea and Mott, 1989; Mott and Stea, 1994; Stea *et al.*, 1996). Manassette Lake (Table I; 19; Fig. 7) is only one kilometre from the Collins Pond site. This lake does not record an obvious mineral or TOC oscillation, except for a slightly organic seam in the lowermost clay (Unit 3; Fig. 2). The thick lacustrine clay deposit (Unit 3) likely formed just after the rapid and short-lived Collins Pond advance which produced a till and deformed peat at Collins Pond (Figs.7, 8; Stea *et al.,* 1996). Many lakes in the upland regions of northern Nova Scotia, and Prince Edward Island (Anderson, 1985) do not record the basal MS or TOC oscillation, and are presumed to have formed after the retreat of Younger Dryas ice, *ca.* 10 ka.

LAKE SEDIMENTATION DURING THE YOUNGER DRYAS

Lake basins in northern Nova Scotia lowlands generally record a significant MS oscillation (Fig. 1). Mott *et al.* (1986) proposed that the flux of inorganic sediment was a result of mass wasting from the drainage basin during the Younger Dryas, as plant cover decreased. It was not clear, however, what would cause long term destabilization of plant cover. Levesque *et al.*, (1994) and Cwynar and Levesque (1995) describe lakes in southwest New Brunswick and Maine, where gravel and coarse sand were deposited during the Younger Dryas interval, and invoked intense erosion, without elaborating on the process. The influx of inorganic sediment into the lake basin is likely a combination of destabilization of plant cover, and sediment-laden meltwater, entraining and transporting coarse clastic sediments into the lake basins. Persistent ice or permanent snow cover and seasonal melting must be essential in the formation of this mineral sediment layer; both to provide the water source and prevent plant re-colonization. The similarity between inorganic lake sediments deposited around 13-12 ka (when ice was still extant) and the Younger Dryas indicates that meltwater was involved. Inorganic sedimentation cannot be solely explained by cooling-related plant cover destabilization. Continuous gyttja accumulation has been recorded in Nova Scotia lake basins which have experienced vegetation removal due to forest fires in the Holocene (Green, 1981). Suppression of organic activity during the Younger Dryas due to cooling alone is unlikely, as gyttja is presently forming in high arctic lakes (Blake, 1981). The presence of a snow-bed pollen taxon (*Empetrum*) was noted in the mineral layer of lakes in Newfoundland (Wolfe and Butler, 1994). The marked decrease in diatoms in a New Brunswick lake during the Younger Dryas (Rawlence, 1988) may be interpreted as a response to permanent ice cover, although a southern Nova Scotia lake did not exhibit a similar response (Wilson *et al.*, 1993).

In northern Nova Scotia lakes with MS oscillation record YD sedimentation rates around 0.19 cm/year (Fig. 2), an order of magnitude increase from Allerød rates of sedimentation (0.018 cm/year). Southern Nova Scotia lakes with a TOC oscillation (Bower Lake for example; Fig. 2) also record higher sedimentation rates during the Younger Dryas. Notwithstanding these data, Spooner (1997) noted a decrease in sedimentation rates in a southern Nova Scotia lake, and little variability in quartz grain size between Allerød and Younger Dryas sediments. He proposed that during the Younger Dryas organic sedimentation was merely suppressed, and no ice or snow was extant. The results of this study and others, however, (Mayle *et al.*, 1993; Wilson *et al.*, 1993; Levesque *et al.,* 1994; Mayle and Cwynar; 1995) clearly indicate sedimentation rate increases during the Younger Dryas interval. The discrepancy between these studies may be explained by proximity to residual ice and snow, whereby significant sedimentation rate increases (and mineral sediment flux) were largely recorded in northern Nova Scotia, Maine and New Brunswick (Fig. 1). The MS oscillation, rather than the TOC oscillation records the presence of permanent snow within the drainage basin, and possibly permanent ice cover over the lake itself. Stea *et al.* (1992) inferred the distribution of Younger Dryas ice based on the distribution of lakes with and without an oscillation, without defining the exact nature of this basal oscillation. Subsequent coring of lakes in the Antigonish Highlands (22-Table I; Fig. 1) where Younger Dryas glacial ice was postulated (*cf.* Stea *et al.*, 1992) has revealed a thick MS oscillation. A complete lake stratigraphic record through the late-glacial precludes erosive glaciers, but not permanent snow, with sub-pack meltwater presumably acting as a sediment source. Figure 10 provides a pictorial summary of the glacial and periglacial sedimentary environments of the Younger Dryas in Nova Scotia.

An average inception date of the Younger Dryas based on peat and wood ages from all the organic sections and previously published lake sediment AMS dates is 10.8 ka (Mayle *et al.*, 1993; Mott and Stea, 1994). This compares favourably to the 10.7 ka inception time obtained by linear regression of the AMS lake basin data in this study.

DEGLACIATION OF NOVA SCOTIA

15,000-13,000 yr BP

Figure 11 is a series of paleogeographic reconstructions of ice retreat and sea level change from 15 ka to 10 ka. During initial deglaciation in the Late Wisconsinan (*ca*. 18-15 ka) ice was drawn out of the Gulf of Maine, isolating an ice mass over Nova Scotia which later became an active centre of outflow (Scotian Ice Divide-Ice Flow Phase 3; Stea *et al.*, 1987, 1996). The Scotian Shelf End Moraine Complex (SSEMC) formed at the margin of this glacier (King, 1996). Ice receded

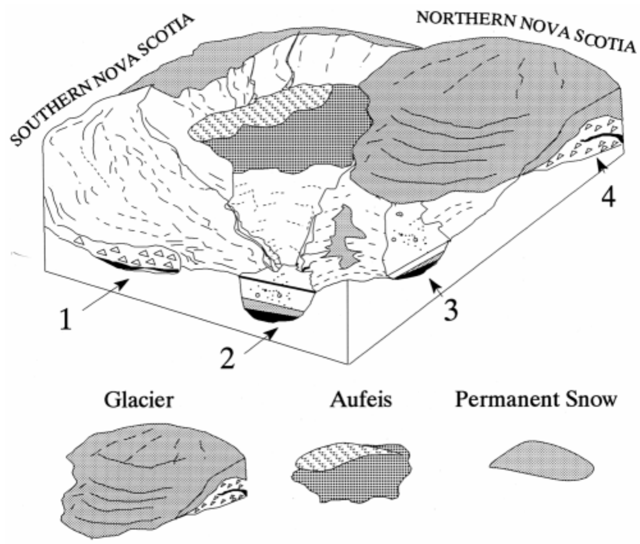

FIGURE 10. Late-glacial sedimentary environments on Nova Scotia.): Mass wasting in a periglacial environment. Colluvial deposits emplaced over peats adjacent to steep slopes (*e.g.* Hirtles Beach; Fig. 1) 2) Lake basins with mineral oscillation, formed adjacent to permanent ice masses/glaciers. 3) Sand/Clay facies deposited in meltwater ponds/lakes over pre-existing bogs (*e.g.* Blomidon; Lismore; Judique; Fig. 1). 4) Subglacial diamicton (*e.g.* Collins Pond; Fig. 1).

Les milieux sédimentaires tardiglaciaires en Nouvelle-Écosse. 1) Mouvement de masse (solifluxion) en milieu périglaciaire. Près des pentes abruptes, les colluvions sont sus-jacentes à la tourbe (Hirtles Beach, fig. 1). 2) Les bassins lacustres ayant connu une oscillation minérale se font formés au droit de masses de glace permanente ou de glaciers. 3) Des faciès de sable ou d'argile se sont déposés dans des lacs ou étangs de fonte glaciaire sur des tourbières pré-existantes (Blomindon, Lismore, Judique ; fig. 1). 4) Diamicton sous-glaciaire (Collins Pond ; fig. 1).

out of marine areas, depositing ice proximal and distal glacial marine sediment (Emerald Silt facies A and B) in the inner shelf basins (King, 1994). By 14 ka, active ice was located on the inner continental shelf off Nova Scotia (Gipp and Piper, 1989; Gipp, 1994; Piper, 1991; Stea *et al.*, 1996) and may have cleared out of much of the Bay of Fundy (Stea and Wightman, 1987). The formation of a calving bay in the Bay of Fundy, rapid ice withdrawal from the Minas Basin and the blocking effect of the North Mountain cuesta, transformed the Scotian Ice Divide into distinct ice caps over the Antigonish Highlands and South Mountain which merged with remnants of ice from the Magdalen Plateau region (Chignecto Glacier; Chalmers, 1895).

The rate of ice retreat after the last glacial maximum was probably modulated by climatic variability at various scales and by fluctuating sea levels. As ice melted, the glacial isostatic DeGeer Sea (1-Fig. 11) inundated the sunken coastal regions and formed beaches such as those at Cape Chignecto along the Bay of Fundy which are now stranded more than 30 m above present sea level (Stea *et al.,* 1987; Grant, 1989).

13-12.5 ka

The lake data from this study and age ages on marine infauna from the DeGeer Sea (Stea and Wightman, 1987; Nicks, 1988; Smith and Hunter, 1989) show that areas adjacent to the Bay of Fundy began to clear of ice around 13.5 ka. Climatic warming gave further impetus to glacier retreat after 13.5 ka (Lowe *et al.,* 1994). In the western Annapolis Valley (Fig. 11), massive, brown clay covers most areas below 35 m. Raised beaches on the adjacent Bay of Fundy coast, are evidence of higher sea levels during the De Geer Sea submergence phase (Stea *et al.,* 1987). Although Bailey (1898) reported marine shells from the Annapolis clay, it has subsequently been interpreted as a glacial lake deposit, based on low pore-water salinity, and lack of marine foraminifera (Ogden, 1980). According to Ogden, the Digby Gut outlet was blocked by a late-melting or re-advancing local glacier. The lake ages provided in this study suggest that the outlet was deglaciated between 14 and 13 ka, and was never overridden by late ice. With both eastern and western valley outlets blocked by ice, an ice-dammed lake would likely have exceeded 35 m in elevation. The correspondence between maximum clay elevation (35 m) and marine limit (40 m) strongly suggests a marine origin for the Annapolis clay.

A series of ice margins were established along the north shore of the Minas Basin and in regions to the north (Figs. 7, 11b). The oldest margin is marked by glaciofluvial and glaciomarine sediments (Five Islands Formation) that form raised and terraced outwash plains and deltas (Swift and Borns, 1967; Wightman, 1980; Stea *et al*., 1986; Fig. 11). AMS radiocarbon ages on the Spencers Island delta (Fig. 1) range from 14,300 yr BP to 12,600 yr BP. (Stea and Wightman, 1987).

A late-glacial ice readvance into the Bay of Fundy from adjacent lowland regions is recorded by striae, eskers and drumlins (MacNeill *in* Prest, 1972; Stea and Wightman, 1987; Ice Flow Phase 4-Chignecto Phase; Stea *et al*., 1987, 1992, 1996) and postdates the Minas Basin outwash deltas (Stea *et al*., 1986). Nicks (1988) proposed that a glacier readvanced over the Sheldon Point Moraine near St. John, New Brunswick, between 12.9 and 12.5 ka. Hickox (1962) inferred a late-glacial readvance from a kame superimposed on a delta at Margaretsville, Nova Scotia (Fig. 1). Stea *et al*., (1987) noted discontinuities in marine limit north and south of this inferred ice margin, and proposed that lower marine limits north of the margin were due to late-melting ice. The Phase 4 advance margin is bracketed by basal accumulation model ages south of Digby (13.2 ka) and northeast and south of Margaretsville (*ca.* 12 ka; Table I).

A cross-valley moraine at Gilbert Lake in the Parrsboro Valley (Fig. 5) has been interpreted as marking a recessional or readvance margin of a late-glacial readvance (Ice Flow Phase 4-Chignecto Phase-Stea *et al.,* 1986; 1992). This late-glacial event is bracketed by basal accumulation model ages from Leak Lake to the south (13.2) and Welton and Gilbert Lakes (10.0 and 11.4 ka; Fig. 1; Table I). On the inner Scotian Shelf, DeGeer moraines that represent the southern

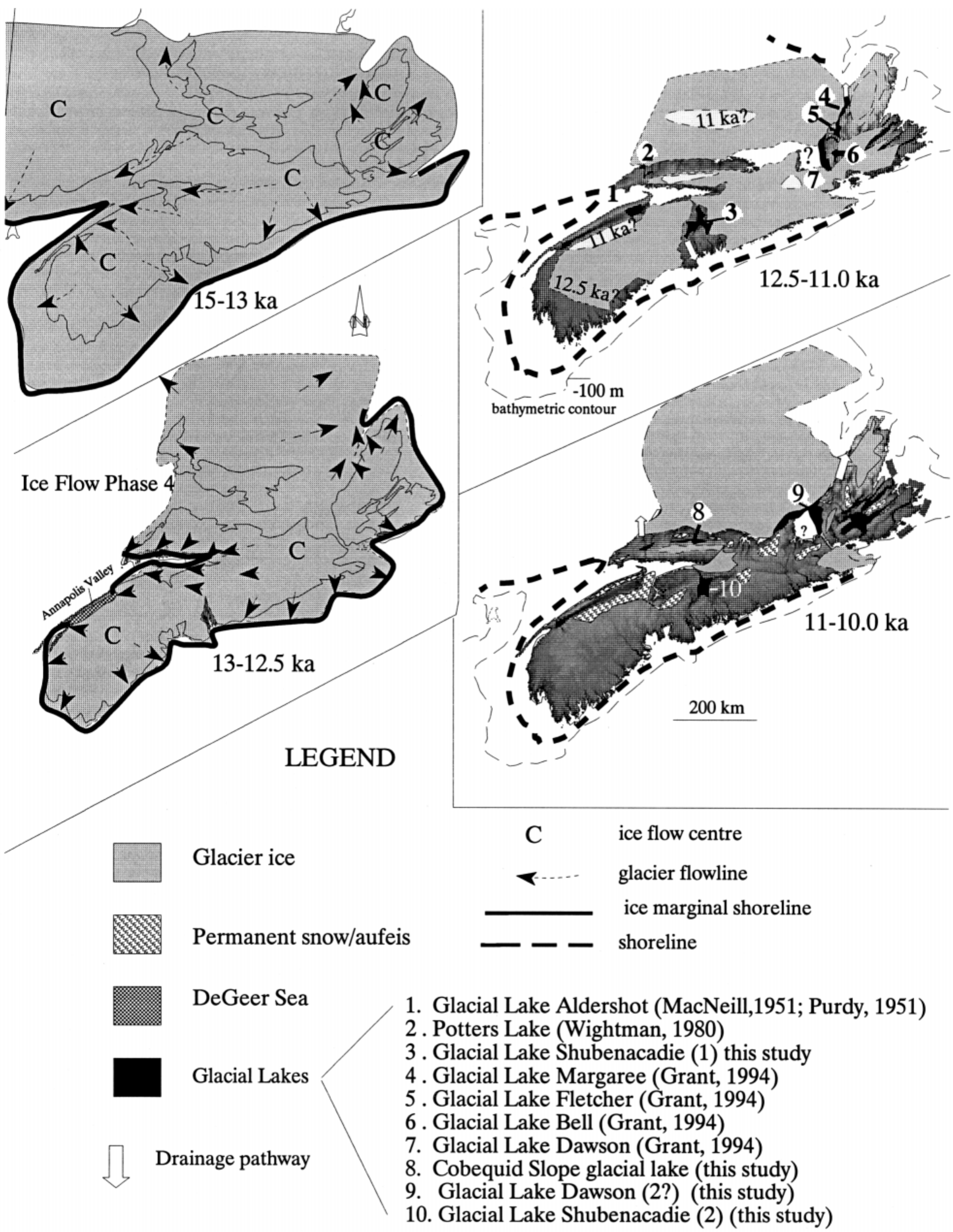

FIGURE 11. Paleogeographic reconstructions of the late-glacial period in Nova Scotia (14,000-10,000 yr BP) (Cape Breton deglaciation after Grant, 1994).

Reconstitution paléogéographique de la période tardiglaciaire en Nouvelle-Écosse (14-10 ka BP). La déglaciation du Cap-Breton est de Grant (1994).

margin of the Antigonish Highland ice cap formed between 14 and 11.6 ka, the time of formation of the Scotian Shelf End Moraine Complex (further seaward) and the lowstand of sea level which destroyed them (Stea *et al.,* 1996). Stea *et al.* (1996) estimated the age of this advance at around 12.8 ka, and considered it a climatic event based on correlative glacial re-advances described in Newfoundland, New Brunswick, and Maine (Borns and Hughes, 1977; Brookes, 1977; Nicks, 1988). These advances may be synchronous with the Port Huron advance of the midcontinent, as Borns and Hughes (1977) first suggested, and may have northwest Atlantic counterparts (Elverhôi *et al.,* 1995). At this time in deep ocean cores off the Scotian Shelf, a cooling signal was recorded in foraminiferal oxygen isotope records (Keigwin and Jones, 1995)

12.5 to 11 ka

Local centres of glaciation were active *ca*. 12.5 ka (Fig. 11. Basal ages from sites in eastern Cape Breton Island suggest an early deglaciation (Table I; Fig. 1). Truro, at the head of the Minas Basin became ice free as early as 12.0 ka (Mott and Stea, 1994), but ice must still have blocked the lower Shubenacadie River valley sometime before 11.7 ka as basal peat dates at Lantz suggest (Fig. 7). Glaciolacustrine sediments were deposited in a large glacial lake henceforth termed "Glacial Lake Shubenacadie 1" that would have encompassed much of the Shubenacadie Valley below 30 m elevation (Fig. 9) and formed sometime between 12.5 and 11.7 ka. This lake lasted for a few hundred years (varve counts at Brookfield-Fig. 7; V. K. Prest, pers comm., 1984) with a southern outlet to the Atlantic through the Lake William-Lake Charles-MicMac Lake system, present site of the Shubenacadie Canal (Fig. 9). Ice was persistent on the south side of the Minas Basin as no record of marine submergence exists there, comparable to the raised deltas of the north shore. Once the ice corridor blocking the Shubenacadie outlet was breached, flow through the outlet would have rapidly drained the lake after 11.7 ka. Although a marine connection may have been established through the Minas Basin, a marine origin for the clay deposits seems unlikely for reasons discussed earlier. After 11.7 ka, relative sea level in the Gulf of Maine had rapidly dropped below the present coastline to a lowstand of -60 m at 11 ka (Barnhardt *et al.,* 1995). The role of differential uplift in the abandonment of outlets in this region is minimal as the retreating ice front was parallel to the Shubenacadie Valley outlets (Fig. 11) and much of the isostatic recovery had already occurred (Stea *et al.,* 1994).

The Atlantic coast was deglaciated *ca.* 12 ka, with slightly earlier deglaciation in southern Nova Scotia, and in southeastern Cape Breton Island (12.3-12.5? ka). The absence of raised marine features in the region is considered to be a result of late deglaciation, rather than minimal ice (Stea *et al*., 1994).

In Cape Breton Island, Grant (1994) proposed several large ice-dammed lakes within the Margaree and River Denys valleys dammed by ice in the lowlands to the south (Fig. 11). These lakes probably formed during the initial deglaciation phase 13-12.0 ka (Fig. 11). Low shrub and herbaceous vegetation were the first to occupy the coastal landscape, usually low, wet protected areas or where soils were more suitable for plants to colonize (Mott and Stea, 1994). Trees began to migrate into the southern Nova Scotia, around 12.0 ka (Mott, 1991; Miller, 1995).

Pollen analysis from northern Nova Scotia lake and buried peat sites indicate that forest growth was accentuated after 12 ka (Jetté and Mott, 1989; Mayle and Cwynar, 1995). Spruce, sometimes preceded by poplar-aspen trees, in the form of open woodlands occupied suitable sites south of Truro and westward along the Minas Basin coast. The tree line lay somewhere in northern mainland Nova Scotia, but had not reached Cape Breton Island.

Caribou and other large land mammals migrated to the Minas Basin area by this time, through the Bay of Fundy corridor (Davis, 1991). Relative sea level (RSL) was near its lowest point. A lowstand shoreline is recorded along the Atlantic coast (*ca.* 11.6 ka) at -65 m (Stea *et al*., 1994). This shoreline can be traced around Nova Scotia to the Fundy coast where it appears to be tilted upwards to -35 m (Fader *et al.,* 1977). The first human migrations probably followed the lowstand corridors into the Bay of Fundy that had been deglaciated first. Davis (1991) argues that human occupation of the Debert site began around 11 ka, during the pre-Younger Dryas time of climatic warming, in spite of radiocarbon ages which average around 10.6 ka (MacDonald, 1968). The younger ages at the site, Davis argues, may have been contaminated by laboratory pre-treatment.

Nova Scotia was virtually ice free by 11,000 years ago. Small remnant ice masses may have persisted in the Cobequid and the Cape Breton Highlands.

11-10 ka

At 10,800 yr BP, an abrupt and dramatic climatic cooling lasting several hundred years strongly affected the landscape and its vegetation cover. Small glaciers were reactivated and produced ice dams that led to flooding of lowland areas, sites of former peat bogs. Lake and bog basins were inundated by clay and silt, emanating from the base of melting glaciers or permanent snowpacks, leaving a distinct marker horizon in lake sediment cores. In Nova Scotia, spruce trees were decimated if not completely eliminated at some sites (Mott, 1994). Shrub birch communities reverted to willow dominance, and herbaceous taxa increased or dominated at some sites.

The extent of the Younger Dryas ice masses is a matter of some controversy. King (1994) proposed a major Younger Dryas glaciation of the continental shelf, based on correlation of an unconformity in offshore glaciomarine sequences, but later recanted based on evidence presented by Stea (1995), namely the continuity of a lowstand shoreline across the proposed offshore Younger Dryas ice margin. On land the glacier margins are not well defined, with the possible exception of ice marginal? ribbed moraine in the vicinity of the stratotype locality at Collins Pond (Stea and Mott, 1989; Fig. 1). Stagnant masses of ice over the Magdalen Plateau region (Figs. 1; 11) are necessary for the formation of Younger Dryas glacial lakes in central Nova Scotia, western Cape Breton and the northern edge of the Cobequid Highlands (Fig. 11). A lake formed again in the Shubenacadie valley (Glacial Lake Shubenacadie 2) as the mouth of the Shubenacadie River was blocked by aufeis/glacier (Fig. 11). The extent of this lake is less certain than Glacial Lake Shubenacadie 1, the first phase of lake formation around 12 ka. Clay deposits at Judique discovered by Robert Turner in 1992 (Figs. 1, 7) may indicate an extensive Younger Dryas ice marginal lake system, throughout the coastal lowlands south of Port Hood, which would also include the Campbell site (Mott and Stea, 1994). Late ice over the Gulf of St. Lawrence area is inferred by a lack of pre-10 ka organic sediment in lake basins in Prince Edward Island and glaciomarine sediment in the Laurentian Channel dated at 10.6 ka (Loring and Nota, 1973).

THE YOUNGER DRYAS AS AN INCIPIENT GLACIATION

The sedimentary record outlined in this study suggests that during the Younger Dryas Chronozone, embryonic glaciers were created, then abruptly aborted with the onset of Holocene warmth. Maritime Canada, has the highest precipitation in eastern Canada, and harboured autonomous ice centres during the Quaternary (Prest and Grant, 1969; Stea *et al.*, 1992; Piper *et al.*, 1994), some of which extended to what is now the offshore bank areas (Stea *et al.,* 1996).

In Nova Scotia, there is a marked concordance between present-day areas of thick, residual snowpacks (F. Amirault, pers. comm., 1992) and the Younger Dryas ice /snow regions (Fig. 11). If net snow accumulation exceeded summer melting then ice sheets would have formed in these regions first. During most of the late-glacial period the region around Prince Edward Island was emergent (Loring and Nota, 1973). The lack of a moderating ocean may have contributed to ice buildup in the northern Nova Scotia/Prince Edward Island region. De Vernal *et al.* (1996) calculated summer temperatures of 1-5°C for the Gulf of St. Lawrence region. An increase in storminess, during the Younger Dryas, as noted in offshore records (Piper and Fehr, 1991; Gipp, 1994; Stea *et al.*, 1996) coupled with marked summer cooling in northern Nova Scotia, is a recipe for ice accretion. Glacierization in the Maritimes during the Younger Dryas proceeded simultaneously in the uplands and lowlands by concomitant upland snowfield and lowland aufeis accretion (Loewe, 1971), with small remaining outliers of Late Wisconsinan ice acting as "seeds" for incipient glaciers. The early Wisconsinan throughout Eastern North America, is characterized by thick glacial lake deposits, which preceded the advance of ice and the deposition of tills (Clark *et al.,* 1993). Denton and Hughes (1981) described mechanisms of ice buildup in maritime areas relating to relatively rapid and thick buildup of ice from impounded water in lowland regions. The Younger Dryas in Nova Scotia, can be viewed as a glaciation that was aborted in it's initial stages, and the sedimentary record in Nova Scotia as an indication for glacierization as a mechanism for the buildup of Appalachian ice sheets.

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