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HALIFAX 2005

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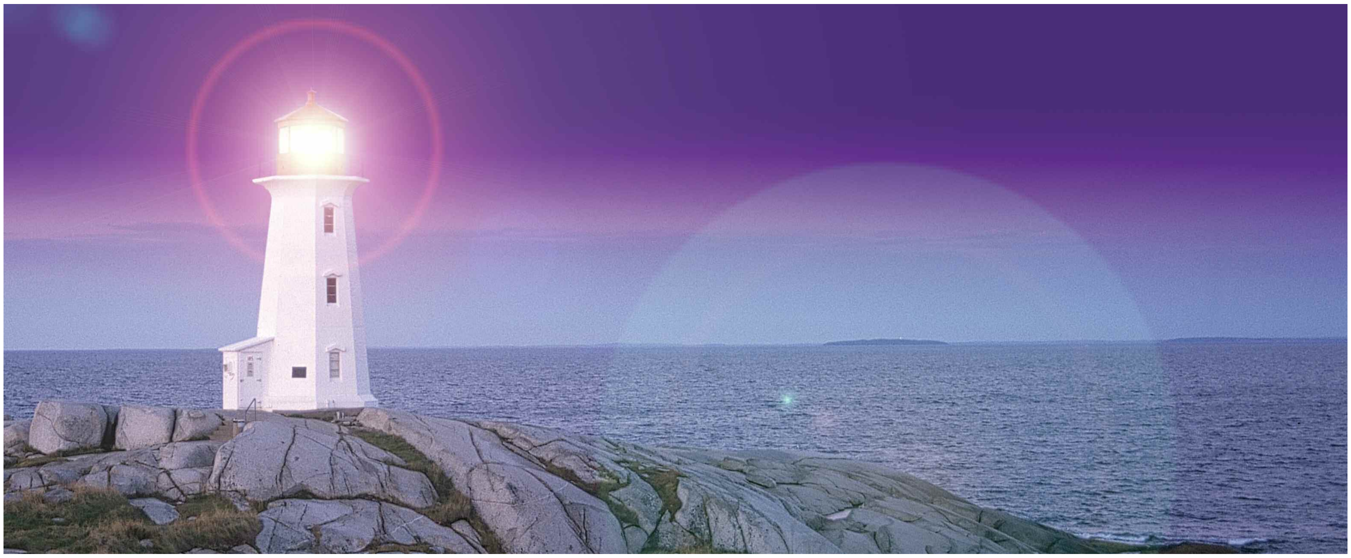

Field Trip A3

Glaciation and landscapes of the Halifax region, Nova Scotia

Ralph Stea and John Gosse

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Geological Association of Canada
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Joint Meeting - Halifax, May 2005

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Glaciation and landscapes of the Halifax region, Nova Scotia

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SAFETY PROCEDURES

Field trip participants are asked to read and heed the following safety procedures. We will attempt to make the trip as safe as possible but we need your co-operation.

1. SUITABLE CLOTHING. The weather is normally cool and wet in Nova Scotia in mid May. Participants should be prepared for heavy rain. Many of the outcrops are along exposed shorelines so your footwear should be suitable for walking on wet, slippery rocks.

2. PICKS AND HAMMERS. Please respect the outcrops and do not pick or hammer indiscriminately. Make sure no-one is standing close to you.

3. SAFETY GOGGLES. A few pairs of safety goggles will be provided. Please use them when hammering or picking - these rocks are extremely "chippy"!

4. ROAD CUTS. Several stops occur along or near busy highways. The bus and vans will pull well off the road, but in some cases, the shoulder may be quite narrow. Please use extreme caution when walking around the vans, and avoid crossing the highway.

5. QUARRY SECTIONS. We will be visiting a quarry section where hard hats and goggles are required by law. The field trip leaders will be providing these. Please do not stray from the defined field trip areas as there are many hazards, including steep sink holes and cliffs.

6. COASTAL SECTIONS. Several stops are at coastal sections. Although high tides do not normally represent a danger at these localities, in stormy weather tides may be higher than usual, and some outcrops may be inaccessible. The rocks may be very slippery near the tide line. Several steep till cliffs will be visited. Please do not dislodge boulders when others are standing below. Please follow the advice of the field trip leaders. At Peggy's Cove it is imperative to stay well away from the waters edge, following the posted signage. Many have perished when engulfed by rogue waves even on apparently calm days.

7. TICKS-MOSQUITOS-OTHER BUGS. Nova Scotia is infested with ticks and, unfortunately, late May-June is the peak of tick season. They are small brownish insects (3-7 mm) resembling beetles. These insects infest broom, sweet fern, and other low shrubs. When bushes or tall grass are disturbed, the ticks are brushed onto passers-by, and subsequently burrow their heads into the skin. It is very important to inspect your body and clothing after each day in the field, to remove any ticks - they generally take about a day to burrow in, so the sooner you find them the better. Nova Scotia had its first case of Lyme disease last year , which can be a debilitating illness. First symptoms are a red "bulls eye" rash. Mosquito, deer and black fly bites are normally just a nuisance, but some people may have various allergic reactions. For precautions from ticks, mosquitos and other biting insects wear long pants tucked into socks, and we recommend the use of a DEET repellant sprayed on outer clothing.

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INTRODUCTION

Welcome to Nova Scotia! The main themes of the field trip are Quaternary glaciations and landscape development of the Halifax region. We will take a voyage from the Mesozoic to the present and discuss the development of our hills and valleys and the formation and development of large local glaciers in the Maritime Provinces collectively known as the Appalachian Glacier Complex (Fig. 1). This field guide is in black and white, but many of the stops are featured in living colour on the virtual field trip of the landscapes of Nova Scotia (www.gov.ns.ca/natr/meb/field/start.htm) and the Dalhousie University cosmogenic dating studies site (www.dal.ca/~cnef).

Just after the glacial theory was born, a controversy emerged about the nature of glaciation in Maritime Canada which still resonates today. Was the ice local, originating in upland areas and confined to the land masses, or was the ice part of a great continental ice sheet from Québec which crossed the Bay of Fundy? The Reverend D. Honeyman, curator of the provincial museum in the late 1800s, discovered basalt boulders near Halifax that had been carried 130 km from the North Mountain, a basalt cuesta that rims the Bay of Fundy (Fig. 2). He used the observation to support the concept of a Québec-based ice movement that crossed the Bay of Fundy (Honeyman 1883). At the turn of the last century Robert Chalmers (1895) of the Geological Survey of Canada was the first to systematically map glacial deposits and landforms in Eastern Canada. He mapped glacial grooves and striae and proposed that northern Nova Scotia had been glaciated largely by local glaciers with floating ice a secondary agent in low-lying areas. Chalmers (1895, p. 95) stated this 'minimalist' position:

"The depression of the Bay of Fundy was not crossed by land ice from southern New Brunswick . Neither has Nova Scotia been glaciated by extra -peninsular ice from the north or north-east."

L.W. Bailey (1898) and W.H. Prest (1896), working in mainland Nova Scotia, observed erratics from both local and New Bunswick sources that supported both previous views. Bailey stated the compromise position (1898, p. 26):

"As in other parts of south-western Nova Scotia the facts connected with the glaciation of Digby Neck are, in the opinion of the writer, best explained upon the supposition of submergence beneath a continental glacier moving southward and bringing debris even from the other side of the Bay of Fundy, followed by a period of more local and restricted distribution, when the higher portions of the Peninsula became themselves the center of the movement, the latter now occurring in all directions."

In the early and middle part of the Twentieth Century many geologists, most notably J. W. Goldthwait (1924), abandoned the idea of local Maritime glaciers in favour of an omnibus Québec-based glaciation but then thinking came around when MacNeill & Purdy (1951) and Hickox (1962) revived the concept of local glaciers in Nova Scotia by mapping striae identifying distinctive South Mountain granite erratics that were transported northward to the North Mountain from a Nova Scotia-based ice cap (Fig. 2). Regional analyses of glacial features from air photographs by Prest & Grant (1969) led them to propose a unifying theory with several large

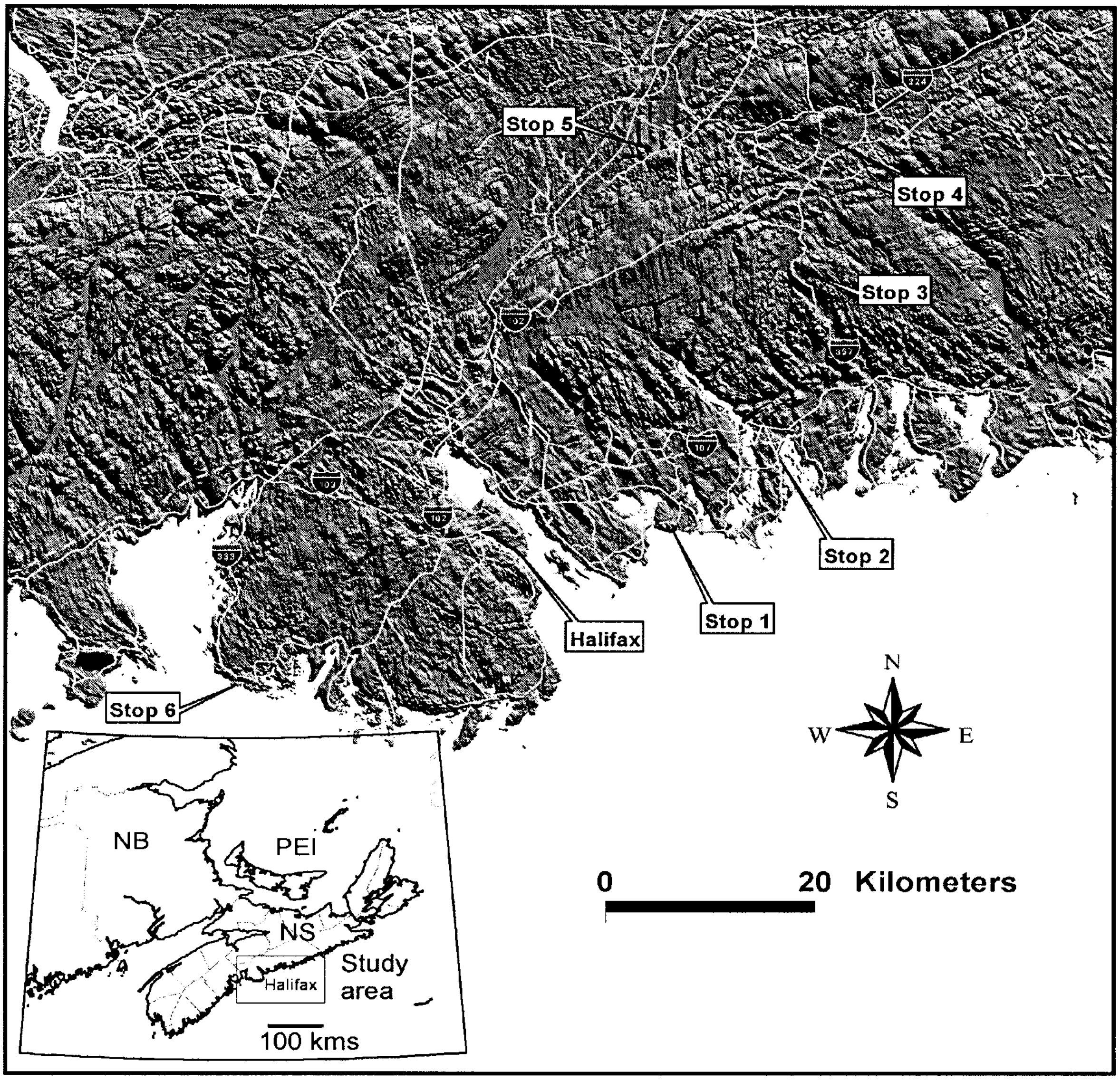


Figure 1. Route map of the glaciation and landscapes field trip.

local glaciers acting more or less synchronously, termed the Appalachian Glacier Complex. They hypothesized that the local Appalachian ice caps "held off" the Laurentide ice sheet or directed it into the Laurentian Channel.

The two opposing models of glaciation for the Maritime provinces have come to be known as the 'minimum' and 'maximum' models. The minimum model evolved from the recognition of autonomous local glaciers in Maritime Canada, into a concept of thin, Late Wisconsinan glaciers restricted to lowland terrestrial areas, and terminating just offshore of the land areas (Grant, 1977; Dyke & Prest, 1987; Grant,1989; 1994). The minimum model was evoked because of apparently unglaciated highland regions (nunataks), such as the Cape Breton Highlands, and the absence of raised shorelines on the coasts of the Gulf of St. Lawrence. The lack of raised marine features was thought to imply minimal ice loads, with the 0 'isobase' used as a proxy for the Late Wisconsinan ice margin in eastern Prince Edward Island (Grant, 1977). The isobase proxy for ice cover was based on the twin assumptions of synchroneity of marine limits and rapid glacier retreat (Wightman and Cooke 1978; Grant, 1980; Tushingham and Peltier 1991). Radiocarbon dating of glacial limits in the Bay of Fundy (Stea and Wightman 1987), the

Scotian Shelf and Gulf of St. Lawrence (McLaren1988; King and Fader 1988; Amos and Miller 1990; Mosher et al. 1989; Gipp and Piper 1989; King 1996; Josenhans and Lehman, 1999) have subsequently rendered the minimum model untenable (Fig. 2). Mapping studies in the onshore areas have documented several glacial movements on the Caledonia and Cobequid Highlands, thought to be Late Wisconsinan nunataks, and a model of larger, local ice caps has emerged (Stea 1983; Rampton et al. 1984; Pronk et al. 1989; Seaman, 1991; Stea et al. 1987; 1992a, b; Broster et al. 1997). Previous studies have also emphasized the presence of powerful ice streams in the Bay of Fundy and Laurentian Channel draining local ice divides (e.g. Mayewski et al. 1981; Belknap et al. 1989; Grant 1989; Stea et al. 1998; Shaw 2003). Whether Laurentide ice crossed the region at any time during the Wisconsinan is still an important, unresolved question (Stea et al. 1998; Dyke et al. 2002).

Stea et al. (1998) presented an empirical model of glaciation, based on a collation and correlation of glacial landforms and deposits onshore and offshore Maritime Canada, a work begun by Grant and King (1984). In Maritime Canada ice margins are offshore, and the method of glaciological investigations vary widely between the land and offshore. Stea (1995) and Stea et al. (1998) attempted to match mapped glacial flowlines from the land areas with marginal morainal deposits in the offshore, using the lithic properties of the marine and terrestrial deposits as a means of correlation. Flow lines were established in the land areas by mapping erosional and depositional glacial landforms and erratic dispersal patterns (cf. Stea and Pe-Piper, 1999). The offshore limits were dated by using 'till tongues' (cf. King and Fader 1986; King 1996), wedges of till that are rooted in submarine end moraines and interfinger with fossiliferous glaciomarine sediments. If offshore-onshore correlations are robust, the oxygen isotope chronology developed offshore can be linked to the deposits on land. Since the Stea et al. (1998) publication there have been a number of important onshore and offshore studies (e.g. Josenhans and Lehman 1999; Seaman 2000; Shaw 2003) that have greatly increased our database of the glacial geology of Atlantic Canada and must be incorporated into an empirical model. A logical next step is to test these inductive models with conceptual ice sheet models, a process now underway starting with the Atlantic Canada Ice Dynamics Workshop at Dalhousie University (ACID Group 2002). Note that the chronologies presented in this field guide are in radiocarbon years (^{14}C) with calibrated ages (CAL) designated in several summary figures.

PRE-WISCONSINAN EVENTS

The oldest Quaternary deposits in Maritime Canada are believed to be represented by the iron-cemented Bridgewater and Mabou conglomerates, assigned by various authors to the Tertiary to early Pleistocene (Prest et al. 1972; Grant 1989). A pre-Illinoian interglacial interval (200-300 ka) is indicated by amino acid racemization dates from a shell-bearing diamicton in southwestern Nova Scotia (Wehmiller et al. 1988). Lack of evidence for pre-Illinoian glaciations in Nova Scotia is puzzling, considering the preservation of unconsolidated Early Cretaceous outliers in lowland fault basins (Stea and Pullan 2001) and the offshore record of glaciations spanning 1 Ma (Piper et al. 1994). Karst topography in many lowland areas provides ample opportunity for the preservation of older glacial and non-glacial deposits, and these have been made accessible by gypsum mining and shoreline erosion.. Nonetheless, the Quaternary record appears to begin at the penultimate glaciation. A possible explanation for the lack of deposits, discounting erosion, may be the development and persistence of local, cold-based ice caps.

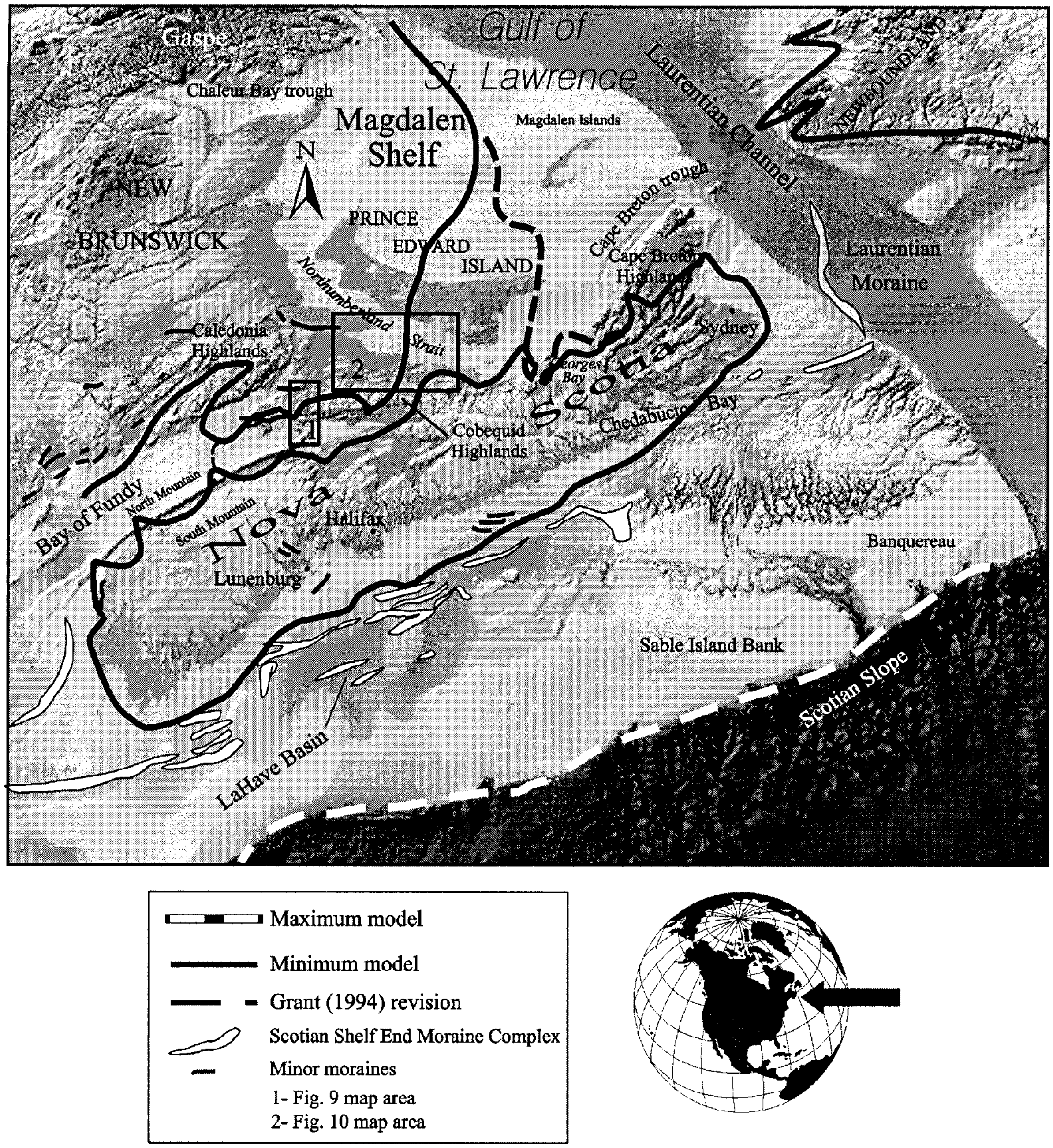


Figure 2*. Location of the area and major morainic systems on land and off-shore. Previous models of the extent and configurations of Late Wisconsinan ice masses (after Grant 1977; 1994; Mayewski et al. 1981; Dyke et al, 2002). Digital image courtesy of John Shaw and Robert Courtney (Geological Survey of Canada, Atlantic). Box 1 Fig. 9 location . Box 2 Figure 10 location.*

Organic deposits that have been assigned to the last interglacial rest on till of Illinoian (Marine Oxygen Isotope Stage (MIS) 6) age (Mott and Grant 1985; Stea et al. 1992b). Deposits assigned to the last interglacial interval are widespread throughout Nova Scotia (Fig. 3). This last interglacial interval, the Sangamonian Stage (75 -128 ka), as defined by Fulton (1984), can be correlated with MIS 5 (Grant and King 1984). During this interval, the climate fluctuated considerably. An early climatic optimum (MIS 5e) was followed by less temperate cycles (MIS 5d to 5a) that culminated in glaciation (MIS 4 to 2). Figure 3 is a summary of the stratigraphic

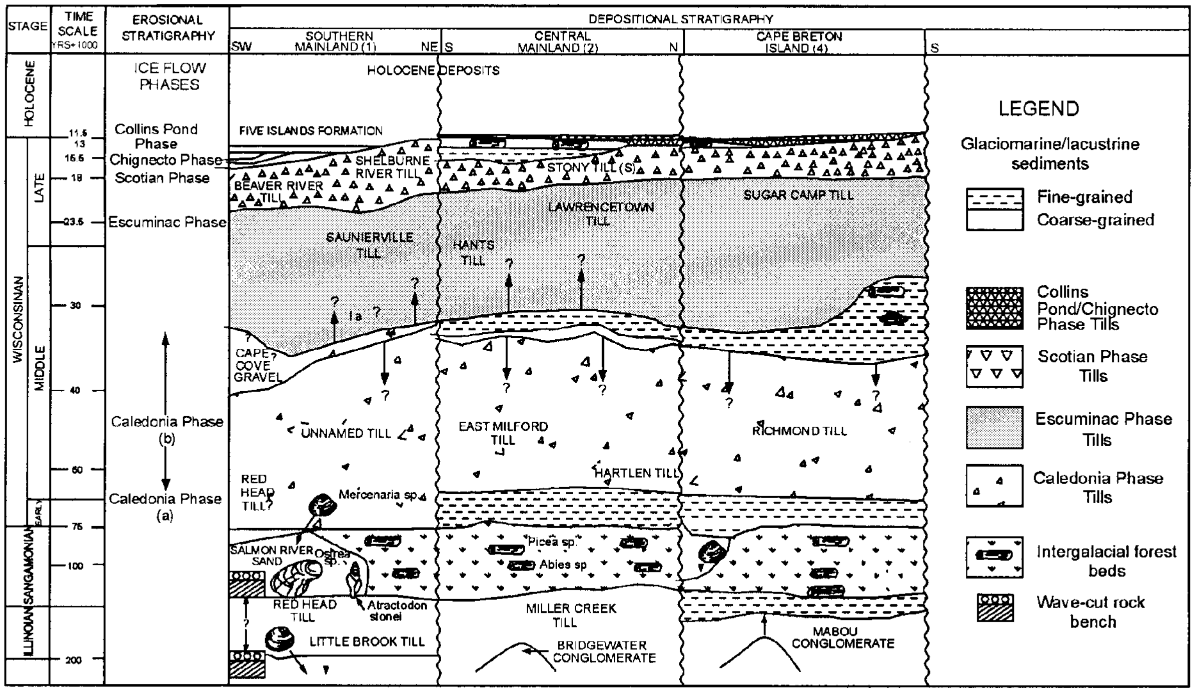


Figure 3*. Time-space diagram of terrestrial Quaternary deposits of Nova Scotia with erosional and depositional "phases" and Pleistocene Stage names (modified after Stea et al., 1992a, 1998). Time scale is in calendar years.*

and temporal relationships of Nova Scotia's Quaternary deposits. The early climatic optimum of the Sangamonian was characterized by sea level 4-6 m higher than at present (Grant 1980). This higher sea level cut a shoreline whose remnants are flat, wave-cut rock benches upon which non-glacial and glacial sediments have been deposited (Fig. 4). Numerous organic deposits, none of which completely span this lengthy non-glacial interval, have been studied by Mott and Grant (1985) and deVernal et al. (1986) who differentiate three types of pollen spectra, termed Palynostratigraphic Units 1 - 3. Unit 1 spectra are characterized by taxa indicative of climatic conditions warmer than present and forests containing abundant white pine and thermophilous hardwood taxa. The spectra of Unit 2 suggest climate similar to the present followed by cooler conditions during a stratigraphically younger interval. Spectra typical of boreal coniferous to woodland and tundra differentiate Unit 3. The possibility of a glacial event within this lengthy interval was also proposed by Grant and King (1984).

Infinite radiocarbon dates suggest a Sangamonian age for at least the temperate climate beds (Table 1). Finite dates greater than 40 ka are not reliable because the associated pollen spectra show the age to be untenable, or the material dated was suspected of contamination by embedded rootlets or some other form of "young" carbon (R.J. Mott, pers. comm. 2002).

Thorium/uranium disequilibrium age determinations on wood from several deposits (Causse and Hillaire-Marcel 1986) have shed some light on the age-dating problem. The uranium series dates confirm a Sangamonian age for most of the beds, but the youngest unit (Unit 3) produced some Middle Wisconsinan ages. DeVernal et al. (1986) proposed an extended interglacial interval with fluctuating climate and no glacial interruptions until the Middle Wisconsinan. The Middle Wisconsinan U/Th dates, however, are considered minimum ages

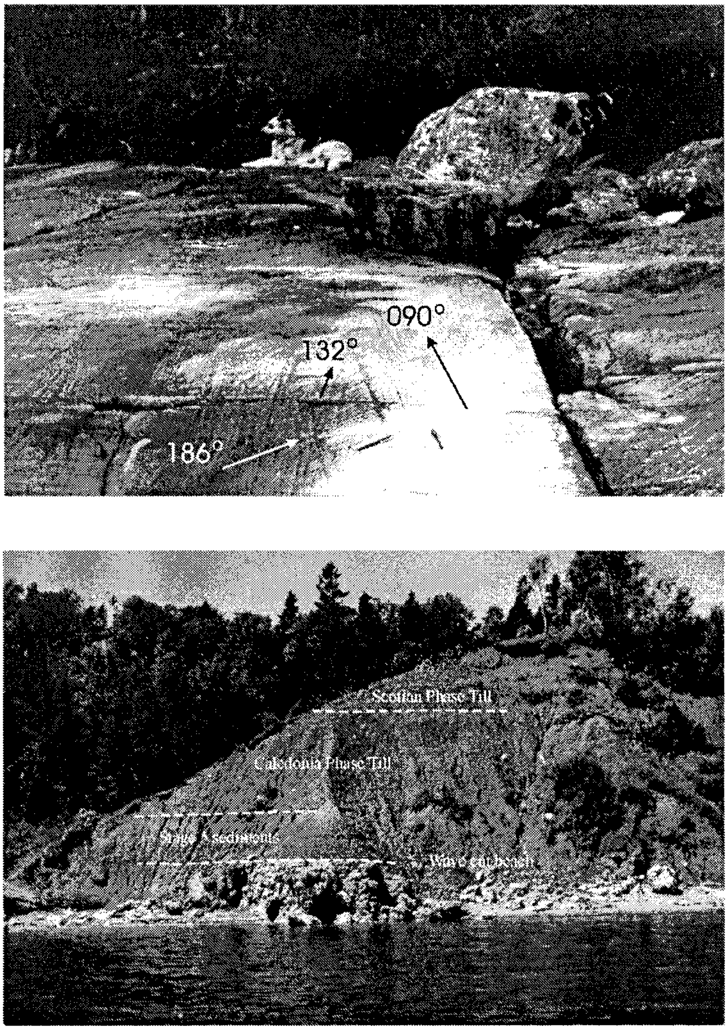


Figure 4. *(A) Photograph of a reference section for erosional stratigraphy at Charles Island (striae directions 090°; 140°, 185°; Phases la-lb, 2; (B) MIS 5e wave-cut bench (Cape George Nova Scotia overlain by sand and gravel and two till sheets, (Caledonia and Scotian Phase tills)*

because of the strong possibility of post-depositional U migration (Stea et al. 1992b). Godfrey-Smith et al. (2003) have presented optical luminescence and electron spin resonance dating of sediments and mastodon bone and dental enamel at the type section in East Milford, which have confirmed the U/Th dates from the same stratigraphic interval (Table 1: Fig. 5). They showed that a mastodon was living in a boreal/tundra forest environment at the end of the last interglacial (MIS 5a; ~75 ka; Table 1).

CALEDONIA PHASE (PHASE 1)

The timing of glacial onset in eastern Canada is uncertain (Clark et al. 1993). Grant (1994) presented cogent evidence that glaciers developed in some highland areas such as the Cape Breton Highlands before coalescing with regional ice sheets in the Early Wisconsinan. Glacial lakes, requiring local ice dams, also developed in many lowland regions after the interglacial and before the deposition of till (Grant 1994).

The oldest pervasive ice flow patterns on land in Maritime Canada are east- and southeastward (Caledonia Phase, Figs. 5 and 6). Eastward flow patterns (Phase la) are prevalent along the shores of the Gulf of St. Lawrence, and were originally assigned by Chalmers (1895) to the Appalachian-based Northumberland Glacier. This ice configuration later changed to a long-lived ice centre in New Brunswick (Gaspereau Ice Center; Rampton et al. 1984; Pronk et al. 1989) which produced southeastward striation patterns on the Caledonia Highlands in southern New Brunswick. These east and southeast flow patterns relate to an extensive ice sheet that eroded the tops of all highland areas in Maritime Canada (Grant 1977; 1989;1994) and deposited overconsolidated, matrix-rich tills (McCarron Brook Till-Hartlen Till-East Milford Till; cf. Williams et al., 1985) found throughout Nova Scotia (Fig. 3). These 'mature' tills have been linked to the eastward and southeastward flow pattern through till fabric and provenance studies, (Nielsen 1976; Stea 1984; Alcock 1984; Stea et al. 1986, 1992b; Graves and Finck 1988; McClenaghan and DiLabio 1995, 1996; Grant 1994; Stea and Pe-Piper 1999). The source of the Caledonia Phase glacier is not certain. There

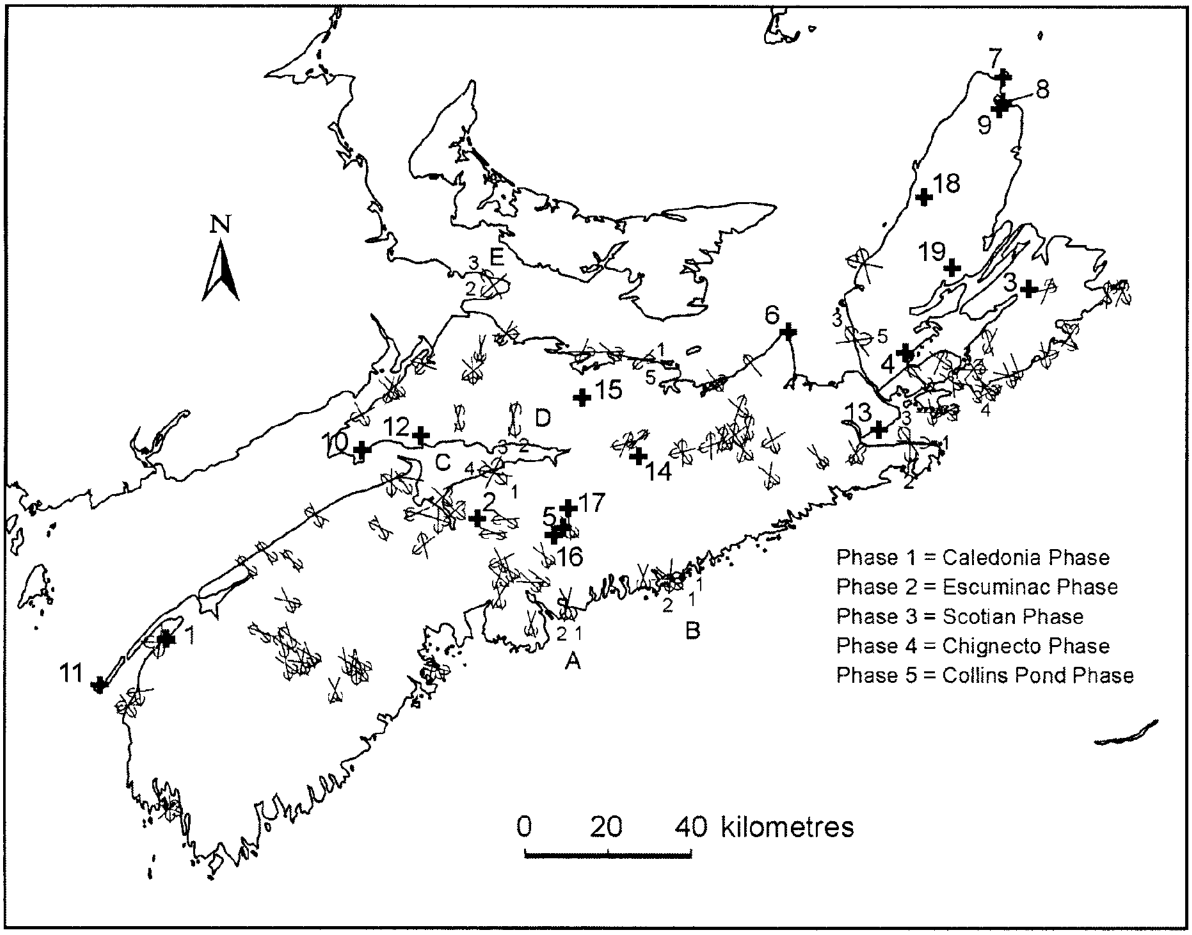


Figure 5. *Type cross-striated outcrops in Nova Scotia and relative ages on outcrop (after Stea et al. 1992a; Grant 1994). A-Charles Island, B-West Lawrencetown (Phases 1, 2), C -Tennycape Quarry.(Phases 1, 3, 4), D Cobequid Highlands (250 m) Phases 1, 2, E Cape Tormentine (Phases1, 3 Rampton & Paradis, 1981; Stea et al, 1987). Location and numbers refer to radiocarbon-dated reference sections in Table 1.*

is a lack of striae or erratic evidence indicating the passage of a southeastward-flowing Laurentide Ice Sheet from Québec across central and northeastern New Brunswick (Rampton et al. 1984; Pronk et al. 1989; Lamothe 1992), although Lamothe (1992, p. 29) cautions that Canadian Shield erratics may be indistinguishable from local high-grade metamorphic rocks. The oldest striae patterns in north-eastern New Brunswick are an eastward Appalachian flow that was directed into the Baie des Chaleurs from highlands to the west and southwest (Pronk et al. 1989; Parkhill and Doiron 2003). This contrasts with Charbonneau and David's (1994) data on erratic dispersal patterns in the Gaspé Highlands to the north (Fig. 2) suggesting a "long episode of regional and consistent south- eastward ice flow". It may be possible that that the southeastward flow on the 1000 m high Gaspè Highlands is relict from the Illinoian glaciation and similar evidence in NE New Brunswick was largely removed by Wisconsinan ice flows.

Eastward flow patterns can be found in southeastern New Brunswick (Foisy and Prichonnet, 1991), the Caledonia Highlands (Munn et al. 1996; Broster et al..1997) and southwest New Brunswick (Seaman 2000). If these early Caledonia Phase flow patterns are correlative (Fig. 6), they indicate an early Wisconsinan ice buildup in the northern Appalachians, as Chalmers (1895) first suggested.

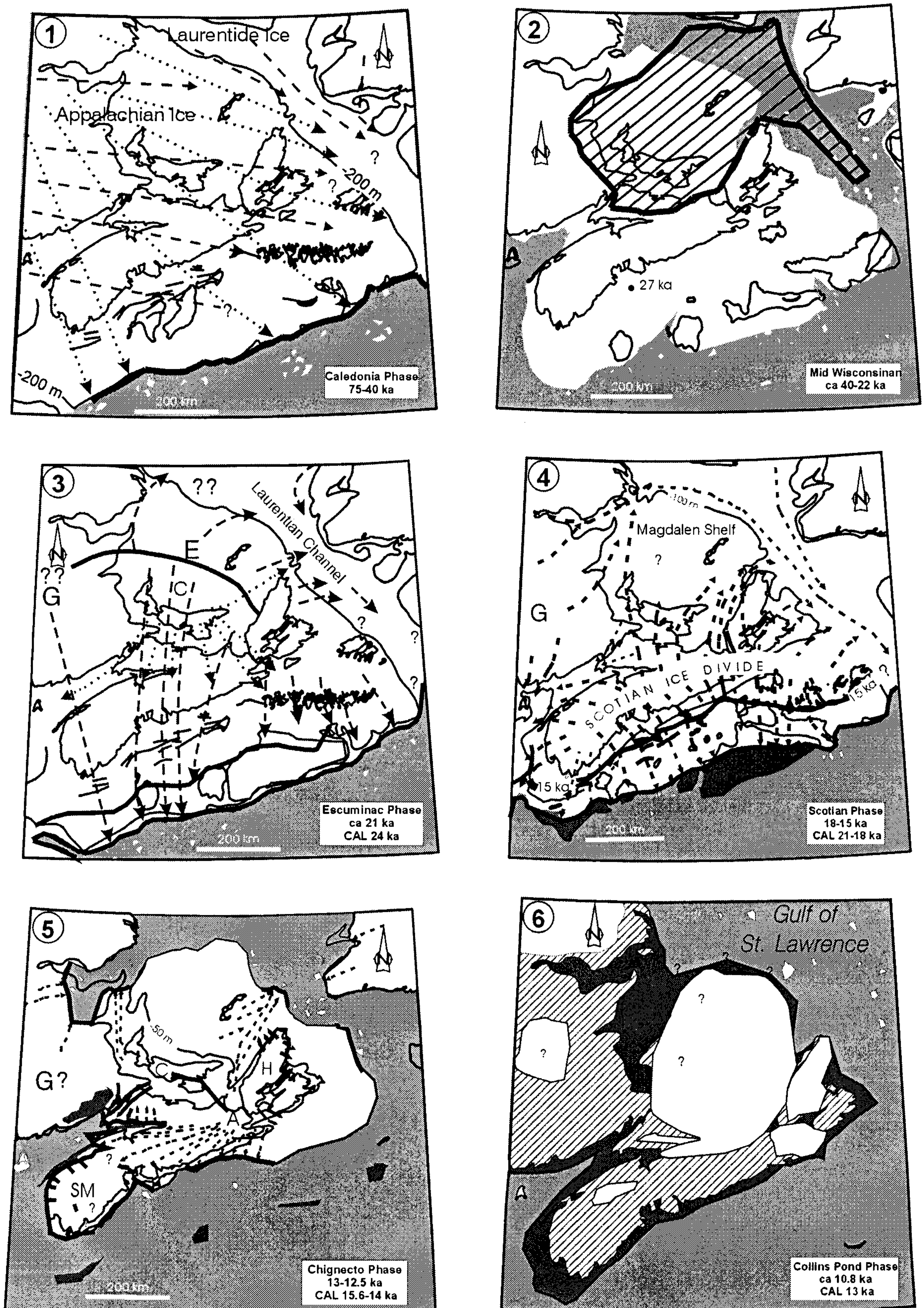

1
Laurentide Ice
Appalachian Ice
-200 m
-200 m
200 km
Caledonia Phase
75-40 ka
2
27 ka
200 km
Mid Wisconsinan
ca 40-22 ka
3
??
Laurentian Channel
E
C
G
200 km
Escuminac Phase
ca 21 ka
CAL 24 ka
4
-100 m
Magdalen Shelf
G
SCOTIAN ICE DIVIDE
15 ka
15 ka
200 km
Scotian Phase
18-15 ka
CAL 21-18 ka
5
-50 m
G?
C
H
A
SM
200 km
Chignecto Phase
13-12.5 ka
CAL 15.6-14 ka
6
Gulf of
St. Lawrence
Collins Pond Phase
ca 10.8 ka
CAL 13 ka

Figure 6 *Evolution (advance and retreat) of ice divides and domes over Maritime Canada during the Wisconsinan (75-10 ka B.P.) Note: white = ice, medium = ocean; dark = emergent land areas; dashed arrows = flowlines; dotted arrows = later flowline during same "phase" of flow. Calibrated year (CAL) conversions are shown below radiocarbon age chronology. (1) Caledonia Phase-Phase 1) in Atlantic Canada and margins (early eastward flow designated Phase la) (2) Middle Wisconsinan ice retreat and distribution of Carboniferous red beds (hatched area) of the Magdalen Shelf. (3) Escuminac Phase (2) from the Escuminac Ice Center and Divide (E) on the Magdalen Shelf and the Gaspereau ice center in New Brunswick (G). A retreat margin over Sable Island, Western, Emerald Banks (after McLaren 1988). Chignecto glacier (C) active for a short time prior to Phase 3. (4) Scotian Phase (3) (advance and retreats) (Scotian Ice Divide), Cross-hatched area = emergent marine landscapes. (5) Chignecto Phase (4) from local centers over the Antigonish Highlands and Chedabucto Bay (A); the South Mountain (SM); Cape Breton Highlands (H) and Prince Edward Island (C). (6) Younger Dryas readvance (Collins Pond Phase (5);(modified after Stea and Mott1998).*

THE LIMIT OF THE EARLY-WISCONSINAN? CALEDONIA PHASE GLACIER

Caledonia Phase tills in Nova Scotia are generally thicker than Late Wisconsinan tills and have a substantial far-travelled erratic component, prompting earlier workers to postulate that the Early Wisconsinan glaciation was the most extensive in eastern Canada (Grant 1977; Grant and King 1984). North Mountain basalt erratics found at Sable Island Bank (King 1970) indicate that a southeastward flow may have indeed extended out at least to the outer banks and probably farther (Fig. 6). The limit of relict iceberg furrowing on the outer banks occurs in water depths of 300-600m and indicates that ice extended out to the shelf/slope topographic break (Dodds and Fader 1986). Debris-flow deposits, interfingering with ice proximal glaciomarine deposits related to a series of tidewater ice margins, have been found on the Scotian Slope (Fig. 7). Wedge 2, a pre-Late Wisconsinan deposit, has been interpreted as Early Wisconsinan (Dodds and Fader 1986; Mosher et al. 1989; Table 2) based on sedimentation rate extrapolation. Scott et al.(1989) interpret a major tunnel-valley forming event at Sable Island Bank as an early Wisconsinan glaciation. Ice rafting records in the North Atlantic also indicate that glaciers existed in the Laurentian Channel during the Early-Middle Wisconsinan (Bond and Lotti 1995).

ESCUMINAC PHASE (PHASE 2)

Striated outcrops in New Brunswick, Nova Scotia and Prince Edward Island (PEI) record a major shift in ice flow from eastward and southeastward during the Caledonia Phase to southward and southwestward during the Escuminac Phase (Fig. 5). Goldthwait (1924) attributed the major southward flow over mainland Nova Scotia to a lobe of Laurentide ice from eastern Québec crossing the Laurentian Channel (Acadian Bay Lobe). The absence or lack of reports of Canadian Shield erratics over wide areas of northern Nova Scotia and Cape Breton (eg. Grant 1994), and the persistence of southward trends across northern mainland Nova Scotia (Fig. 5), imply a Maritime ice divide rather than Laurentide ice. This idea is reinforced by the strong southwest ice flow trends into the Bay of Fundy, clearly derived from an ice center on the Magdalen Shelf (Figs. 2, 5). An undated anorthosite boulder in Western Prince Edward Island associated with boulders derived from Devonian and Pre-Cambrian granites and gneisses in

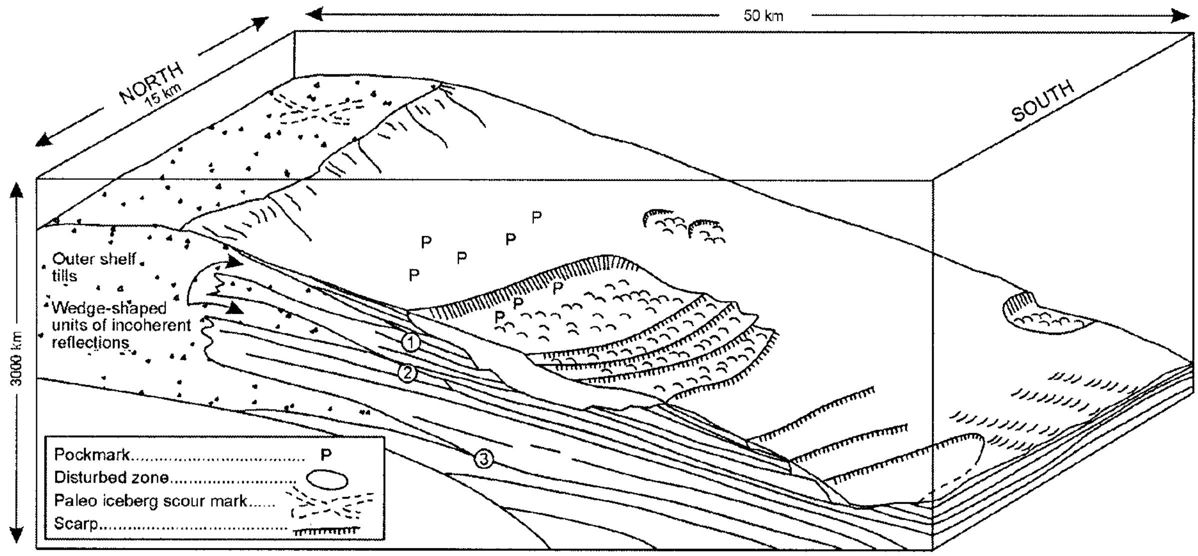


Figure 7. *Interpretative diagram of till and diamicton wedges on the outer Scotian Shelf and Slope (after Mosher et al. 1989; 1991). Interfingering of wedge-shaped, massive, incoherent seismic units (with iceberg-furrowing above 300 m) with laminated acoustic units. Three wedge units are numbered from youngest to oldest 1-3. Wedge 1 is Late Wisconsinan (MIS 2), Wedge 2 is interpreted as Early Wisconsinan (MIS 4); Wedge 3 - Illinoian? (MIS 6).*

western PEI (Prest and Nielsen 1987) may indicate an Appalachian source rather than Laurentide, as these lithologies are documented in the Caledonia Highlands of New Brunswick (Fig. 2; Barr and White, 1999; C.E. White, pers.comm., 2003) but recent erratic sampling in PEI by the authors suggests a Laurentide incursion of unknown age and extent. It is thought that the Laurentide Ice Sheet was confluent with ice from the Appalachian Glacier Complex in the Laurentian Channel (Fig. 6) based on abundant anorthosite and granulite erratics in the mid-Gulf region (Loring and Nota 1973; Stea 2001).

The south and southwestward flow trajectories were the result of an ice center situated near Prince Edward Island, which has been termed the Escuminac Ice Center (Rampton and Paradis 1981; Rampton et al. 1984; Fig. 6). It funnelled ice southwestwards into Chignecto Bay and the Bay of Fundy (Chalmers 1895; Prest 1973; Rampton et al. 1984; Foisy and Prichonnet, 1991) and radiated north into the Chaleur trough and over the Magdalen Islands, merging with the Laurentian Channel ice stream (Pronk et al. 1989; Occhietti, 1989). Parent and Dubois, (1990) and Josenhans and Lehman (1999) describe tills formed by the Escuminac Ice Center, on the Magdalen Islands and in the Laurentian Channel. During the Escuminac Phase the Lawrencetown Till was deposited over the pre-existing Caledonia Phase Hartlen Till along the Atlantic coast of Nova Scotia (Fig. 3). The red mud matrix of the Lawrencetown Till was derived in part from Carboniferous red, hematitic mudstones from the Magdalen Shelf and northern Nova Scotia and distinctive erratics from small source plutons in the 300 m high Cobequid Highlands (Fig. 2). These data verify a southward flow trajectory across Nova Scotia (Stea and Pe-Piper 1999). The Escuminac Ice Center may have been coeval with the Gaspereau dispersal center in central New Brunswick, responsible for southward flow across the Caledonia Highlands (Broster et al. 1997). The dividing line between Gaspereau- and Escuminac-derived flows in Nova Scotia is found in the Bridgewater area where red mud Lawrencetown Till drumlins with sources to the northeast, abruptly change into grey till drumlins with north-northwest sources (Grant 1963). Ice outflow from the Escuminac Ice Center during the Escuminac Phase was likely in the form of rapidly flowing ice streams or ice 'currents', as first envisioned by Grant (1963; 1976). Rapid ice stream flow may be inferred from the properties of the Lawrencetown Till drumlins which have a consistently high erratic content, as if dilution with local debris along its flow path was suppressed (Grant 1976; Finck and Stea 1995). Clark (1987) for example, suggested that rapid ice sheet flow inhibits basal mixing, and produces long dispersal trains. The Lawrencetown Till is also associated with relatively narrow drumlin fields and streamlined drift in low-lying areas of Nova Scotia, consistent with ice stream-generated landforms inferred elsewhere (Patterson 1998). Drumlin orientations and Escuminac Phase striae patterns suggest that the ice streams converged into inter-bank channels on the outer shelf, drawn by calving of the ice margin in deeper water (Piper 1991; Fig. 6).

THE MARGINS OF THE ESCUMINAC PHASE GLACIER

Escuminac Phase ice streams must have crossed the inner Scotian Shelf to the outer banks, as they appeared to have crossed the 300 m high Cobequid and Caledonia Highlands (Stea et al. 1986; Broster et al. 1997) and there are no further topographic barriers to prevent their passage. Nova Scotia-derived metamorphic erratics have been noted in glacigenic deposits sampled at the shelf/slope margin (Fig. 7; Mosher et al. 1989). The distinctive Carboniferous red mud that makes up the Lawrencetown Till along the Atlantic Coast can be traced in cores across the Scotian Shelf to the shelf edge (Cok 1970; Stanley et al. 1972; Hill 1981; Amos and Miller 1990). Based on this correlation a Late Wisconsinan age is assumed for the Escuminac Phase because it represents the last major shelf-crossing glacial event (Fig. 6). The age of the Late Wisconsinan calving ice margin at the shelf edge is between 18 and 21 ka as determined by shell dates from piston cores in marginal glaciomarine deposits, interfingering with debris flow deposits relating to these ice margins (Mosher et al. 1989; Piper 1991; Baltzer et al. 1994; Fig. 7). The maximum extent of the Escuminac Phase, shown in Figure 6, is comparable to the 'maximum model' (Fig. 2) and is consistent with ice thickness estimates on the inner Scotian Shelf (1 km; Stea 1995).

Throughout most of the Escuminac Phase and later flow phases, the Magdalen Shelf was probably an ice rise bypassed by ice streams in the Laurentian Channel, Cape Breton and Chaleur troughs (Fig. 2). Ice divides and cold-based ice located over this region may explain thin till cover over Prince Edward Island (Prest 1973) and the style of glaciotectonic deformation on the Magdalen Islands (Dredge et al. 1992; Fig. 2). Josenhans and Lehman (1999) demonstrated that the Laurentian Channel was filled with ice prior to 15 ka, thus defining the maximum extent of Late Wisconsinan ice streams within the Laurentian Channel. A critical terrestrial section at Bay St. Lawrence in northern Cape Breton (Fig. 5) can be used to defend a 'minimalist' position based on a radiocarbon date from a shell-bearing sand indicating ice retreat at 21 ka (Grant 1994). Local wood dates ranging from 32-24 ka seem to confirm this retreat event (Table 1). Diamicton above the shell-bearing unit has been interpreted as periglacial colluvium (deVernal, et al. 1986; Grant 1994) or till (Newman 1971; Stea et al. 1992b; 1998). Further examination of the sections has shown that both locally-derived till and colluvial facies post-date the shell diamicton unit (S. Occhietti, personal communication, 1998). We can infer from these data that after Mid-Wisconsinan ice retreat, the Laurentian Channel was filled with an ice stream that was fed by

Cape Breton Highland, Escuminac, Newfoundland and Laurentide ice (eg. Shaw, 2003).

SCOTIAN PHASE (PHASE 3)

An ice divide (Scotian Ice Divide) formed over Nova Scotia as ice thinned rapidly in the Late Wisconsinan, due to enhanced ice stream flow into the marine channels bordering Nova Scotia, including the Gulf of Maine-Bay of Fundy System and the Cape Breton Channel (Prest and Grant 1969; Mayewski et al.1981; Stea et al. 1998; Fig. 6). In this manner Nova Scotia ice was cut off from Escuminac and Gaspereau ice sources. Flow from the Scotian Ice Divide was northwestwards into the Bay of Fundy and southeastward over the Atlantic coast of Nova Scotia. It was also funnelled northward into Georges Bay and the Cape Breton Channel from Cape Breton and the Nova Scotia mainland (Myers and Stea 1986; Stea et al. 1992a). At an early stage of development ice streams from the Scotian Ice Divide merged with streams in New Brunswick and Prince Edward Island to flow northward through the Shediac Channel into the Laurentian Channel. This assertion is based on late northwestward-trending striae on the Northumberland Strait coast (Rampton and Paradis1981; Stea et al. 1987) and Prince Edward Island (Fig. 2; Prest 1973). This "erosional stratigraphy" can be linked with surface tills in Nova Scotia to better elucidate a sequence of events. The provenance of surface tills in sections along the Georges Bay coast of northern Nova Scotia and southern Cape Breton indicate a northeastward Scotian Phase flow (Fig. 4; Stea et al, 1989; McClenaghan and DiLabio, 1996). The distinctive, clast-dominated, locally-derived, Beaver River Till which overlies both Escuminac and Caledonia Phase tills along the Atlantic coast of Nova Scotia was deposited by southward and southeastward ice flow from the Scotian Ice Divide (Grant 1976; Grant and King 1984; Graves and Finck, 1988; Finck and Stea, 1995). The Middle Wisconsinan ice dome off Cape Breton Island, proposed by Grant (1977), was probably part of the Late Wisconsinan Scotian Ice Divide whose northwestward flow pattern can be traced across Chedabucto Bay to mainland Nova Scotia (Fig. 6).

LIMIT OF THE SCOTIAN PHASE GLACIER

The Scotian Phase glacier margin is believed to be a series of submerged moraines termed the Scotian Shelf End Moraine Complex (King 1969; King and Fader 1986; King 1996), the site of former ice retreat margins from earlier advances (Figs. 2, 6 and 8). The southern margin of the Scotian Ice Divide on the inner Scotian Shelf may have been a short-lived ice shelf, with ice rises briefly acting as centers of outflow during recession (Gipp and Piper 1989). Northward shelf-based ice flow indicators have not been documented along the Atlantic coast with the exception of a controversial study of drumlin orientations in southwest Nova Scotia (Gravenor 1974; Everett 1976).

Scotian Phase ice flow patterns, and the Scotian Shelf End Moraine Complex, were linked by moraine orientations and lithological studies of morainal deposits at sea and tills on land (Stea 1995; King 1996; Stea et al. 1998). These are extremely important correlations because they enable a flow phase or flow lines on land to be linked with a dated ice margin offshore. The

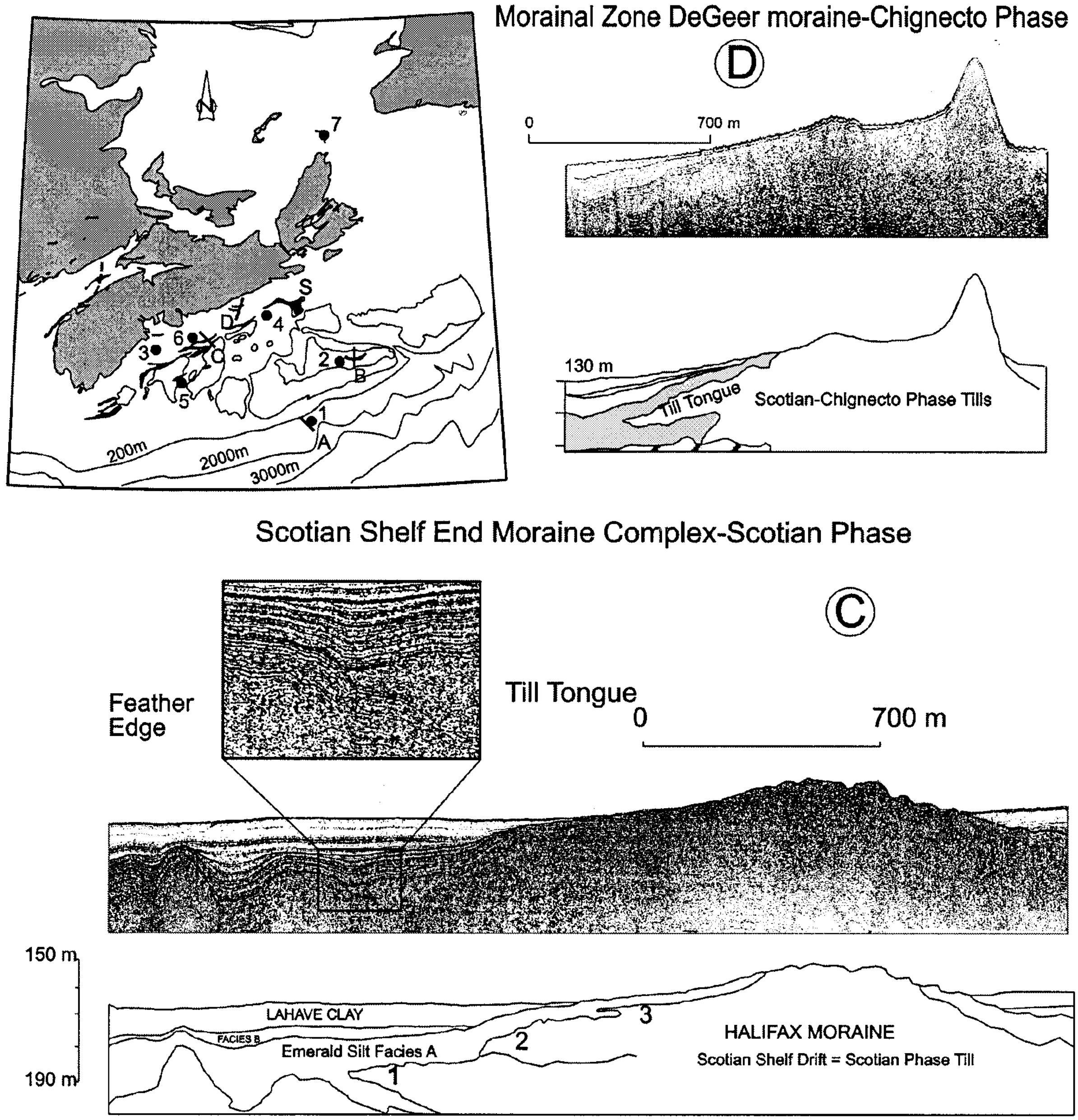


Figure 8. *Location map of four off-shore glacier margins in eastern Canada, relating to the four phases of ice flow from ice centers and divides of the Appalachian Glacier Complex. Margin A: Caledonia and Escuminac Phases outer shelf margin (cf. Figure 6.). Margin B: Sable Island Bank Escuminac (retreat phase?) and Scotian Phase? Margins Sable Island Bank (McLaren, 1988; Stea et al., 1998). Margin C: Scotian Phase, inner shelf margin. Sparker acoustic profile and interpretation of a segment of the Halifax Moraine, part of the Scotian Shelf End Moraine Complex (Fig. 1). Three till 'tongues' can be delineated in the southern or distal portion of the moraine. (4) Margin D: Chignecto Phase. Huntec (boomer) seismic profile and interpretation of a submarine moraine ridge from the inner Scotian Shelf along the eastern shore of Nova Scotia (after Stea 1995). Positions of dated piston cores in Table 2.*

Beaver River Till deposited under the Scotian Ice Divide consists almost entirely of local metamorphic rocks (Grant 1976; Finck and Stea 1995), as do the off-shore diamictons which can be differentiated lithologically from Caledonia and Escuminac Phase tills (Stea 1995; Stea et al. 1998). The Scotian Shelf End Moraine Complex formed between 15.5 ka in the northeast and 17 ka in the south-west (King and Fader 1988; Gipp and Piper 1989; Piper and Fehr 1991; King 1996). Keigwin and Jones (1995) note a peak in ice-rafted debris (IRD) production in Scotian Slope cores at around 16 ka, without a concomitant decrease in ^{18}O values, suggesting a cooling-related ice advance that may relate to the Scotian Phase.

CHIGNECTO PHASE (PHASE 4)

Cross-striated bedrock exposures along the coast of the Bay of Fundy (Fig. 5) record an intermittent shifting of ice flow from north-eastward during the Scotian Phase, to northwestward and finally west- and southwestward during the Chignecto Phase (MacNeill, in Prest et al. 1972; Stea and Finck 1984; B. McClenaghan, pers. comm.1994; Figs. 5 and 6). Rampton et al. (1984) first used the term to describe southwest-trending ice flow patterns in south-east New Brunswick. Small ice caps formed over southern Nova Scotia, (South Mountain Ice Cap-MacNeill and Purdy,1951), the Northumberland Strait area (Chignecto Glacier-Chalmers 1895; Rampton and Paradis 1981), the Antigonish Highlands (Myers and Stea 1986) and Cape Breton Highlands (Grant 1994) in northern Nova Scotia as a result of calving bay ingress into the Bay of Fundy and Cape Breton trough and drawdown or deflation of the Scotian Ice Divide (Fig. 6). Areas of northern Nova Scotia around the Gulf of St. Lawrence that featured small Chignecto Phase ice caps today experience some of the highest snowfall accumulation in the Maritimes (F. Amirault, pers. comm. 1991).

LIMITS OF THE CHIGNECTO PHASE GLACIERS

On the Atlantic coast of Nova Scotia, ice flow during the Chignecto Phase was southwestward from a center in the Antigonish Highlands (Myers and Stea 1986; Stea et al.1992a; Fig. 6). The margin of the Antigonish Highlands glacier was on the inner Scotian Shelf, represented by NW-SE oriented, submarine moraines, perpendicular to the Chignecto Phase flow pattern (Stea 1995; Stea, et al. 1998; Fig. 9). This offshore margin is estimated to be around 13 ka, based on age dates of the offshore lowstand shoreline truncating the moraines and glaciomarine sediment with the moraines (Stea et al. 1996). A glacier lobe or ice stream emanating from the Antigonish Highlands and highland ice caps in Cape Breton flowed northwards and terminated in 200 m water depth at the mouth of the Cape Breton trough in the Laurentian Channel leaving a distinctive 'till tongue' marking its readvance (Josenhans and Lehman 1999). This till-tongue margin was radiocarbon dated at around 13.2 ka (reservoir corrected; Josenhans and Lehman 1999; Table 2). The till which comprises the till tongue was correlated to the Chignecto Phase by pebble provenance (Stea 2001). The Chignecto Glacier, centred somewhere in the Prince Edward Island, Northumberland Strait area, flowed south-westward into Chignecto Bay, at approximately the same time as a residual ice cap over the Cobequid Highlands fed ice into the southern lowlands of the Minas Basin-Bay of Fundy (Fig. 6). Ice margins along the north

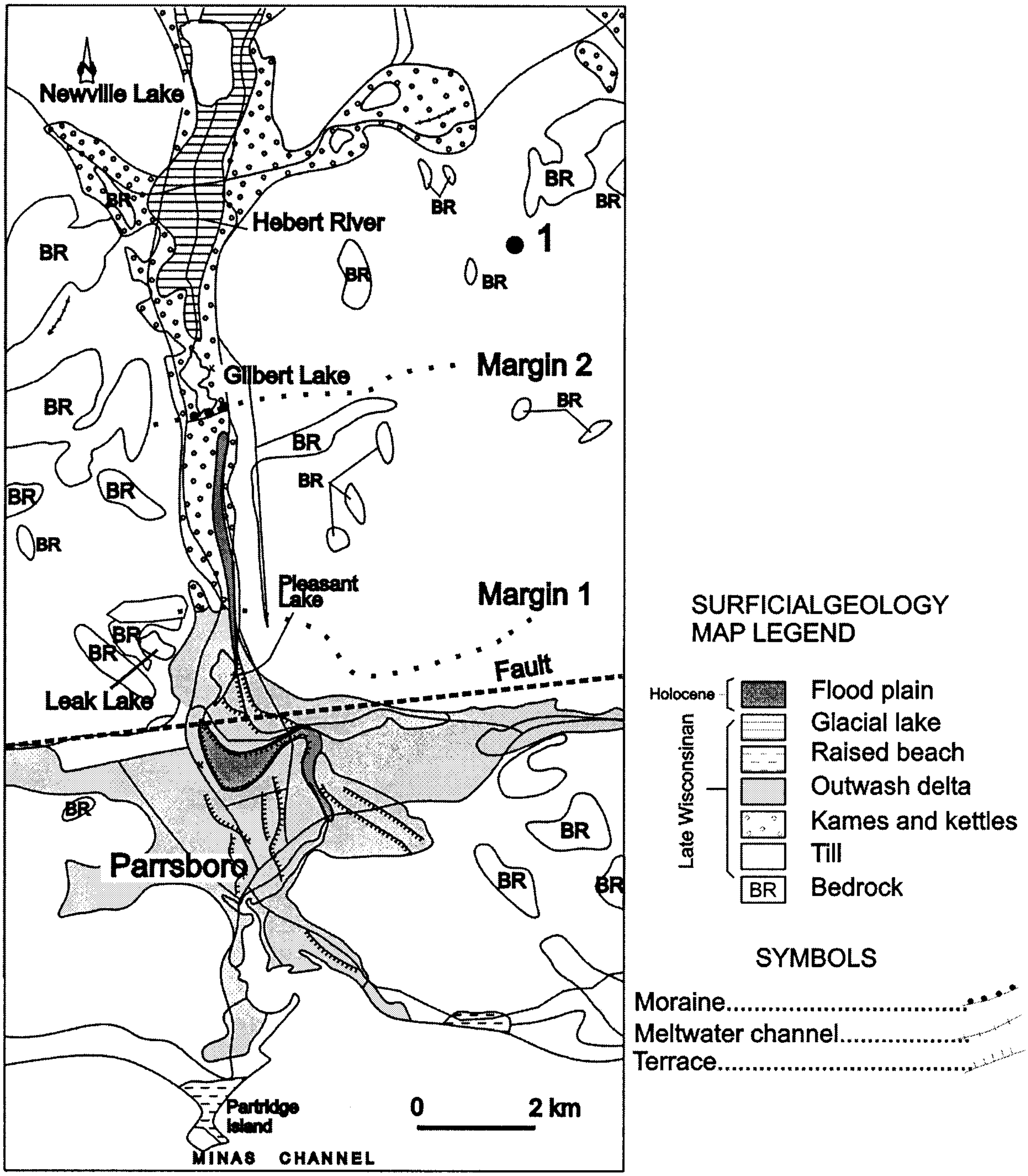


Figure 9*. Chignecto Phase (Late-glacial) ice margins in the Parrsboro Valley, Nova Scotia (location on Fig.2). Two ice margins are located north of Leak Lake (ice-contact head) and at the southern end of Gilbert Lake (moraine). (after Swift and Borns 1967; Wightman 1980; Stea et al. 1986).*

shore of the Minas Basin, are represented by outwash delta deposits termed the Five Islands Formation and Minas Terrace (Fig. 9; Swift and Borns 1967; Wightman 1980). These deposits have been dated around 13 ka by in situ *Portlandia arctica* shells in the deltaic sediments (Stea and Wightman 1987) and AMS (wood) basal lake dates from kettle lakes at the margin (Stea and

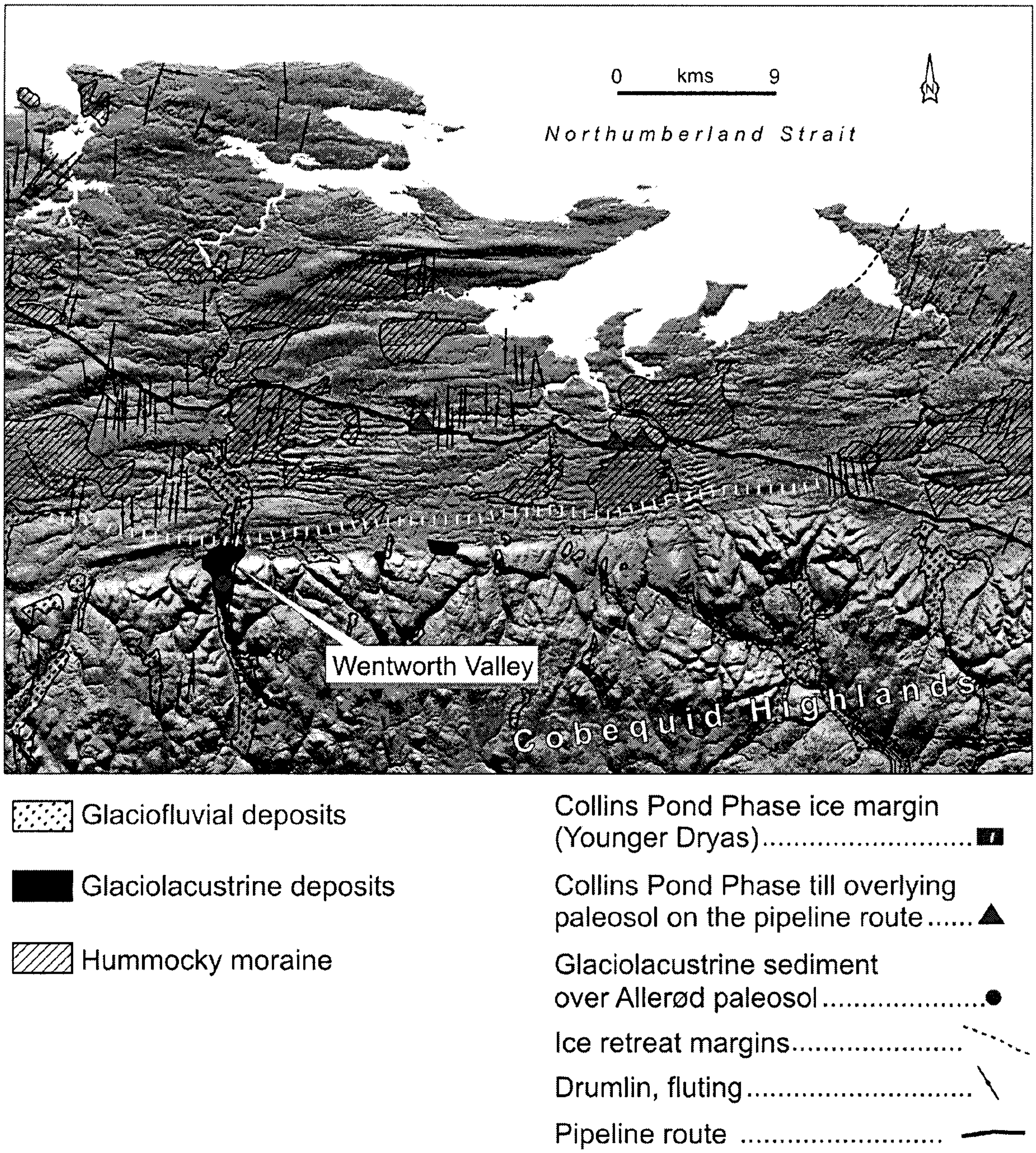


***Figure 10**. Collins Pond Phase (Younger Dryas) ice margin in northern Nova Scotia stemming from a shelf-based glacier in the Gulf of St. Lawrence .*

Mott 1998). These outwash fans were previously thought to relate to the Late Wisconsinan glacier maximum in the 'minimum model' (Grant 1977; Fig. 2). Cosmogenic dating of boulders from Peggys Cove -south of Halifax on landscapes deglaciated after the Chignecto Phase produced ages of 16.0 CAL (~13 ka radiocarbon) (MacDonald 2003)

A secondary ice readvance (Shulie Lake Phase; Stea et al. 1986) after the main

Chignecto Phase relates to an ice margin on the Chignecto Peninsula region north of the Minas Basin that is marked by the pinch-out of a distinct till and the Gilbert Lake moraine in the Parrsboro Gap dated between 13 and 11.6 ka (Fig. 9; Stea and Mott 1998).

COLLINS POND PHASE (YOUNGER DRYAS CHRONOZONE) (PHASE 5)

A prolonged period of climatic warming and ice retreat post-dated the Chignecto Phase with the dissipation of most glacier ice (Mott 1994). An abrupt and pronounced phase of climatic cooling then occurred, dated just before 11 ka, that strongly affected the terrestrial landscape, its vegetation cover and the nature of sedimentation and fauna of adjacent oceanic basins (Mott et al. 1986; Stea et al. 1996; Mott 1994). Various authors have proposed that glaciers formed and were reactivated during this time (Borns 1966; Stea and Mott 1989; Grant 1989; Lamothe 1992; Mott and Stea 1993; Grant 1994; King 1994; Stea and Mott 1998), based on ~30 sites with glaciogenic deposits overlying organic beds. Field evidence for an ice readvance during the Younger Dryas is best represented at Collins Pond, Nova Scotia (Stea and Mott 1989; Mott and Stea 1993; Stea et al. 1996) and central New Brunswick (Lamothe 1992) where glacial till overlies and deforms peat beds. Stea and Mott (1998) also describe glacial lake sediments in south-west Cape Breton that overlie an 11 ka peat bed. Offshore, Piper and Fehr (1991), King (1994), Gipp (1994) and Stea et al. (1996) described a distinctive seismic marker horizon formed during the Younger Dryas Chronozone.

LIMIT OF THE COLLINS POND PHASE GLACIER

The margins of the Younger Dryas glacier complex were not well understood, largely because of the lack of till-buried organic sites, that would allow for correlation of a till sheet with ice marginal deposits (Stea and Mott, 1998). This situation changed, however, in the summer of 1999, when the Maritimes and Northeast Pipeline Company excavated a continuous, 3 m trench across northern mainland Nova Scotia, to host the Sable Island gas pipeline (Fig. 10). A paleosol, with preserved A horizon (peat and wood), was found buried under 2- 10 m of surface till at six sites over a wide area of the pipeline route. Till fabric analyses in the surface till sheet from 3 sites, indicate a strong fabric parallel to regional late south- and southwestward glacial lineations. Radiocarbon dates on wood from the buried paleosol cluster around 10.9 ka (Table 1). The regional till sheet overlying the soil can be traced to ice-marginal deposits near the Cobequid Highlands to the south, including ice-dammed glaciolacustrine sediments overlying peat found along the coasts of northern Nova Scotia and south-west Cape Breton Island (Fig. 10). The margins of Younger Dryas glaciers have been modified from Stea and Mott (1998) to reflect the evidence of a widely distributed Younger Dryas till sheet in northern Nova Scotia. The source of this southward flowing ice must be near Prince Edward Island, the site of previous Wisconsinan ice centres of the Appalachian Glacier Complex. Grant (1994) and Stea and Mott (1998) had proposed a reactivated glacier from the Gulf of St. Lawrence region to account for the formation of ice marginal glacial lakes in the lowlands of western Cape Breton. King (1994) had also proposed a much more extensive off-shore ice advance terminating at Sable Island, but there is a lack of evidence for Younger Dryas-age glaciers on the inner Scotian Shelf (Stea et al. 1996). The formation and reactivation of both highland and lowland ice caps during the Younger Dryas in Nova Scotia may be a key to the understanding of Wisconsinan glaciation throughout the region. The concepts of Laurentide incursion in Maritime Canada, prevalent in the past, were based on a sometimes unstated assumption that the region was too far south for the development of ice sheets. The development of Younger Dryas glaciers in eastern Canada, suggests that glacierization of large regions of eastern Canada, including present-day shelf areas may have been part of the evolution of the large Appalachian ice sheets that formed over the region during the Wisconsinan Stage (eg. Stea and Pe-Piper 1999). During periods of climatic cooling, permanent snow and ice became established over the northern parts of eastern Canada. Albedo-feedback cooling and formation of blocking high pressure cells over these high latitude regions forced the jet stream and storm tracks southward over the Maritime-Appalachian region (Isarin and Renssen, 1999). Moisture-laden air masses from the Gulf of Mexico following the northeast jet stream path may have produced heavy snowfall accumulations, nourishing the incipient glaciers.

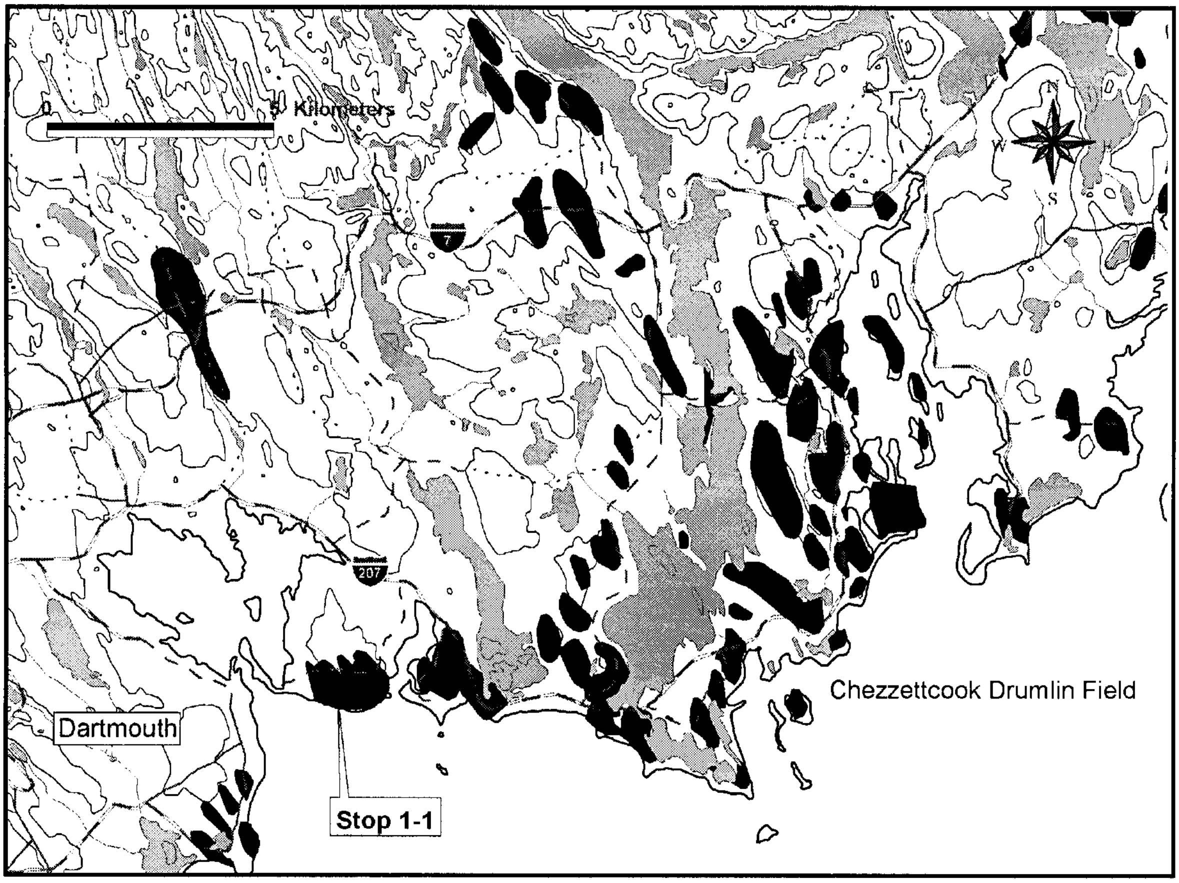


Figure 11. Location of Stop 1 and a geology map of the Chezzetcook drumlin field.

STOP 1-1: WEST LAWRENCETOWN SITE

PURPOSE: To examine a reference drumlin section with three distinct tills and a nearby striated outcrop showing evidence of the Caledonia (1) and Escuminac Phases (2) of the Appalachian Glacier Complex. Discuss till genesis, drumlin formation, glaciation models and sea-level change for the area.

ROUTE: Depart Dalhousie University, Halifax, at 8:00 AM. Cross A. M. MacKay Bridge. Cross to Dartmouth and proceed along Highway 111 through the MicMac exchange to the Portland Street exit. Proceed 12 km along Portland Street-Cole Harbour Road (Highway 207) and turn right at West Lawrencetown Road. Travel another 3 km.. Turn in the driveway and park at the end of the road. Arrive 8:45 AM. See map (Fig. 11).

EN ROUTE TO STOP

Halifax is the capital of Nova Scotia. A British garrison established the town of Halifax and fortress in 1749, with a lumber mill on the opposite side of the harbour. This British North Atlantic military stronghold countered the French fortress of Louisbourg on Cape Breton Island. The Dartmouth town site was the base for commercial whaling operations from 1785 to 1792 by the former Nantucket Whaling Company.

Although the establishment of Halifax and Dartmouth was motivated by international politics, their specific location was chosen because of geological attributes. The most distinctive feature of the Halifax-Dartmouth site is its large, deep, navigable harbour, carved by rivers, enlarged by glacial action and inundated by geologically-recent relative sea level rise. The depth and extent of the harbour lead early explorers to estimate that it could hold "1000" ships, an estimation borne out in World War II when it was used as a staging area for cross-Atlantic ship convoys. A second feature of the Halifax town site was a drumlin hill known as Citadel Hill which provided a sweeping view of the harbour and its approaches. Another important physiographic attribute of Halifax is the portage canoe route to the Bay of Fundy through a chain of lakes in Dartmouth (Lake Williams, Lake Charles, Lake Fletcher, Banook Lake) representing former spillways of large glacial lakes in central Nova Scotia. This route was used as a migration corridor for early MikMaq who fished on the Atlantic in summer and wintered in the Bay of Fundy. This natural corridor was eventually transformed into the Shubenacadie Canal System to link the Bay of Fundy with the Atlantic Ocean. The first railways in the early 1800's rendered this canal route obsolete.

From the A. M. MacKay bridge several drumlins are visible in the harbour to the south including Georges and McNabs Islands. These islands formed part of the defense system of Halifax in the early 1800s. The Naval Dockyard, where tide-gauge records have been collected since the 1890s, can be seen in distance under the other bridge. As we cross the bridge the Bedford Basin is on the left. During World War II this basin was a staging point for the convoys that crossed the North Atlantic. Just south of the bridge in the harbour narrows, a Belgian relief ship the "Imo" collided with the French munitions ship the "Mont Blanc" during World War I, creating the largest man-made explosion before the A-bomb. Sixteen hundred people died in the blast which levelled the north end of Halifax.

Bedford Basin is a glacially-moulded trough. Cores of the bottom sediments have shown that the basin was a freshwater lake before 7700 years ago. It was inundated with salt water due to rising sea levels about 5800 years B.P. (Miller et al. 1982). Bedrock exposures along the circumferential highway are striated and grooved, with several directions of ice flow indicated. South of the MicMac exchange we pass from this rough bedrock terrain onto a rolling terrain characterized by a series of southeastward-trending lakes and drumlins. (For more information see Lewis et al. 1998)

INTRODUCTION

The West Lawrencetown drumlin section was first described by Doug Grant (1963) and later by Eric Nielsen (1976). The drumlin has a v-shaped, lobate form with a main axis trending 150° and another axis trending 185° (Fig. 12). Between the two lobes is a circular depression, partially eroded by the sea, with a base resting on a lower till unit. Two exposures separated by the inter-lobe depression reveal two compositionally-distinct till units, the Hartlen and

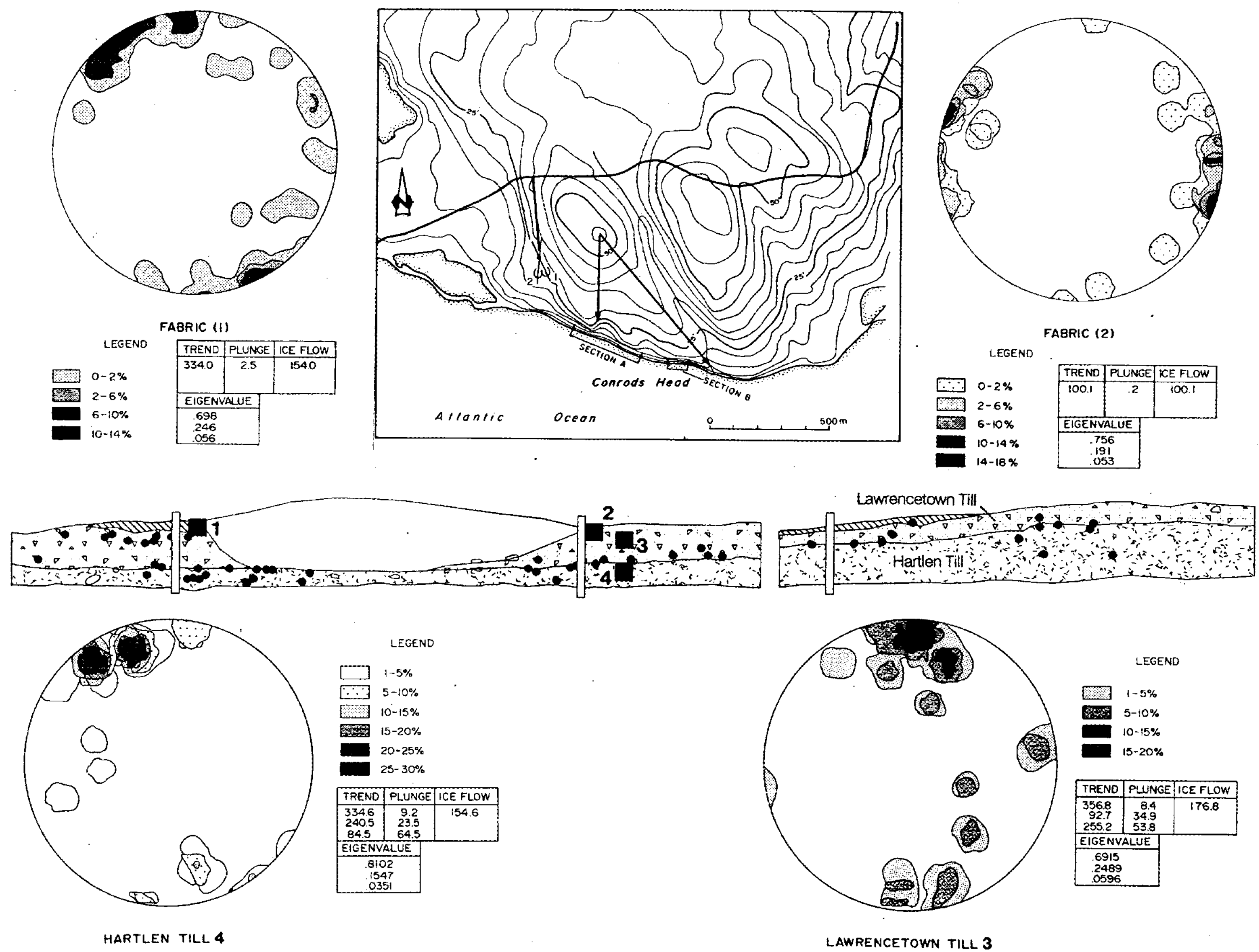


***Figure 12**. Detailed map of the West Lawrencetown sections showing striae locations, lithological sampling sections and till fabric localities. Cross-hatched area on top of the Lawrencetown till is the local, sandy facies.*

Figure 13. *Closeup of striated outcrop adjacent to the palimpsest West Lawrencetown drumlin, showing two sets of striae, 185° (younger) and 155°.*

Lawrencetown tills. Bedrock outcrop on the western flank of the drumlin complex reveals wide, parallel grooves 10-40 cm in width, and striae that trend 155° cut by another distinct set of slightly divergent, finer striae trending 180°-190°. Notice that the fine striae skip across down-glacier facing sides of grooves formed during the earlier movement, and are best developed on the up-glacier facing part of the groove (Fig. 13). The relative ages are determined not by cross-cutting relationships, which can be ambiguous, but by the upper surface development of the finer southward-trending striae.

DESCRIPTION

The lower Hartlen Till (Fig. 12) is grey, matrix-supported and highly compacted, with a strong southeastward till fabric parallel to the oldest striation set on the adjacent outcrop. Pebble counts (>1000) in the Hartlen Till average 59% metagreywacke (derived from of the underlying Cambro- Ordovician Meguma Group), 23% whitish, megacrystic Meguma Zone granitic pebbles, and 17% foreign (non-Meguma Zone) pebbles. Foreign lithologies include North Mountain basalt, red and grey siliclastic rocks, pink alkali-feldspar granitoids, diorite and felsic and mafic volcanic rocks. A distinctly reddish, muddy, matrix-supported till (Lawrencetown Till) overlies

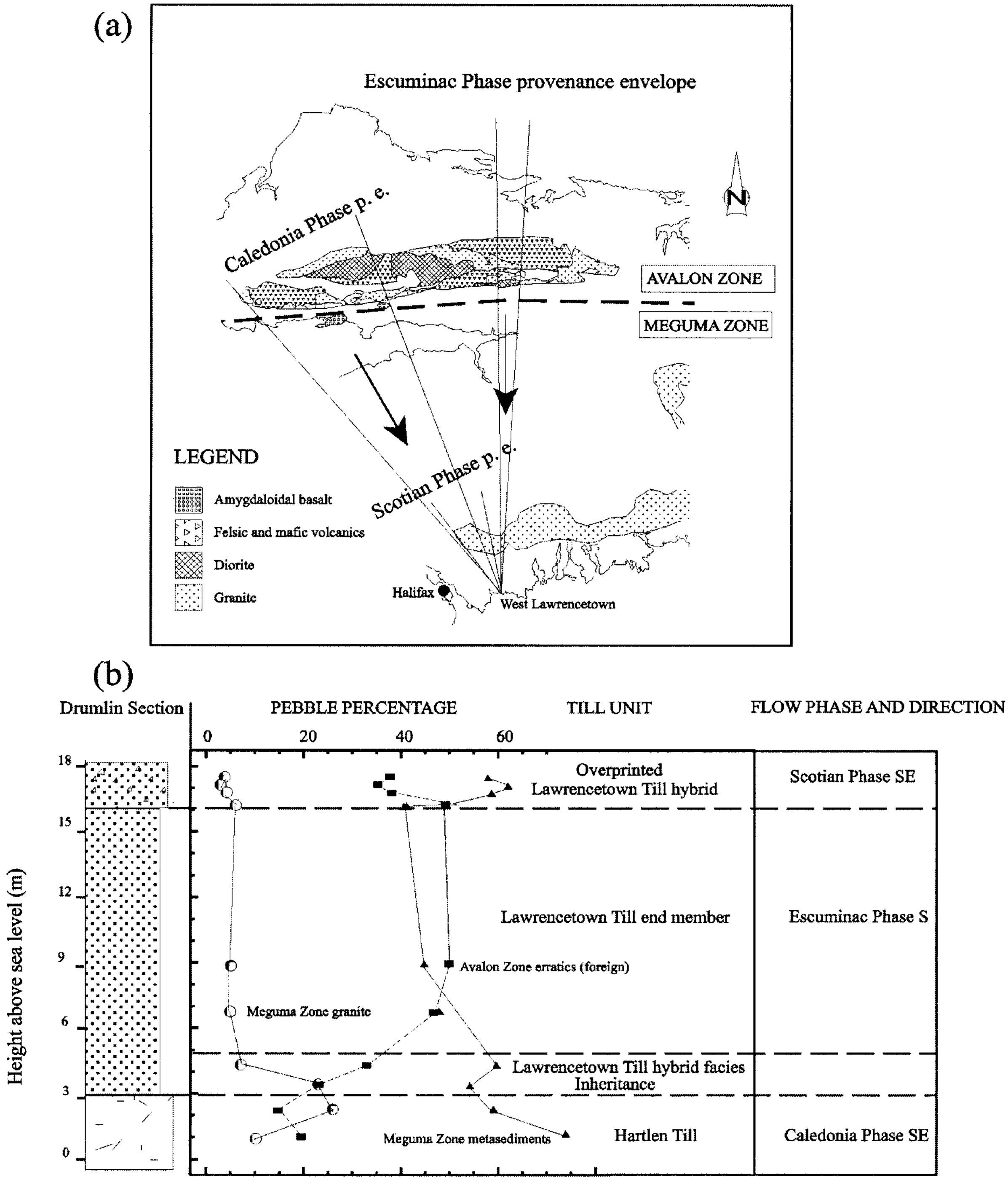


Figure 14*. Map of bedrock source areas or "provenance envelopes" (cf. Stea and Pe-Piper 1999) for the till units in the drumlin section at West Lawrencetown (location-Figs. 2 and 6). These source area envelopes were established using the overlapping ranges of source areas for erratics identified using petrology and whole-rock geochemistry. (b). Three till units are exposed at the section. The diagram shows vertical variations in pebble lithologies from the source areas above. The uppermost Lawrencetown "hybrid" Till formed by overprinting of the Lawrencetown Till during the Scotian Phase.*

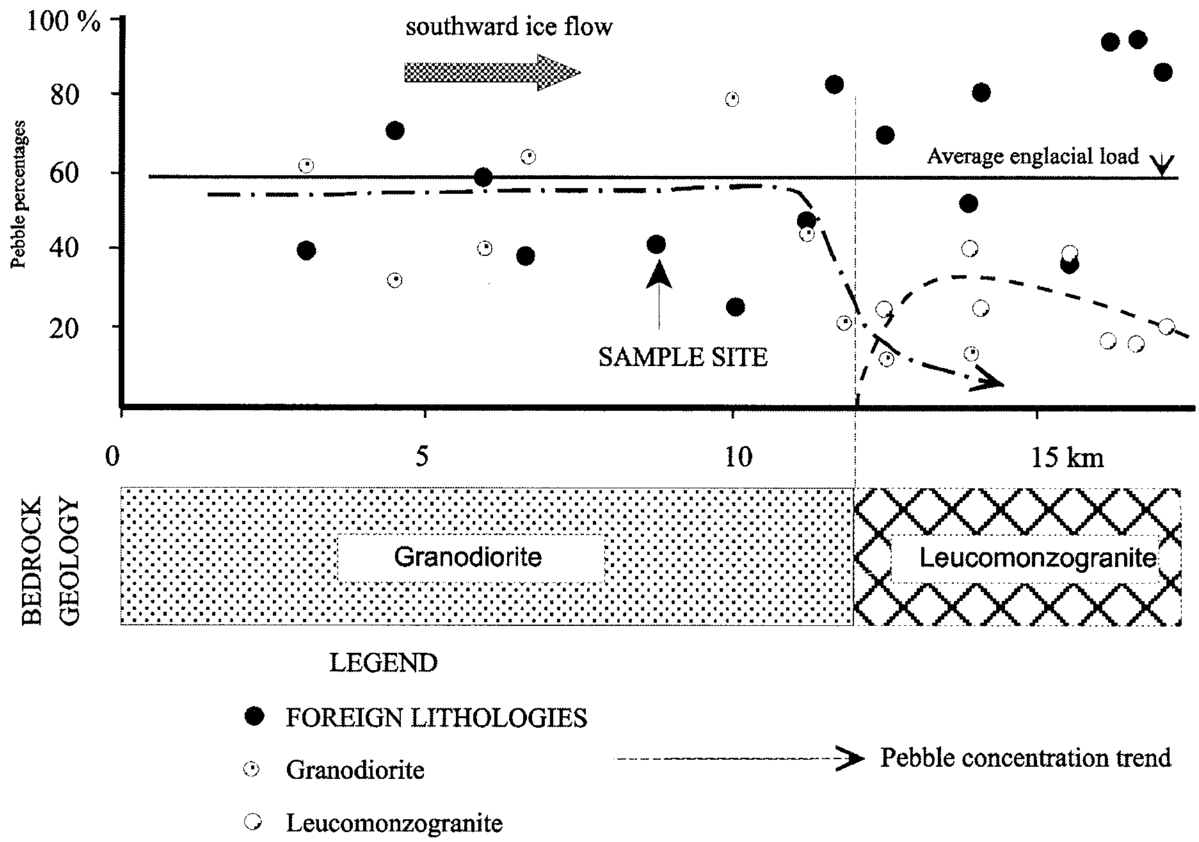


Figure 15*. Dispersal (decay and uptake) of local granite facies in the pebble fraction (>2.5 cm) of the Lawrencetown Till over the South Mountain Batholith and the persistence of the englacial load. (after Stea and Finck 2001)*

the Hartlen Till with a knife-sharp contact. The Lawrencetown Till can be divided into an upper, sandy facies and a lower, clay rich facies with sand inclusions. The lower facies exhibits a southward-trending till fabric, parallel to the south lobe of the drumlin landform and the youngest striae set on the adjacent bedrock outcrop (Fig. 12). Bullet-boulder long axes, and striae on the upper surface of these boulders, reflect the imposition of a southward ice flow. The upper facies of the Lawrencetown Till has a till fabric that trends southeastward, subparallel to the Hartlen Till, and it is enriched in local Meguma Zone pebbles. This till fabric was probably reset during the penultimate SE ice flow from a divide over Nova Scotia which also overprinted the Lawrencetown till with a flux of local bedrock (Scotian Phase-Stea et al.1998). Pebble counts (>1000) average 49% metagreywacke, 46% erratics, and 3% Meguma Zone granite, but local metawacke clasts increase to 70% in the upper hybrid facies. Foreign pebbles in the Lawrencetown Till include, in order of abundance, salmon-pink, hornblende bearing granite, granodiorite, diorite, and mafic and felsic volcanics..

SOURCE OF DRUMLIN ERRATICS AND ICE STREAM TILLS

Laurentide erratics like those in the eastern Grenville Province (high-grade metamorphic rocks and plutonic rocks-anorthosite and charnockite) have not been identified at this section or anywhere else in Nova Scotia. A Canadian Shield source and by implication a Laurentide-based ice flow is unlikely. All erratics at this section have been demonstrated to have Maritime Appalachian bedrock sources . North Mountain basalt erratics are found exclusively in the Hartlen Till and suggest a southeastward flow from known outcrops along the Bay of Fundy. The Lawrencetown Till contains distinctive pink granite erratics with sodic amphibole that are derived from the Hart Lake-Byers Lake pluton near Debert River in the eastern Cobequid Highlands, and porphyritic rhyolite that is probably derived from the adjacent rocks to the northeast (Fig 14; Stea and Pe-Piper 1999). Granite with sodic amphibole is restricted to an area of a few square kilometres around Debert River, and all the other erratic pebbles from the section have a source outcrop within a few kilometres of this amphibole-bearing granite. This narrow range of source lithologies defined as a "provenance envelope" clearly delineates a southward ice flow (Stea and Finck 2001). The drumlin "core" till (Hartlen Till) is derived from a flow across the Bay of Fundy termed the Caledonia Phase (Rampton et al. 1984; Stea et al.1998). This till is interpreted as a lodgement facies (Nielsen 1976; Podolak and Shilts1977), with boulder pavements and sand inclusions near the contact with the Lawrencetown Till, suggesting a thin englacial-meltout till facies near the top of the till unit. The drumlin "mantle" till (Lawrencetown Till) at both drumlin sections records a distinct southward shift in ice flow direction to an ice divide over the Magdalen Shelf, underlain by Carboniferous red-beds (Escuminac Ice Divide). The Lawrencetown Till has an unusually high and consistent erratic (non-Meguma Zone) content (>50%), with only minor dilution by local rocks along flow-lines (Fig. 15; Stea and Finck 2001). This is a strong indication of fast-moving ice streams, under extending flow, characterized by the lack of mixing between the englacial and basal glacier zones (Clark 1987). In fact, Grant (1963) first hypothesized that drumlins in Nova Scotia were formed as a result of warm-based, rapid ice stream flow. The surrounding regions are characterized by locally derived, stony tills, presumably bypassed by the fast moving ice. Nielsen (1976) interpreted the Lawrencetown Till as a basal melt-out till based on the abundance of sand inclusions and long-distance transport.

GROWTH OF SHELF-BASED ICE DOMES

Arguments for rapid accretion of shelf-based ice domes were first summarized in Denton and Hughes (1981). The provenance data compiled in this paper support the hypothesis of Late Wisconsinan glacier growth on the Magdalen Shelf. The formation of calving bays in the deep marine channels that bisected Maritime Canada during the Mid-Wisconsinan may have isolated an ice mass centred over the Magdalen Shelf plateau. This remnant ice mass, plugging drainage divides in the surrounding uplands, rapidly accreted during the Late Wisconsinan aided by diversion of the jet stream and increased snowfall around the larger Laurentide Ice Sheet. Ice streams draining the Escuminac Ice Divide were funnelled into channels between morainal banks on the outer Scotian Shelf and calved into the open ocean at the edge of the continental shelf (Mosher et al. 1989).

DRUMLINS AND CATASTROPHIC FLOODS

Shaw and Sharpe (1987) have proposed that catastrophic subglacial meltwater outbursts

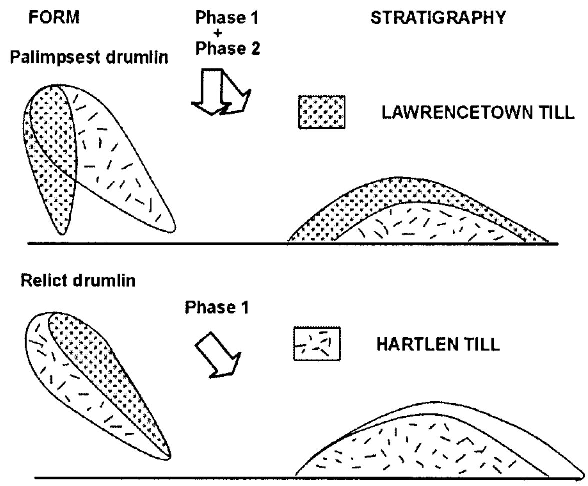


***Figure 16**. The sequential evolution of drumlin form and stratigraphy over several glaciations (modified after Stea and Brown 1989; Stea, 1994).*

form drumlins, both by erosion and deposition. This intriguing theory can be tested with data from Nova Scotia. Drumlins described in this study, and elsewhere in Nova Scotia, are largely made up of till and are surrounded by ice-scoured bedrock, inscribed with two or more sets of crossing striae. Could multiple striation sets form after a Late-glacial subglacial outburst that presumably eroded the till-dominated drumlins?. Drumlin trends are strongly correlative with striae trends in Nova Scotia suggesting a common origin (Stea and Brown 1989). Many drumlins were formed by an ice cap centred over Nova Scotia. This residual ice mass would be an unlikely candidate for the huge meltwater reservoirs envisioned in a regional subglacial flood. The lack of widespread deposits of gravel and sand and absence of unequivocal meltwater erosional forms, such as gravel-filled potholes, all mitigate against a catastrophic outburst. Time-independent process models have dominated much of the previous thinking on drumlins and other glacial landforms. The complex geomorphic effects produced by the multiple glaciations of the Laurentide and Scandinavian ice sheets have been documented by Veillette et.al. (1999) and Kleman (1994). The West Lawrencetown drumlin is considered to be a palimpsest landform (Fig. 16), that has evolved in form over two or more glacier flow phases.

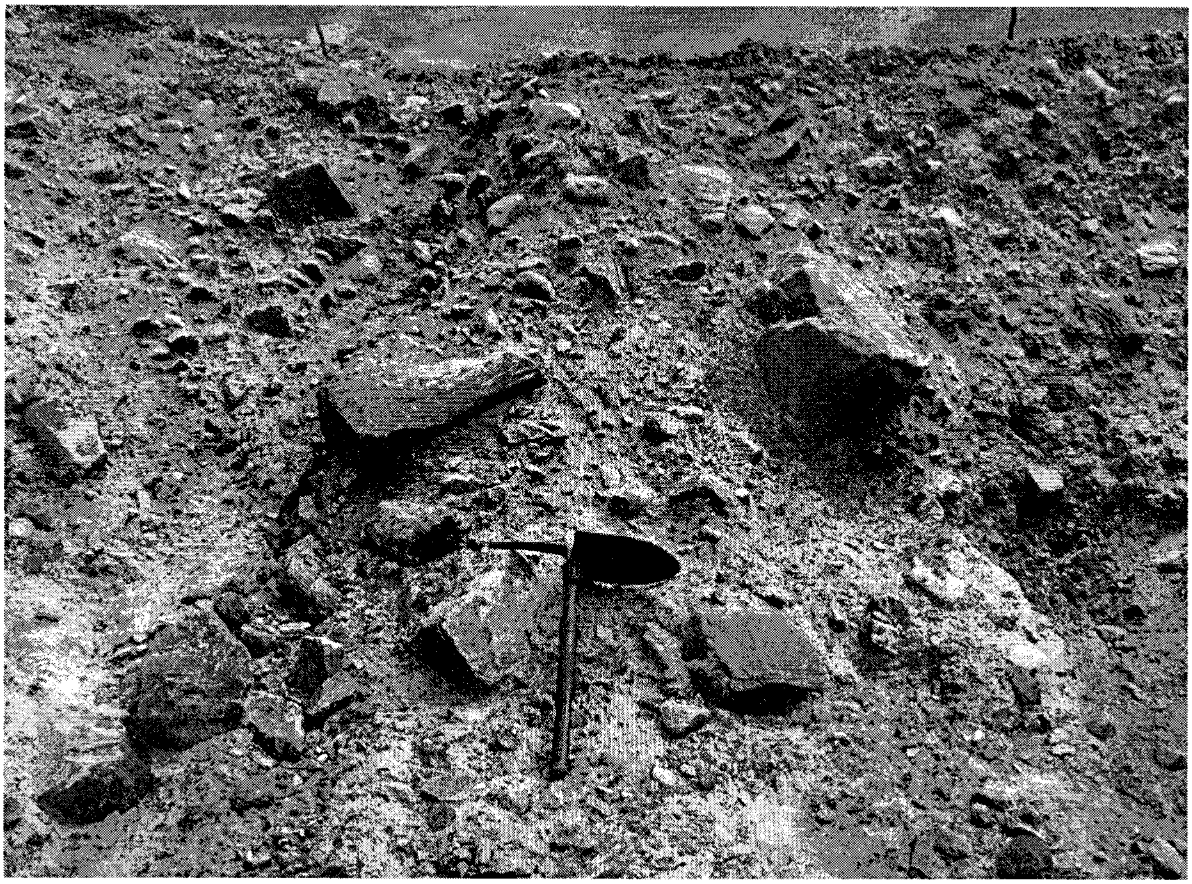


***Figure 17**. Typical example of stony Beaver River Till near Sheet Harbour, Nova Scotia.*

STOP 1-2 LOCAL STONY TILL

PURPOSE: To examine an outcrop of stony till, younger than the drumlin tills formed by the imposition of an ice divide over mainland Nova Scotia during deglaciation.

ROUTE: Leave West Lawrencetown at 10:00 am. Follow road back to trunk highway 207. Proceed eastward along 207 for ~15 kms to West Chezzetcook. Take shore road to the right and proceed for ~ 1 km. Exposure is in a road cut on a logging road. See map (Fig. 1).

INTRODUCTION

A distinctive, sandy, stony till with a high percentage of angular-subangular cobbles and boulders is found in the terrain around drumlins in the Chezzetcook drumlin field (Figs. 11 and 17). It covers some of the drumlins, implying that the deposit is younger than the drumlin tills. Apart from drumlin fields, much of the area underlain by Paleozoic bedrock on the Nova Scotia Atlantic coast is characterized by irregular, hummocky topography or ribbed moraine with large quartzite or granite boulders (from <1 m up to 20 m in diameter) strewn on the surface (Stony Till Plain; Stea et al., 1992a). The till that makes up the Stony Till Plain is called the Beaver River Till (Grant 1980; Finck and Stea 1995). This till facies was interpreted as an ablation or melt-out till

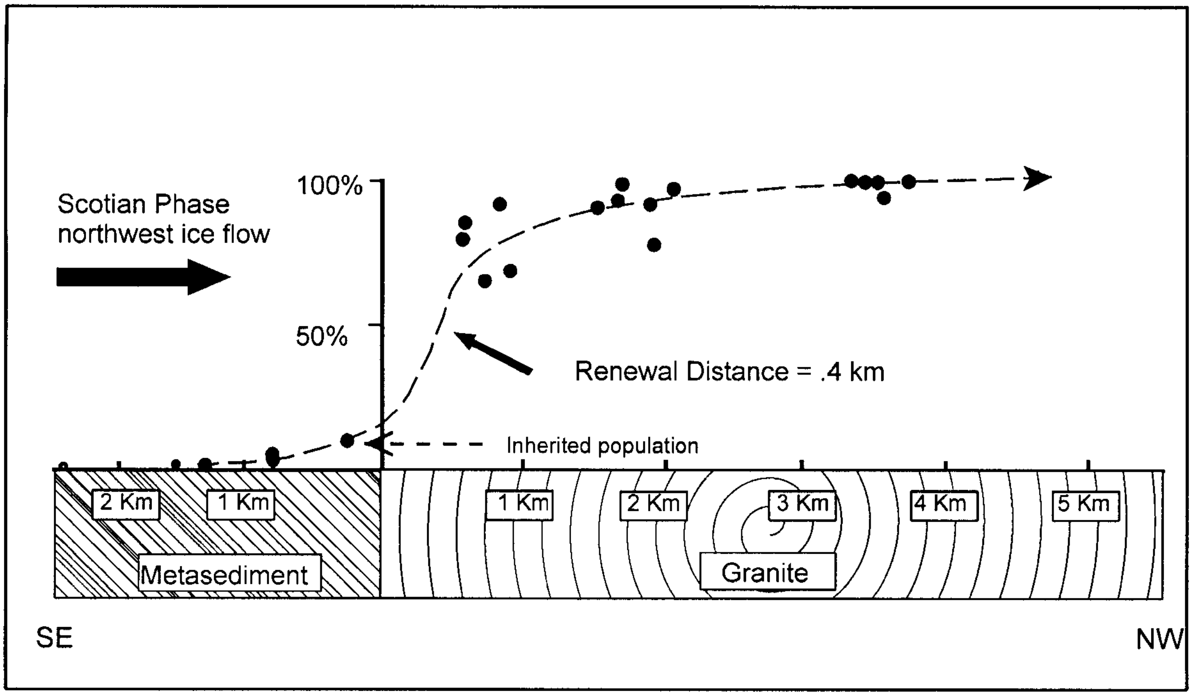


***Figure 18**. Dispersal (decay and uptake) of local granite facies in the pebble fraction (>2.5 cm) of the Beaver River Till across a granite/metasedimentary rock contact (after Stea and Finck 2001).*

based on it's looseness and coarse texture (Nielsen, 1976).

Beaver River Till thicknesses vary widely, generally increasing in lower elevations from 1-5 m in upland areas (>200 m elevation) to greater than 7 m in lower elevations. Excavations of a pipeline corridor across Nova Scotia in the Beaver River Till revealed increasing clast abrasion (clasts with striae and edge rounding) in a down-ice direction towards the coast. The pebble lithology of an "end member" Beaver River Till closely reflects underlying bedrock geology (Fig. 18). End member Beaver River Tills are dominated by a single local bedrock-derived, clast lithology, generally> 90%. Deviations from this idealized end member clast lithology reflect reworking from underlying till ("hybrid" tills) or the presence of up-ice bedrock units. The Beaver River Till can be used as an accurate tool to map bedrock geology (MacDonald and Horne 1987) and also has potential as a placer deposit in areas near gold districts.

The renewal distance, as defined by Peltoniemi (1985), is the distance down-ice over which the proportion of a new rock type increases in the till from 0% to 50%. Renewal distances calculated for the Beaver River Till range from 0.2 to 0. 8 km (Fig. 18). These extremely low renewal distances (e.g. Scandinavian Tills average 5-20 km; Peltoniemi, 1985) are indicative of low-velocity basal ice conditions in proximity to an ice divide (Stea et al. 1989). The angular nature, lack of striae on the clasts, and the predominance of clast modes, suggests that quarrying and fracturing are dominant mechanisms in the formation of the Beaver River Till rather than abrasion (Iverson 1995). Low sliding velocities are implied by the negligible dispersal, reinforcing the idea that these tills are deposited near or under the Scotian Ice Divide. These can be considered "immature" tills (cf. Dreimanis and Vagners 1971) implying that they were formed in a relatively short period of time, with little chance to develop a matrix component through comminution and abrasion. These stony "local" tills are common near former ice centres in central New Brunswick (Broster et al. 1997) and throughout the shield areas of Canada (Dredge 1983). Stea et al. (1989) used the term "ice divide till" for these till facies.

STOP 1-3 GIBRALTAR SPILLWAY AND ATLANTIC UPLANDS PENEPLANE

PURPOSE: To have lunch and take a birds eye view of the Gibraltar Spillway and the Atlantic Uplands Peneplane surface.

ROUTE: Proceed northward along side road to rejoin highway 207. Take exchange 21 and drive eastward along highway 107 to the village of Musquodoboit Harbour. Turn right onto highway 7 and drive about 6 km to the junction of highway 357. Proceed northward along 337 ~20 km to the park entrance driveway. See map (Fig. 1).

INTRODUCTION

Gibraltar Rock is a granite peak that commands a view of the mighty Musquodoboit River. After winding its way westward through the lowlands to the north, the Musquodoboit River takes a sharp southward turn and cuts through a gorge cut into solid granite. The highlands to the south are the Atlantic Uplands, an exhumed upland surface underlain by Cambro-Ordovician slate and metawacke, and Devono-Carboniferous granite. This ancient peneplane slopes from 300 m elevation in the vicinity of Gibraltar Rock to sea-level along the Atlantic coast. During this stop we will discuss the origin of the misfit valley, and evolution of the local landscapes.

Much of central Nova Scotia is drained today through the Shubenacadie River system (Fig. 19) which runs 86 km from the head of Grand Lake into the Minas Basin. At the end of the last ice age between 10 and 12 ka (11.5-14 CAL) glaciers twice blocked the Shubenacadie River outlets and the lakes that formed in the Shubenacadie and Musquodoboit valleys have been named Glacial Lake Shubenacadie 1 and 2 (Stea and Mott 1998). Clay deposits occur sporadically throughout the valleys, deposited during the glacial lake phases. One lake outlet was through a chain of lakes in Dartmouth, Nova Scotia, that end up in Halifax Harbour (Fig. 19). Another outlet was through the Gibraltar gorge, and the Musquodoboit River (Fig. 19). In the digital elevation model (DEM) the regions under 30 m elevation are shaded, representing the known levels of Glacial Lake Shubenacadie based on strandlines and clay deposits. It is clear from the DEM that if the Shubenacadie Valley was blocked, the Musquodoboit Valley would have also filled with water, if it was ice free.

The prevailing concept of landscape development in eastern Canada is that of a regionally extensive low-relief surface, or peneplane, that was uplifted and tilted to the southeast, then set in

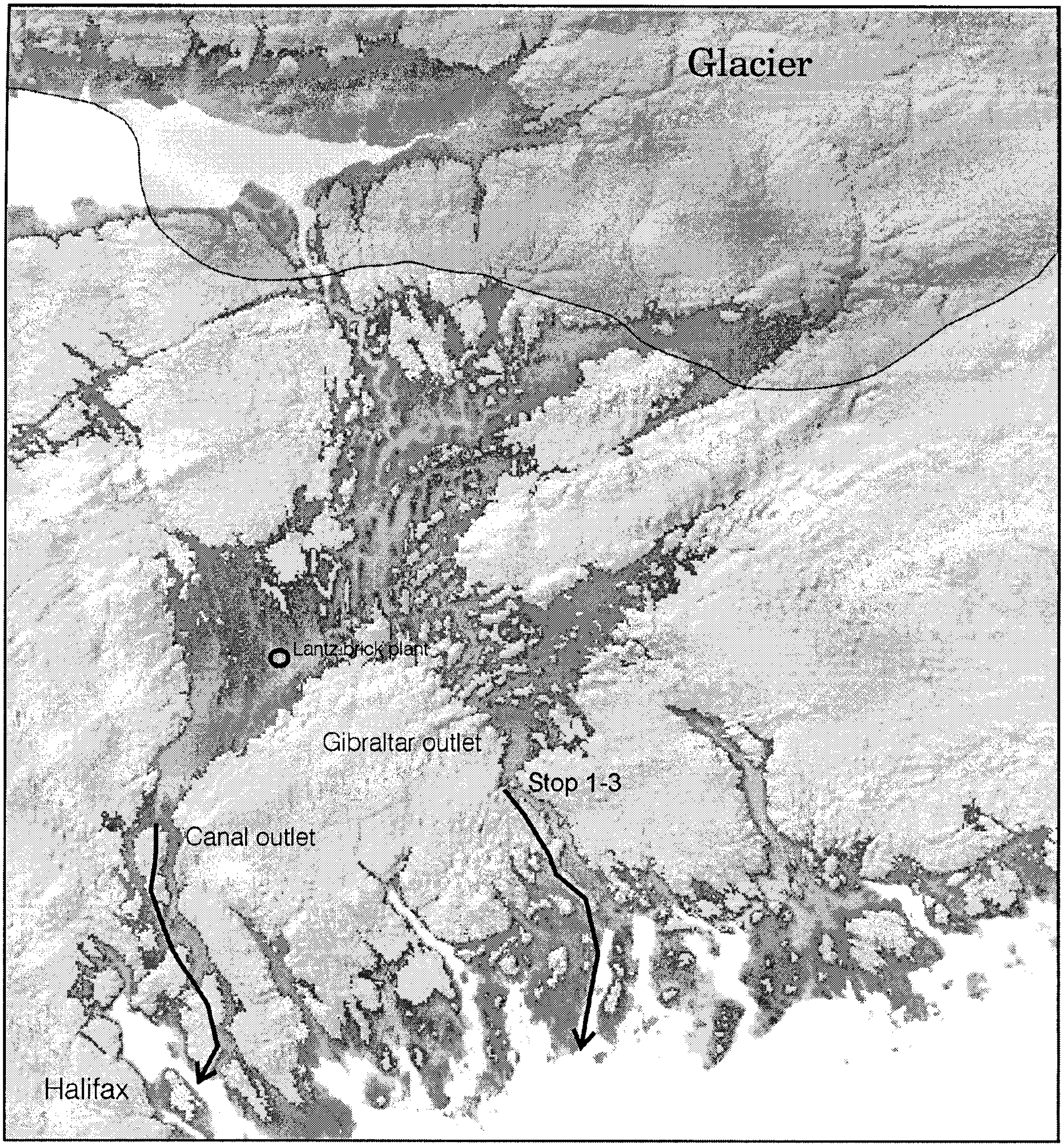


Figure 19. *Digital terrain model of central Nova Scotia. Shaded areas represent the regions under 30m elevation inundated by Glacial Lakes Shubenacadie 1 and 2 and the end of the last ice age. Two main spillways are noted by arrows.*

relief by erosion of weaker rocks (Grant 1994). Goldthwait (1924) proposed that the Musquodoboit River below Gibraltar Rock follows an old southeastward course, an ancient drainage pattern established prior to glaciation and later reoccupied as moraines or drift covered the divide. Lin (1971) came essentially to the same conclusion. Roland (1982) argued that the Musquodoboit River was a tributary of a north-flowing Cretaceous system basing his ideas on the assumption that Early Cretaceous (EK) clay found in the valleys formed after the valleys were first excavated by erosion. Seismic and drillhole data presented by Stea and Pullan (2001) however, led to the conclusion that the erosional model is erroneous. They suggested that the lowland Cretaceous deposits were down-faulted after the Early Cretaceous and the ancient structural valleys were exhumed from under a thick cover of Mesozoic and Cenozoic sediments.

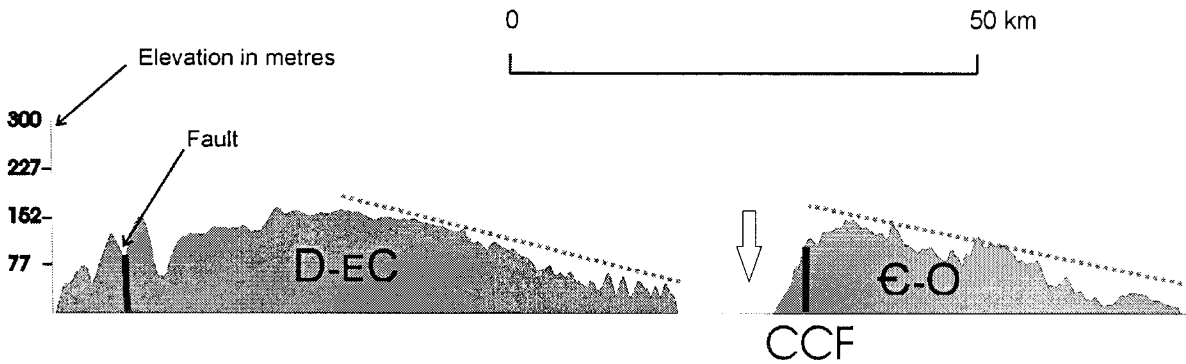


Figure 20*. Fault offset of the Atlantic Uplands Peneplane across the Cobequid Chedabucto Fault system (after Grant 1994;Stea and Horne 2004)*

Using a GIS-generated series of topographic profiles Stea and Horne (2004) further developed Grant's (1994) hypothesis that the Atlantic Uplands Peneplane (AUP) was offset on the fault systems bounding EK outliers (Fig. 20). The AUP also exhibits an inflection point associated with fault systems in southern Nova Scotia, changing to a north-dipping surface. Assuming that these surfaces across the inflection point are coeval, the age of the AUP is younger than Early Jurassic, as it truncates rocks of this age. The lack of Late Cretaceous and Tertiary strata in these structural basins implies that offset-deformation occurred soon after the Early Cretaceous, thus bracketing the AUP between the Late Cretaceous and Early Jurassic. Furthermore, Early Cretaceous deposits can be restored to the AUP surface along basin margin faults. By linking the mature, quartz arenite and kaolin sedimentary deposits characteristic of the EK outliers with the AUP we can envision the peneplane being formed by subareal fluvial processes, weathering and denudation during the Early Cretaceous.

Goldthwait (1924) was partially correct in his assertion that the Musquodoboit River may have re-occupied an old southeastward-flowing river system, but the evidence of an extensive morainal dam in the lowlands to the north is lacking. There are some ablation till deposits in the vicinity of the river, but the hills that cut off the two valley systems to the northeast are rock-cored rather than made up of glacial deposits. It is more likely that the gorge was cut by successive glacial lakes dammed up in the Musquodoboit Valley releasing torrents of meltwater through the relatively narrow outlet at Gibraltar. These lakes formed not only at the end of the last glaciation, but prior to the glaciation as well, as icesheets built up in northern Nova Scotia. Evidence of the clay deposits of pre-Wisconsinan glacial lakes can be found in drillholes and seismic profiles of the Musquodoboit Valley.

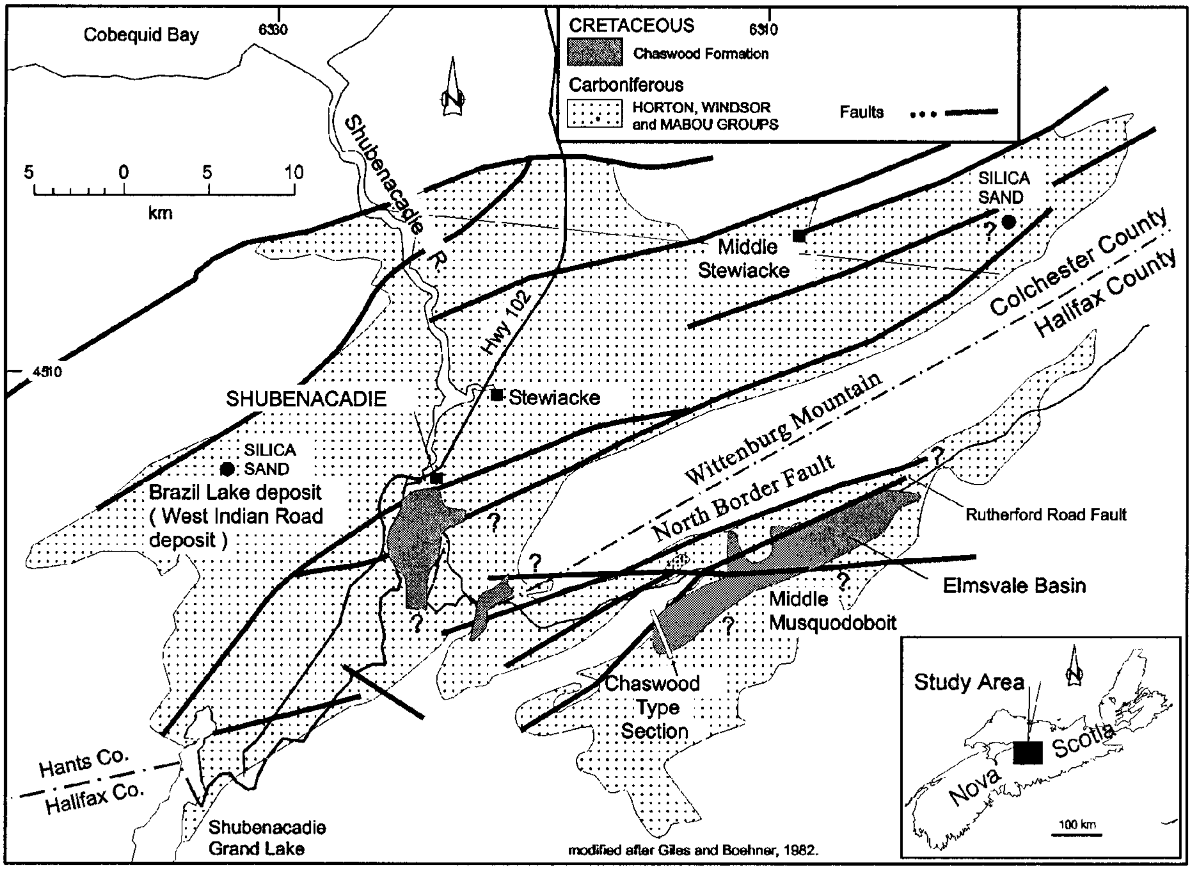


Figure 21. Detailed map of the Chaswood Formation in central Nova Scotia.

STOP 1-4 HIDDEN CRETACEOUS BASIN, CHASWOOD

PURPOSE: To look at seismic evidence for hidden Mesozoic basin remnants and buried Quaternary valleys.

ROUTE: Proceed northward ~22 km along 357 from Stop 1-3 to the town of Middle Musquodoboit. Turn left on 224 and drive 5 km to the Chaswood turnoff. Proceed south 2 km. See map (Fig. 1).

INTRODUCTION

3-D stratigraphic work in the Shubenacadie-Musquodoboit valleys by Stea and Pullan (2001) revealed the existence of "hidden" Mesozoic basin remnants not previously known. These depressions are filled with Early Cretaceous sediments defined as the Chaswood Formation and consist of quartz-rich gravel and sand, fine-grained kaolinitic and organic clay, and lignite. They are "hidden" in deep fault basins, usually obscured by thick Quaternary sediments and only rarely found close to the surface. In the Musquodoboit Valley the Mesozoic deposits cover about 40 km^2, and constitute the "Elmsvale Basin" (Fig. 21). Three "members" have been defined within the Chaswood Formation (Fig. 22), the lower, middle and upper members. The upper and lower

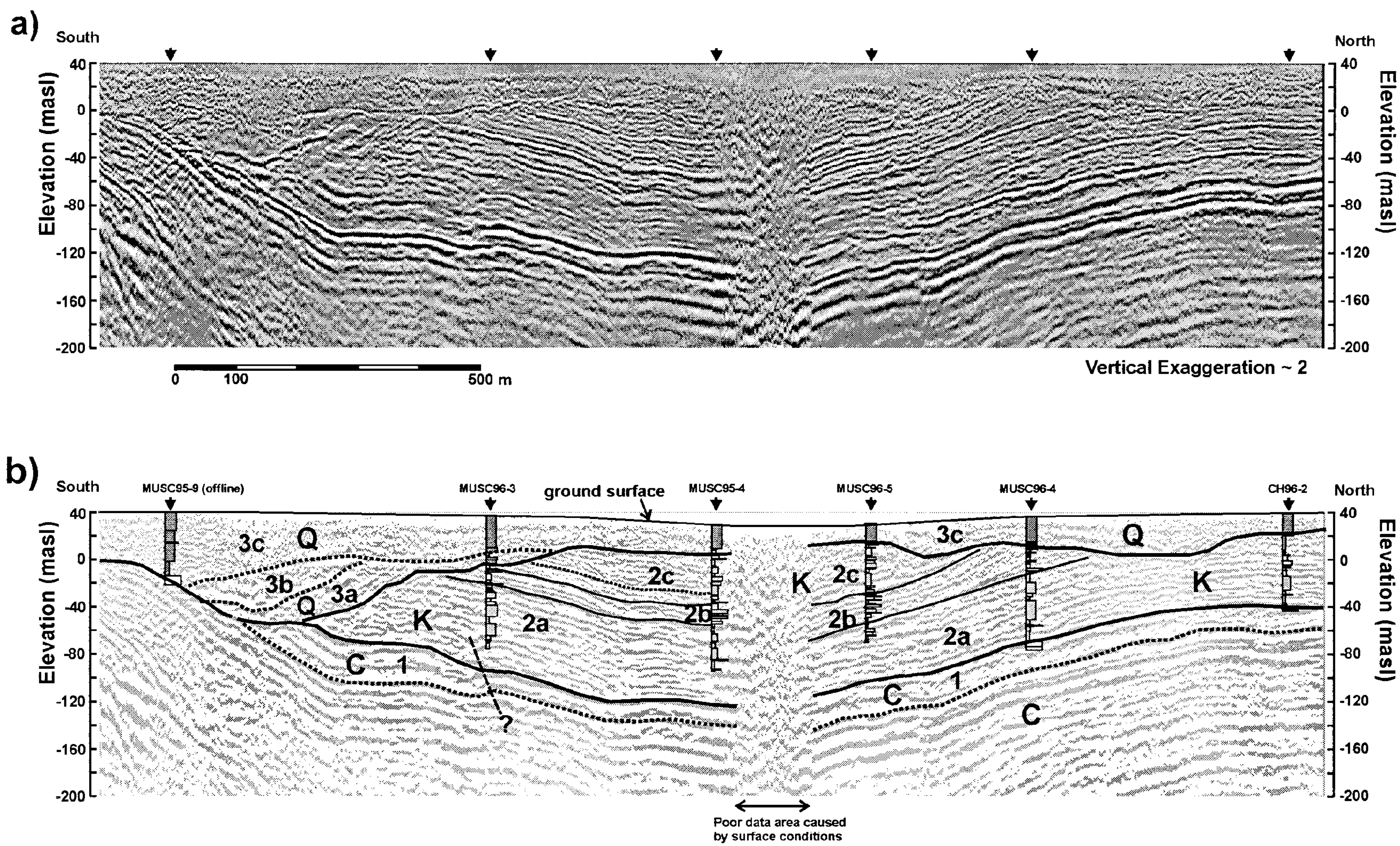


Figure 22. *Chaswood type seismic transect (a) and interpretation (b), displayed in depth. Two-way travel times have been converted to depth and is displayed as actual elevations (masl=m above sea level). The upper profile is an amplitude display where the dark areas are positive amplitude, the same as the dark lines on the variable area display below. Schematic representations of the lithological logs of six boreholes drilled on or close to this seismic line are superimposed on the lower panel. Sequences 1-3 correspond to Carboniferous bedrock (C), Cretaceous sediments (EK), and Quaternary strata (Q).*

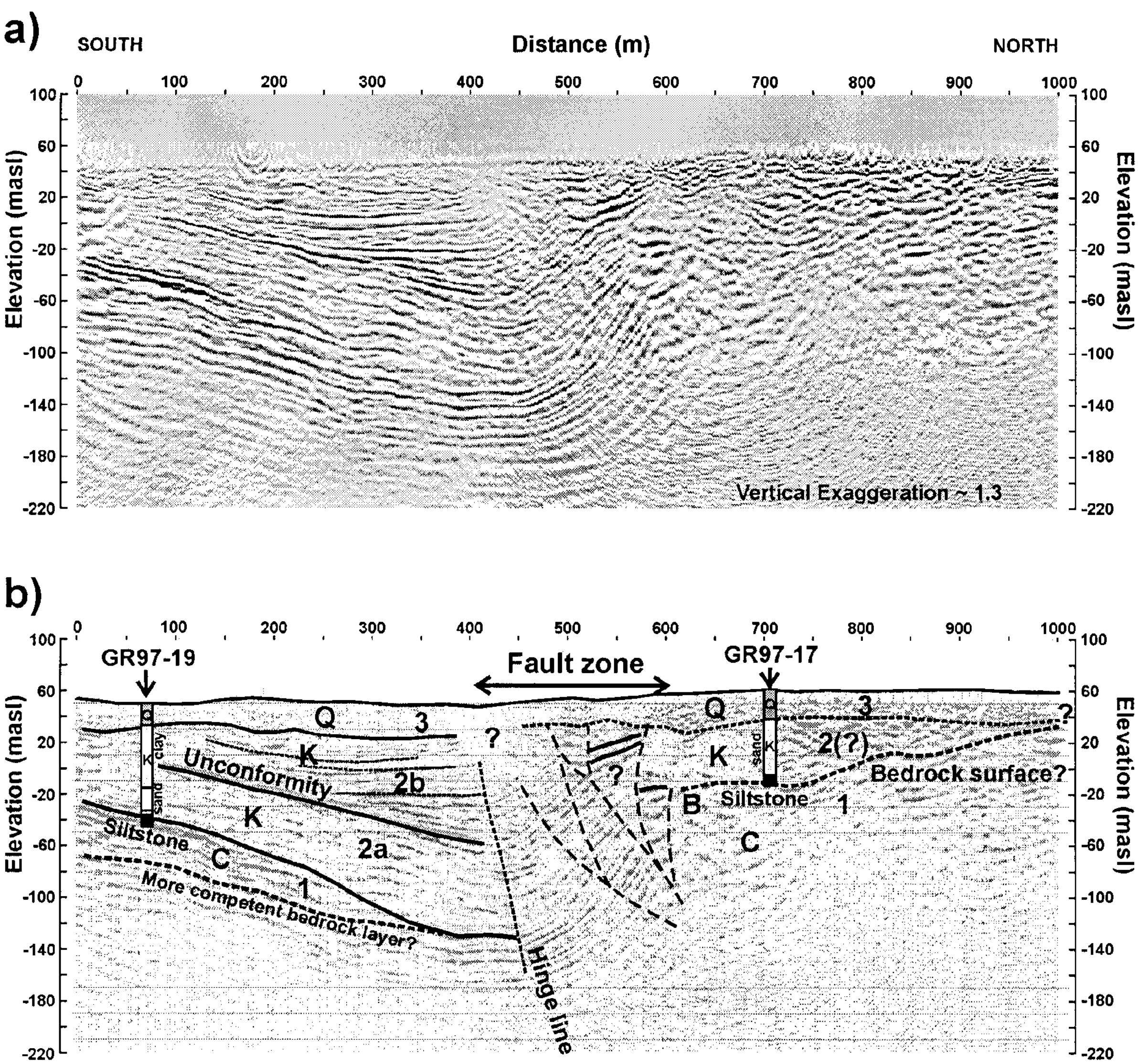


Figure 23. Seismic profile of the fault-bounded northern edge of the Elmsvale Basin. This transect is located along the Glenmore Road in Middle Musquodoboit (Fig. 21). Seismic sequences 1-3 correspond to Carboniferous bedrock (C-1), Early Cretaceous sediments (K-2) and Quaternary strata (Q-3).

members are dominated by inorganic facies and the middle member is characterized by basin-wide lignite beds and organic-rich fine-grained sediments.

We will drive along a valley road which is the site of the type section of the Chaswood Formation; the Chaswood Meadows seismic line and drillhole transect. It is a north-south profile across the Elmsvale Basin approximately 2.5 km in length. In the seismic profile (Fig. 22) we can clearly see a roughly symmetrical basin >2 km wide and 130 m deep overlying Carboniferous bedock (Unit 1) filled with unconsolidated Cretaceous sediments (Units 2a, 2b and 2c). Incised into this larger basin is a smaller, asymmetrical, channel at the south end of the line, approximately 1 km wide and 50 m deep, filled with Quaternary sediments (Units 3a, 3b and 3c). The deepest part of the basin corresponds to the topographic low. To the north of the lie depicted in Figure 22 the Cretaceous section butts up against Cambro-Ordovician bedrock in a near-vertical contact interpreted as a steeply-dipping fault. (eg. Fig. 23)

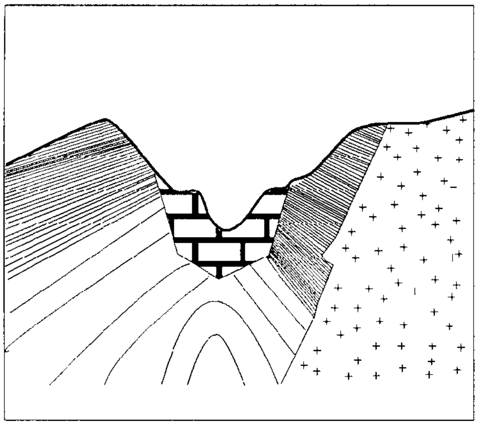
A fault zone (Rutherford Road Fault) has been mapped at the northern margin of the Elmsvale Basin. South of the fault the Chaswood Formation is as much as 140 m thick. Bedrock And Chaswood Formation reflectors terminate abruptly at the fault zone and 100 m of dip slip movement is indicated. The fault zone exhibits several structural styles, including fault-related folds and possibly reverse faults (Fig. 23).

SUMMARY OF LANDSCAPE EVENTS

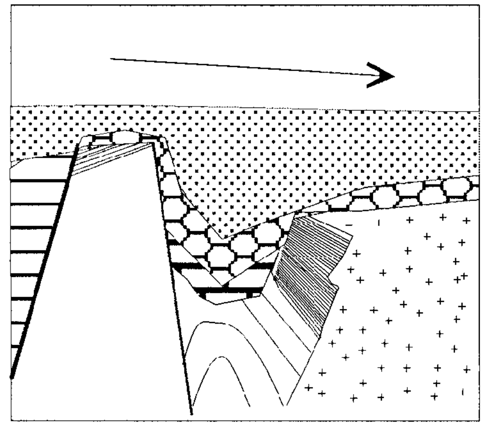
Stea and Pullan (2001) presented a model of landscape development in Eastern Canada termed the structural-exhumation hypothesis, based on the land stratigraphic record. They proposed a substantial cover of Early Cretaceous sediments to explain the preservation of the outliers and the thermal history of the basins (Arne, et. al. 1989). Recent discoveries in New Brunswick (Falcon-Lang et. al. 2003). and northern Nova Scotia (Stea et al. 2003) suggest that Early Cretaceous sediments covered large areas of Eastern Canada, a point reinforced by the petrological similarity of Chaswood Formation sediments (Stea and Fowler 1981; Dickie 1986; Falcon-Lang et. al.. 2003). Within a minimum age range of nearly 17 million years (Barremian-Albian) substantial Early Cretaceous sedimentation across Eastern Canada can be accommodated. The Stea and Pullan (2001) model can be summarized as follows (see also Fig. 24):

1. Early Cretaceous deposition of the Chaswood Formation 140-110 Ma (age range);
2. post-Early Cretaceous faulting->110<80 Ma;.
3. Mid-Late Cretaceous-Early Tertiary uplift/exhumation;
4. Late Tertiary subsidence; and
5. Quaternary deposition.

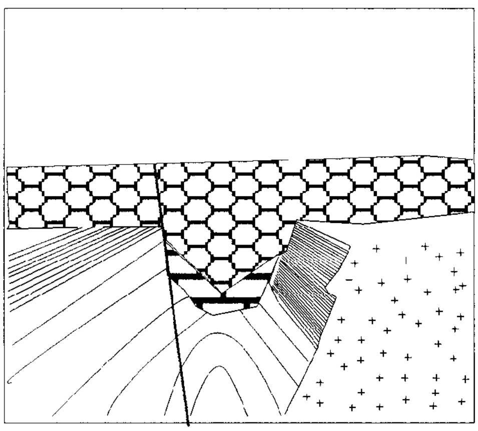
Figure 24 (next page). *Structural-exhumation hypothesis for the evolution of Nova Scotia landscapes. (a) Early Carboniferous sediments deposited in continental-epeiric basins, with (b) pediplain formation in the Late Carboniferous. (c) Triassic-Jurassic rifting and landscape rejuvenation. Deposition of Triassic-Jurassic sediments in Fundy Basin. (d) Deposition of thick Early Cretaceous sediments in a low relief coastal plain fluvial environment and north-south regional consequent drainage. Residuum in upland areas feeding the deltas provides kaolin and quartz. (e) Mid Cretaceous diastrophism creates or reactivates basin faults creating structural valleys (eg. Elmsvale Basin) followed by further Tertiary uplift and denudation. Subsequent valleys formed. (f) Quaternary modification, valley incision.*



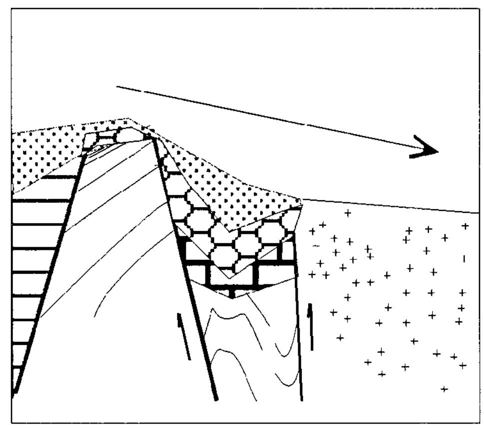
Early Carboniferous



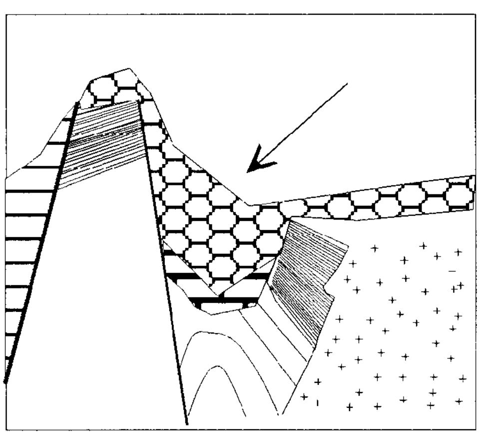
Early -Mid Cretaceous



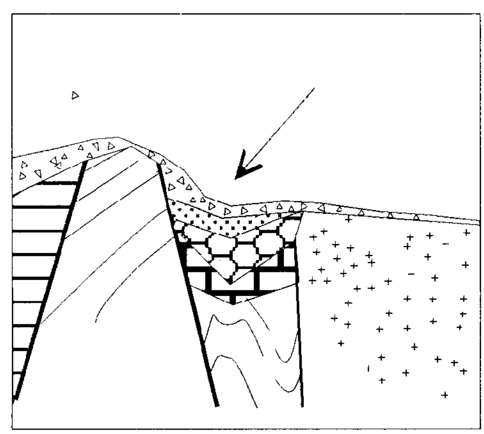
Late Carboniferous



Mid-Cretaceous-Tertiary



Triassic-Jurassic



Quaternary

STOP 1-5: EAST MILFORD QUARRY

PURPOSE: To examine the Quaternary stratigraphic record, especially the end of the last interglacial preserved in karst depressions in the National Gypsum quarry.

ROUTE: Return to highway 224 and proceed westward ~10 km to Gays River. Continue westward on highway 277 until the village of Carrolls Corner. Turn northward on secondary road and continue for 6 km to highway 2. Turn left and head eastward for 4 km. Turn onto quarry road on left. See map and details (Fig. 1).

INTRODUCTION

A few years ago the nearly complete remains of a mastodon were found by a grader operator while he was cleaning the karst gypsum surface in preparation for mining. The excavation of the mastodon was supervised by Bob Grantham, former curator of geology for the Nova Scotia Museum. Over the past 38 years, since the quarry first opened, interglacial/interstadial forest bedshave been frequently unearthed by the stripping operations. The quarry was a field stop during the 1972 and 1987 INQUA congresses. It has been described by Hughes (1957), Take (pers. comm., 1964), Grant (1975), Prest (1977) and Mott et al. (1982). Numerous sections have been uncovered and then destroyed by stripping at the site since 1954 when the pit first opened. We will summarize the observations made by these authors at all these sections.

QUATERNARY STRATIGRAPHY

The oldest recognized surficial sedimentary unit is a black,lignite-bearing clay with thin sand laminae, plastered inside the gypsum sink holes (Fig. 25). Veinlets of gypsum penetrate this unit, which is Cretaceous in age, based on fossil pollen (R. Fensome, personal communication, 2002). The black lignitic clay is similar to Cretaceous deposits at Gays River, Shubenacadie and Musquodoboit in the Hants Lowlands. Dark grey, rubbly or gypsiferous tills have been described on the gypsum surface and at the base of some karst channels. The lowest till is called the Miller Creek Till, named after a type locality in a gypsum quarry near Windsor, Nova Scotia. At the type section the Miller Creek Till underlies a thick peat and wood layer and has soil developed on it, suggesting an interglacial interval of some duration. Within the karst sinkholes the apparent stratigraphic relations are unclear due to solution collapse or subglacial deformation. Infilling some sinkholes is a pebbly clay with shell horizons, and peat lenses. Wood fragments are found throughout, but concentrated in the upper peaty layer. The sinkhole organics include gastropod and pelecypod shell layers and small wood fragments including some beaver-cut sticks and bones. In 1991 Stanley McMullin noticed a tusk in some fill excavated from a sinkhole. Scientists from the Nova Scotia Museum eventually uncovered the nearly complete remains of an adult and juvenile Mastodon (*Mammut americanum*). These include skull, (with tusk and teeth

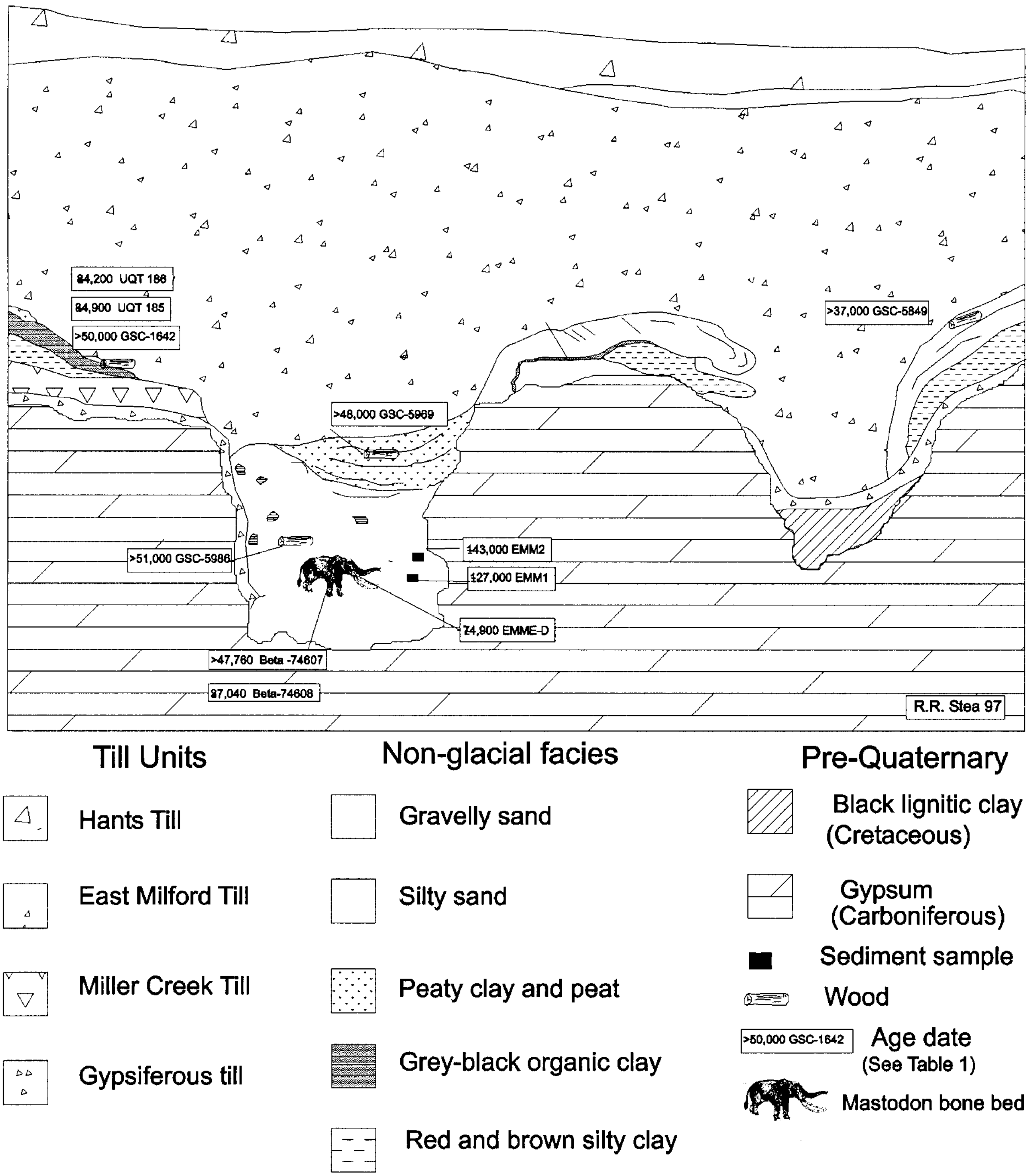


Figure 25*. Schematic drawing of the stratigraphy of the East Milford Quarry. Based on observations from R. R. Stea and R. J. Mott from 1970-2003.*

disarticulated), scapula, thoracic vertebra, ribs and femur bone fragments (Grantham and Kozera-Gillis 1992). A few pollen samples from the sinkhole organics revealed a homogenous conifer-dominated assemblage, consistent with a cool boreal forest environment. Wood (*Pinus banksiana*) found near the mastodon bones produced a radiocarbon age of >51 ka (Table 3).

Peat layers above the gypsum surface can be divided into two pollen zones, a lower zone containing hardwoods, including *Tilia* (basswood), *Fagus* (beech), and *Acer* (maple),and an upper zone characterized by coniferous pollen, mainly *Abies* (fir). The presence of the carabid beetle,(*Pterostichus patruelis*) in the upper zone, with a range in northern Quebec today, indicates a cooler boreal climate. Uranium series ages of 84.2 ka and 84.9 ka were obtained using wood fragments from the upper unit (Causse and Hillaire-Marcel 1986). These organic deposits probably span a lengthy nonglacial interval. In their syntheses, Mott and Grant (1985) and deVernal et al. (1986) differentiated three types of pollen sequences within the organic zones of interglacial deposits throughout Nova Scotia, termed Palynostratigraphic Units 1-3. Unit 1 spectra are characterized by pollen indicative of climatic conditions warmer than present with forests containing white pine and hardwoods. At East Milford, the pollen record is restricted to Palynostratigraphic Units 2 and 3. Unit 2 indicates a climate similar to the present and Unit 3 is characterized by cooler boreal-forest to tundra conditions.

Overlying the organic sediments at the base of the section are two to three tills and waterlain sediments. Organic horizons are overlain by glacio-tectonized silty-clay, sand and gravel containing fossil wood. The overlying reddish silty till, called the East Milford Till, forms much of the topography of the Hants Lowlands averaging between 10 and 20 m in thickness. It has a northern Nova Scotia provenance and formed during the Early to Middle Wisconsinan Caledonia Phase (Fig. 6). Overlying the East Milford Till are clay, silt, sand and gravel beds and the surface Hants Till (Stea 1984), which was initially deposited by southward-flowing Escuminac Phase glaciers (correlative with the Lawrencetown Till on the Atlantic coast) and later reworked by local ice caps and divides.

INTERPRETATION

The forests at East Milford grew during the Sangamonian Interglaciation (125-75 ka). Deposits of the last interglacial are beyond the 50 ka limit of radiocarbon dating. U/Th age dates on wood seem to confirm a Sangamon age for most of the beds, but the youngest palynological unit (Unit 3) produced some Mid-Wisconsinan ages (Mott and Grant 1985; de Vernal et al., 1986). Because of the possibility of parent U isotope migration, the younger dates are thought to be minimum ages (Stea et al.. 1992b). Recent dating of the sinkhole organics by optically-stimulated luminescence has produced ages in the range of 135 ka, slightly older than the last interglacial (Godfrey-Smith et al. 2003). Mastodon tooth enamel, however, produced an age of 74.5 ka, closer to the U/Th wood ages. From this discrepancy, Godfrey- Smith et al. (2003) concluded that the mastodon and older peat mixed together as the animal sank in the mire. This mixing of young bones and older peat was later augmented by glacial loading and tectonism.

A question that one could ask is where are the tills/organics/paleosols relating to previous glacial-interglacial cycles. Unconsolidated Early Cretaceous sediments have been found on the karst gypsum surface directly overlain by Sangamonian organic sediments with no glacial deposits intervening. Twenty-two glaciations have occurred in the last 2 million years in Atlantic Canada, but sedimentary records pre-dating the Illinoian (Marine Oxygen Isotope Stage 6) have only been documented at one locality (Table 1; Wehmiller et al., 1988)

The East Milford section records landscape development since the Mesozoic and the geological events recorded by the section can be summarized as follows:

(1) Karst development (uplift and erosion?) before the Early Cretaceous
(2) Deposition of Mesozoic sediments
(3) Glacial advance, probably during the Illinoian Glaciation (Miller Creek Till)
(4) Glacier retreat, global warming and climatic fluctuations during the Sangamonian Interglaciation.
(5) Glacial advances during the Wisconsinan Glaciation (East Milford and Hants Tills)

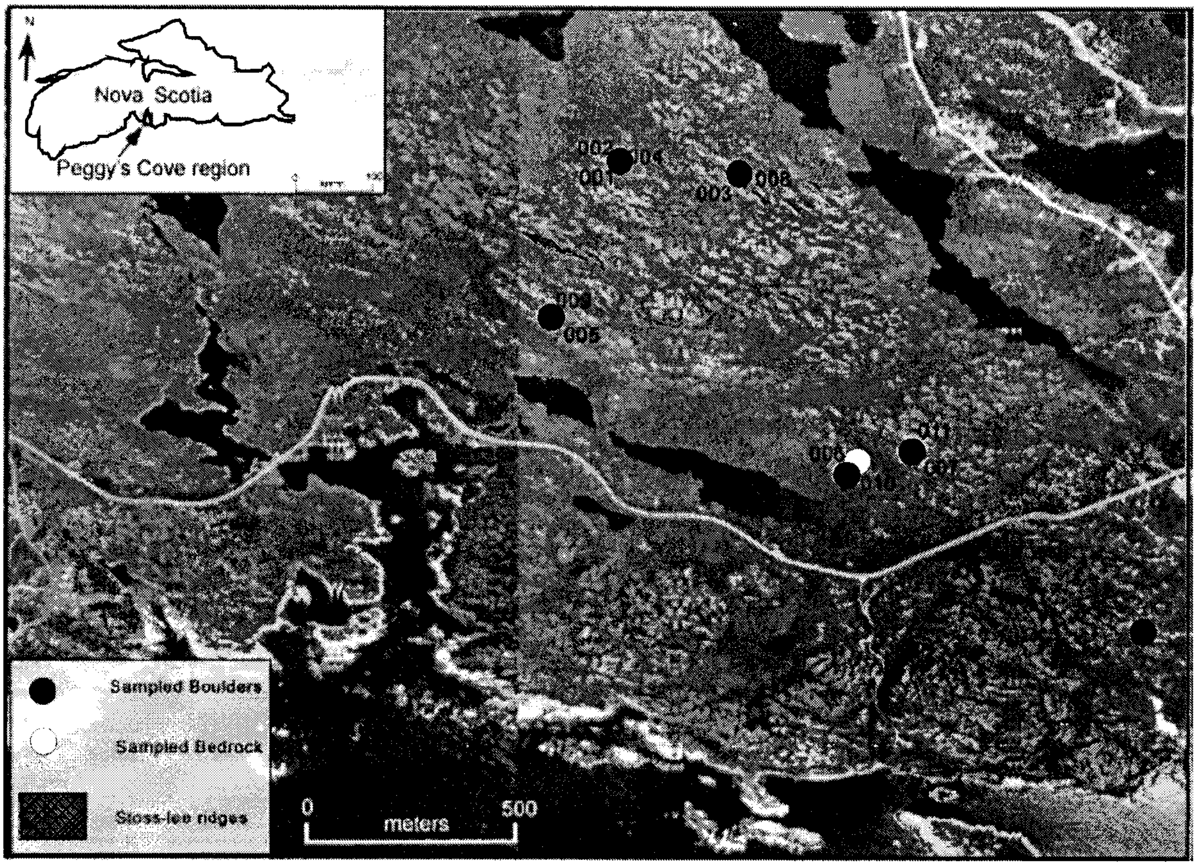


Figure 26*. Location of boulder (001 to 007) and bedrock (008 to 011) sample sites for exposure dating in the Peggys Cove region.*

STOP 1-6: GLACIAL GEOCHRONOLOGY OF PEGGYS COVE

PURPOSE: At this locality we will describe the cosmogenic nuclide exposure chronology on erratic boulders, and place them into a regional deglacial context. The sampling strategy for exposure dating will be discussed. The implications for glacial erosion of exposure ages on bedrock ridges will be presented. We will also consider the possibility that the boulders here represent a large flood event.

ROUTE: See map (Fig. 1). This stop will involve a 2 hour hike over low relief. Although most of the hike will be along a well used trail, some parts of the trail may be marshy, so waterproof boots are recommended.

EN ROUTE:

Before or after the field stops we will visit the Peggy's Cove Lighthouse (officially the Peggy's Point Lighthouse). The current lighthouse was built in 1915, replacing the original lighthouse that was erected in 1868. Please heed safety warnings of the potential for waves to unpredictably reach dangerous heights.

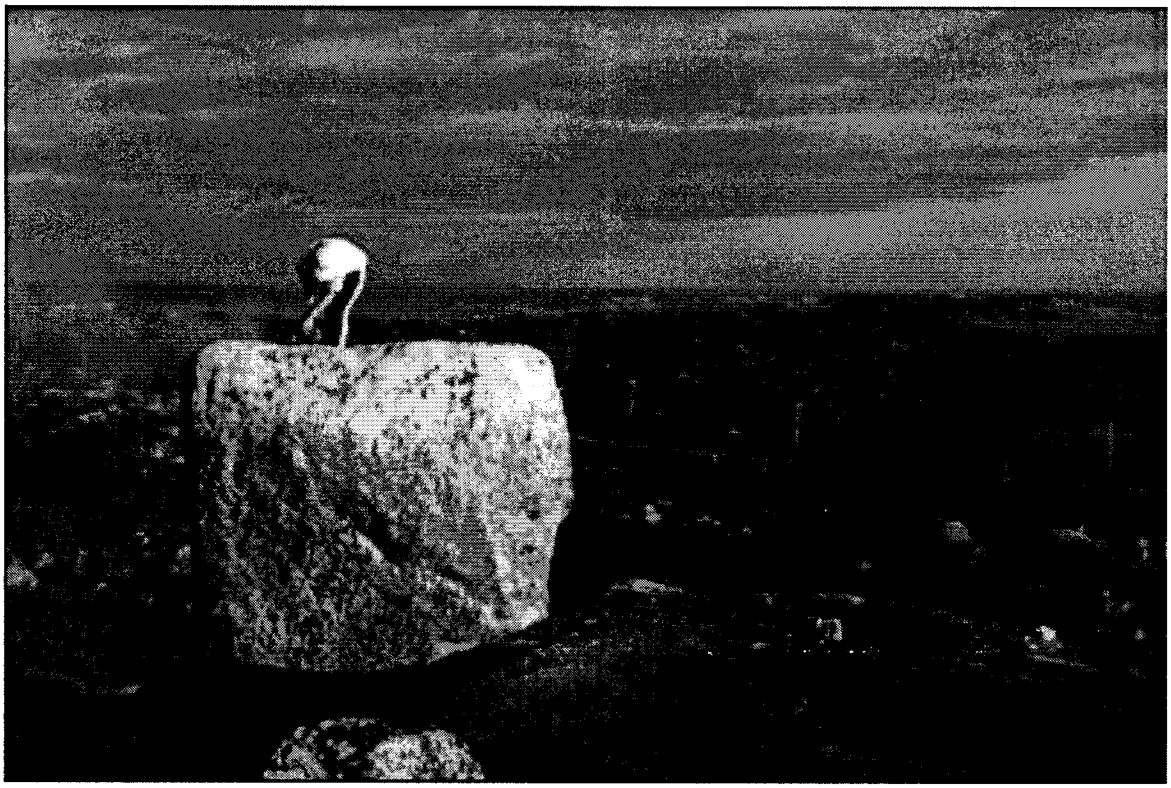


Figure 27. *Erratic boulder sample NS-Peggy-02-07. This boulder has an ideal geometry (>2m height, horizontal flat upper surface, resting on a local protrusion to minimize snow cover). Samples were collected with a gas-powered 10" diamond blade cut-off saw. All samples were a minimum of 60cm from the nearest edge of the boulder.*

TIME OF DEGLACIATION BASED ON BOULDER EXPOSURE AGES

Unlike its offshore counterpart with radiocarbon datable media, the record of onshore deglaciation has limited time control. Useful lake records are uncommon and exposures of macrofossils preserved in tills are rare. OSL has been used in some coastal localities with variable success. As part of the mandate of the Atlantic Canada paleo-Ice Dynamics (ACID) project, cosmogenic nuclide exposure dating will be used in regions where suitable sampling material exist and where the chronology can provide meaningful information in the regional glacial or deglacial context. The first exposure dating experiment in Nova Scotia used ^{10}Be measured in quartz and was attempted here at Peggy's Cove for the following reasons:

1. Records of deglaciation of the Scotian Shelf (e.g. King and Fader 1986;Gipp and Piper 1989 Gipp 1994;) had been previously documented and formed a basis for the regional glacial history compiled by Stea et al. (1998). The shelf record indicated that ice may have persisted in the area as late as 17.4 ka (Stea et al. 1998). The area was known to be ice free well before the Younger Dryas chron, 13.5 ka (Miller and Elias 2000; Stea et al. 1998; Spooner, et al. in press;) based on lacustrine records and the age of the now submerged Sambro Delta, 50 km east of Peggy's Cove. We are currently examining the possibility that a post-Chignecto readvance may have occurred in

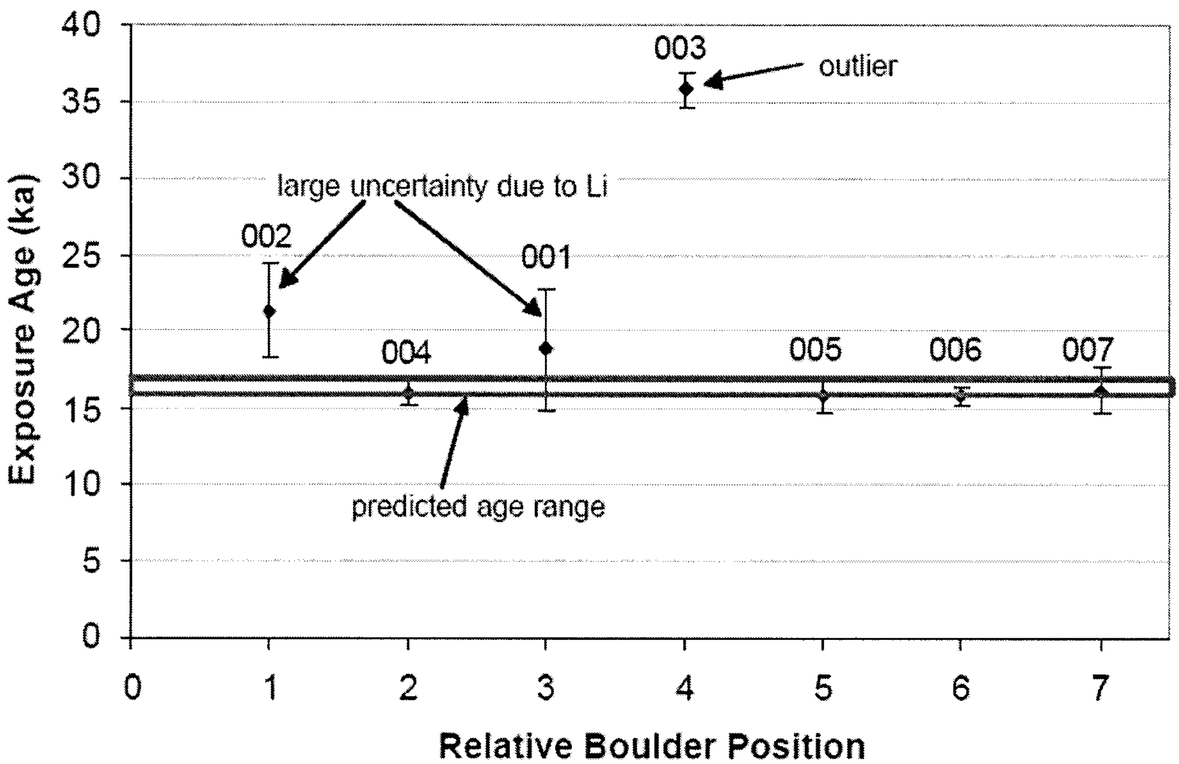


***Figure 28**. Exposure ages of the seven Peggy's Cove boulders sampled in 2002 (McDonald 2003). Calculations assume negligible erosion or burial of the samples, and are based on a cosmogenic ^{10}Be production rate of 5.1 atoms g-1 yr-1 scaled according to Lal (1991) and Stone (2000) for atmospheric and geomagnetic field affects. Relative boulder position refers to the location of the boulders in an ice flow direction (sample NS-Peggy-02-007 is farthest down flow, and closest to the coast). Sample NS-Peggy-02-003 is a statistical outlier. Samples NS-Peggy-02-001 and 002 have large 1s uncertainties due to AMS difficulties believed to be related to high Li-content of the samples. Predicted age range is based on the estimated timing of ice marginal retreat after the Chignecto Phase ice advance (Stea et al. 1998), and on a glacial readvance date of 15.9 ±? 0.3 ka cal. from Cape Breton Island (Josenhans and Lehman 1999).*

this southeastern quadrant of the Nova Scotia ice cap, coeval with the 15.9 ka readvance in the Laurentian Channel (Josenhans and Lehman 1999).

2. The quartz-rich Halifax Pluton granite has a low fracture density that has yielded boulders up to 3 m in diameter. Large (> 2 m height) boulders on morainal or bedrock ridges are necessary for exposure dating as there is a lower probability that the boulders were exhumed sometime after their deposition, they have negligible snow cover, rolling is unlikely, and they have a negligible fast-neutron leakage problem because the rock edges can be adequately avoided.

3. Other favorable conditions include minimal glacial sediment cover (veneer averages less

Figure 29*. Sample NS-Peggy-02-010 is a bedrock ridge sample collected from the top of a striated granite ridge (ice flowed from left to right). Bedrock exposure samples were collected at least 15 cm from the nearest edge, on the flattest portion of the highest ridges in areas with varying fracture densities.*

than 1 m), a simple ice flow history documented by cross-cutting striations (SE and SW-SSW with uncertain relative ages), and no marine incursion above the current sea level. Despite its coarse grain size and proximity to salt water, the post-glacial erosion of the granite surfaces is negligible (dike protrusions are typically 1 cm above the host surface).

The experiment formed the basis of Fiona McDonald's honour's thesis (McDonald 2003). Seven boulders were sampled using a chisel and 10" diamond blade cutoff saw (Fig. 27). The ^{10}Be exposure ages on the boulders are presented in Fig 28. The 16.4 ± 1.7 ka weighted mean age of the boulders ignores the obvious outlier which at 35 ka must have contained a ^{10}Be concentration that was inherited from exposure prior to glaciation. Ignoring the two samples with large AMS uncertainty due to a beam-interference by Li, the mean age is 15.9 ± 0.2 ka. Based on the observed post-glacial erosion, the effect of erosion on the boulder exposure ages would be less than 0.5%, so these ages are not likely a minimum age, but probably reflect close to the actual time of ice marginal retreat. Glacial erosion of the Halifax pluton

The hundreds of unweathered granitic boulders strewn across this landscape may be taken to indicate that glacial erosion by plucking significantly denuded the Halifax pluton during the last glaciation. How much glacial denudation occurred? Exposure ages of the bedrock ridges (Fig. 27 and 28) have provided some insights into the controls and amount of erosion during the last glaciation. The approach here is to determine if the cosmogenic nuclide clock was completely reset during the last glaciation. Cosmic rays are attenuated by a rate of 160 g/cm^2, which means that the flux will half (specifically, it will diminish at a rate of 1/e) every 60 cm in granite. If the glacier did not erode the bedrock, the concentration in we measure would be the sum of post-glacial and pre -glacial nuclide production. If the glacier eroded the granite by more than 3.6 m, there would be an insignificant amount of inherited ^{10}Be relative to the post glacial ^{10}Be concentration, and the bedrock ages would equal the boulder ages (15.9 ka). Granted, the very fact that the ridges form positive relief indicates that they may be less eroded than the surrounding depressions. However, we cannot exposure date the depressions due to the difficulties in compensating for an unknown amount of shielding by till, water, and snow. Four samples were collected from bedrock ridges that are at least 1.5 m above the surrounding ground level. From youngest to oldest, the bedrock ages are 18.0 ±1.0, 35.7 ± 1.7, 92.7 ± 1.8, 102 ± 3.4 ka. These ages are significant because they show::

1. all of the ridges have at least a little inheritance compared to the boulder ages.

2. there was no indication of burial by till or marine sediment that has been since eroded, because no ages are younger than the boulder ages.

3. inherited ^{10}Be implies that the glacier eroded less than 3.6 m of granite on these ridges. Table 4 provides an estimate of glacial erosion rate, assuming a 75 ka exposure period prior to the last glaciation.

Table 4: Modelled glacial erosion rates of bedrock ridges (McDonald, 2003)

Field ID	Z (cm)	Glacial erosion (cm/kyr)
NS-PEGGY-02-008	7	0.1
NS-PEGGY-02-009	85	1.5
NS-PEGGY-02-010	263	4.5
NS-PEGGY-02-011	0	0

Based on these few samples it is difficult to say anything substantial. However, F. McDonald determined that the sites with negligible erosion had the lowest fracture density. Perhaps then, a high fracture density makes it easier for a glacier to pluck blocks. More measurements are clearly needed.

(Note: For colour photos and extra information tour the Cosmogenic Nuclide lab web site! http://www.dal.ca/~cnef/

Age Date Tables

Section/Core	Date	Type	Material/deposit	LabN	Source
Pre- Sangamon (MIS 5) Dates					
(1) Gilbert Cove, N.S	>250000	AAR	Shell in till	------	Wehmiller *et al.*(1988)
Sangamon Interglaciation (Oxygen Isotope Stage 5) dates (pre Phase 1 Tills)					
(2) Miller Creek Quarry	>52,000	^{14}C	Wood under till	GSC-2694	Stea *et al.*(1992b)
(3) Leitches Creek	>52,000	^{14}C	Peat under till	GSC-2678	Grant (1989)
(4) Big Brook	>49,000	^{14}C	Wood in lowest till	GSC-3289	Mott & Grant(1985)
(5) East Milford Quarry	>50,000	^{14}C	Wood under till	GSC-1642	Mott & Grant (1985)
(5) East Milford Quarry	84.0±6.5ka	U/Th	Wood under till	UQT-185	deVernal et al.(1986)
(5) East Milford Quarry	74.9±6.5ka	ESR	Dental enamel	----	Godfrey-Smith et al.(2003)
(6) Cape George	80-110 ka	AAR	Shell in till	----	Stea *et al.*(1992b)
(6) Cape George	99-184 ka	ESR	Shell in till	----	Stea *et al.*(1992b)
Mid-Late Wisconsinan recession? (Stage 3/2) dates (pre-Phase 2/3/4 Tills)					
(7) Bay St. Lawrence	21,920±150	^{14}C	Shell in glaciomarine	TO-246	Grant (1994)
(8) Dingwall	23,700±560	^{14}C	Wood in till	GSC-3381-1	Grant (1994)
(8) Dingwall	32,700±560	^{14}C	Wood in till	GSC-3381-2	Grant (1994)
(9) South Aspy River	24,900±700	^{14}C	Organic sediment	I-3414	Newman (1971)
Late Wisconsinan recession (Stage 2) (Post Phase 3- pre-Phase 4 advance)					
(10) Spencers Island Delta	13,400±300	^{14}C	Shell in bottomsets	BETA13728	Stea & Wightman (1987)
(11) Brier Island Lake	13,200±130	^{14}C	Shell in sand	GSC-4431	Stea & Mott (1998)
(12) Leak Lake	13204	^{14}C	Basal age regression		Stea & Mott (1998)
Late Wisconsinan recession (Allerød) (post Phase 4- pre-Younger Dryas readvance)					
(13) Collins Pond (top)	10,900±100	^{14}C	Peat under till	GSC-4475	Mott & Stea 1994
(14) Millbrook	10,900±110	^{14}C	Wood under till	GSC-6435	Stea & Mott (unpub)
(15) Lake Road	10,800±100	^{14}C	Wood under till	GSC-6419	Stea & Mott (unpub)
(16) Lantz (top)	10,900±90	^{14}C	Peat under lake clay	GSC-3771	Stea & Mott, 1989
(17) Shubenacadie (top)	10,800±100	^{14}C	Peat under sand	GSC-3981	Stea & Mott, 1989
Final Deglaciation (Holocene) (post Younger Dryas)					
(18) Cormier Lake (base)	9970±80	^{14}C	Wood in gyttja	BETA61401	Stea & Mott (1998)
(19) Baddeck Bog (base)	9100±100	^{14}C	Gyttja	GSC-4865	Grant, 1994

Table 1. *Terrestrial radiocarbon dates from Nova Scotia from several sources. Note that these are only selected dates relating specifically to ice margins from much larger lists summarized in the references provided in Stea et al. (1998) and Grant (1989).*

Age Date Tables (con't)

Caledonia Phase advance limit? or older (MIS 6)					
Section/Core	Date	Type	Material/deposit	LabN	Source
Wedge 2 (Fig. 6)	~70 ka		Sedimentation rate		Mosher *et al.* 1989
Mid-Late Wisconsinan recession (MIS 3/2) (pre-Phases 2/3/4)					
(2) G046	33,970±560	^{14}C	Shell in channel gravel	TO-753	Amos & Miller 1990
(2) G046	32,130±270	^{14}C	Shell in channel gravel	TO-754	Amos & Miller 1990
(2) G046	32,200±290	^{14}C	Shell in channel gravel	TO-755	Amos & Miller 1990
(2) SAB-85	37,210±400	^{14}C	Shell in Seismic unit 2	RIDDL-639	Boyd *et al.* 1988
(3) 90002-05	27,990±220	^{14}C	Shell in shelf sand	TO-2038	Piper & Fehr 1991
Escuminac Phase advance limit (Phase 2 -MIS 2)					
(1) 90-015-1	20,780±170	^{14}C	Shell in debris flow	TO-2077	Baltzer *et al.* 1994
Scotian Phase advance limit (Phase 3)					
(4) 79-011-11	14,850±170	^{14}C	Shell in Emerald Silt A	None	King ,1996
(5)88-010-007	17,450±155	^{14}C	Shell In Emerald Silt	Beta-27229	Piper *et al.* 1990
Chignecto Phase advance limit (Phase 4)					
(6) 91-018-53	13,050±140	^{14}C	Shell in Emerald Silt	GX-17635	Stea *et al.* 1996
(7)HU90-028-10	13,650±80	^{14}C	Shell neart till tongue	OS-4865	Josenhans & Lehman, 1999

Table 2. *Marine radiocarbon dates from the Scotian Shelf from several sources. Note that these are only selected dates relating specifically to ice margins from much larger lists summarized in the references provided. These dates are reported without marine reservoir corrections.*

Age Date Tables (con't)

Age dates East Milford Quarry					
No.	Age	Material	Species	Unit	Lab Number
1	>33,800	wood	?	6?	GSC-33
2	>50,000	wood	*Larix sp.*	6	GSC-1642
3	>38,000	wood	*Pinus banksiana*	4	GSC-5308
3	>51,000	wood	*Pinus banksiana*	4	GSC-5986
4	>48,000	wood	?	4	GSC-5969
5	>37,000	wood	Deciduous?	7	GSC-5849
6	>47,760	bone collagen	*Mammut sp*	4	Beta-74607
7	37,040 ± 1730	bone collagen	*Mammut sp*	4	Beta-74608
8	33,810 ± 380	bone	*Mammut sp*	4	TO-2145
9	8610 ± 60	wood	?	9?	GSC-6010
Uranium-Thorium disequilibrium ages					
9	84,900 ± 6500	wood	*Picea sp.*	6	UQT-185
10	84,200 ± 11,300	wood	*Abies balsamea*	6	UQT-186
Optical Luminescence ages					
11	127,000 ± 13,000	sediment	-	4	EMM1
12	143,000 ± 16,000	sediment	-	4	EMM2
Electron Spin Resonance ages					
13	78,200 ± 7300	tooth enamel	*Mammut sp*	4	EMM-Jdep
14	77,000 ± 7300	tooth enamel	*Mammut sp*	4	EMM-Jndep

Table 3. *Age dates from the East Milford Section. Dates derived from Mott and Grant (1985), deVernal et al. (1986); Godfrey Smith et al. (2003) Stea, unpublished data.*

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Pre-conference Field Trips

A1 Contamination in the South Mountain Batholith and Port Mouton Pluton, southern Nova Scotia
D. Barrie Clarke and Saskia Erdmann

A2 Salt tectonics and sedimentation in western Cape Breton Island, Nova Scotia
Ian Davison and Chris Jauer

A3 Glaciation and landscapes of the Halifax region, Nova Scotia
Ralph Stea and John Gosse

A4 Structural geology and vein arrays of lode gold deposits, Meguma terrane, Nova Scotia
Rick Horne

A5 Facies heterogeneity in lacustrine basins: the transtensional Moncton Basin (Mississippian) and extensional Fundy Basin (Triassic-Jurassic), New Brunswick and Nova Scotia
David Keighley and David E. Brown

A6 Geological setting of intrusion-related gold mineralization in southwestern New Brunswick
Kathleen Thorne, Malcolm McLeod, Les Fyffe, and David Lentz

A7 The Triassic-Jurassic faunal and floral transition in the Fundy Basin, Nova Scotia
Paul Olsen, Jessica Whiteside, and Tim Fedak

Post-conference Field Trips

B1 Accretion of peri-Gondwanan terranes, northern mainland Nova Scotia and southern New Brunswick
Sandra Barr, Susan Johnson, Brendan Murphy, Georgia Pe-Piper, David Piper, and Chris White

B2 The Joggins Cliffs of Nova Scotia: Lyell & Co's "Coal Age Galapagos"
J.H. Calder, M.R. Gibling, and M.C. Rygel

B3 Geology and volcanology of the Jurassic North Mountain Basalt, southern Nova Scotia
Dan Kontak, Jarda Dostal, and John Greenough

B4 Stratigraphic setting of base-metal deposits in the Bathurst Mining Camp, New Brunswick
Steve McCutcheon, Jim Walker, Pierre Bernard, David Lentz, Warna Downey, and Sean McClenaghan

B5 Geology and environmental geochemistry of lode gold deposits in Nova Scotia
Paul Smith, Michael Parsons, and Terry Goodwin

B6 The macrotidal environment of the Minas Basin, Nova Scotia: sedimentology, morphology, and human impact
Ian Spooner, Andrew MacRae, and Danika van Proosdij

B7 Transpression and transtension along a continental transform fault: Minas Fault Zone, Nova Scotia
John W.F. Waldron, Joseph Clancy White, Elizabeth MacInnes, and Carlos G. Roselli

B8 New Brunswick Appalachian transect: bedrock and Quaternary geology of the Mount Carleton – Restigouche River area
Reginald A. Wilson, Michael A. Parkhill, and Jeffrey I. Carroll

B9 Gold metallogeny in the Newfoundland Appalachians
Andrew Kerr, Richard J. Wardle, Sean J. O'Brien, David W. Evans, and Gerald C. Squires