HANSON

53RD ANNUAL REUNION

FRIENDS OF THE PLEISTOCENE

HALIFAX, NOVA SCOTIA

MAY 25-28, 1990

QUATERNARY GEOLOGY OF NOVA SCOTIA

GUIDEBOOK FOR FIELD EXCURSION

NOVA SCOTIA DEPARTMENT OF MINES AND ENERGY
OPEN FILE REPORT 90-008

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Department of Mines and Energy

GEOLOGICAL SURVEY OF CANADA

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ITINERARY

Day 1 - May 26, 1990

Leaders: Grantham, Finck, Mott, Scott, Stea

Time	Event
7:45 AM	Muster in lobby of Shirreff Hall; leave for first stop
8:45 - 9:45 AM	Stop 1-1
10:15 - 10:45 AM	Stop 1-2
11:30 - 1:00 PM	Stop 1-3
1:15 - 2:15 PM	Stop 1-4
3:00 - 4:00 PM	Stop 1-5
4:45 - 5:15 PM	Stop 1-6 (optional)
6:00 PM	Arrive in Truro, stay at the Nova Scotia Agricultural College

Day 2 - May 27, 1990

Leaders: Brewster, Davis, Finck, Mott, Stea

Time	Event
7:45 AM	Muster in lobby of Frazier House; leave for first stop
8:10 - 8:40 AM	Stop 2-1
9:10 - 10:00 AM	Stop 2-2
11:00 - 12:30 PM	Stop 2-3
1:15 - 2:15 PM	Stop 2-4
2:40 - 3:20 PM	Stop 2-5
3:30 - 4:00 PM	Stop 2-6
4:30 - 5:00 PM	Stop 2-7
5:20 PM	Arrive in Antigonish, stay at St. Francis Xavier University

Day 3 - May 28, 1990

Leaders: Mott, Stea

Time	Event
8:15 AM	Muster in lobby of Bishop Hall; leave for first stop
9:20 - 10:30 AM	Stop 3-1
10:40 - 1:00 PM	Stop 3-2
1:15 PM	Drive back to Halifax; end of field trip.

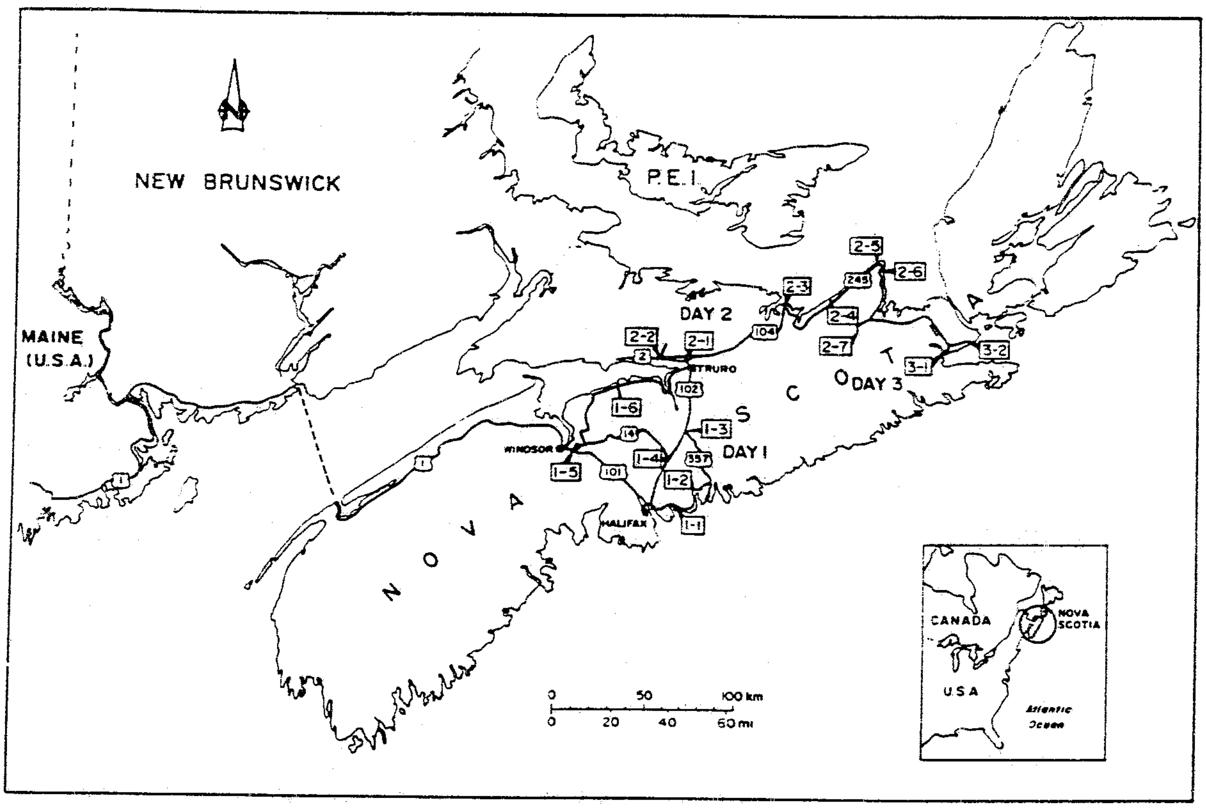

Figure 1. Route map for the Friends of the Pleistocene field trip.

SAFETY PROCEDURES

For personal and group safety reasons, field trip participants are advised to read and heed the following safety related procedures. The field trip leaders will endeavour to make the trip as safe as possible but can do so only with the co-operation of the participants.

1. PICKS AND HAMMERS. Do not indiscriminantly hammer and do not swing the hammer or pick wildly. Use downward blows and make sure no one is standing close to you. Do not pick at rock or till above your head.

2. SUITABLE CLOTHING. Participants should have adequate footwear and protection against both wet and cold, including hat and gloves. Adequate clothing is particularly important if there is an accident.

3. SAFETY GOGGLES. Although safety goggles are a sensible precaution, they are not widely used by professional geologists. Participants are advised that safety goggles are recommended. Safety goggles are particularly important when hammering fine grained, hard rocks.

4. HARD HATS. Except in mines and quarries, most professional geologists do not wear hard hats. However, hard hats do provide some protection against falling rocks and are recommended. They are strongly recommended if work is to be done on cliff sections or in quarries. Hard hats can be loaned to participants on request.

5. ROAD CUTS. Avoid crossing highways to examine road cuts. Examine only the side on which you are parked. Pull right off the road on 100 series highways in Nova Scotia Do not park on the hard shoulder and do not allow participants on the hard shoulder. On any highway, park in a safe manner and do not stray onto the road.

6. FALLING ROCKS. Falling rocks are a major hazard on field trips. Every situation should be individually assessed, but the following are useful guidelines:
   (a) avoid obviously unstable or overhanging cliffs,
   (b) do not hammer above your head or above others,
   (c) if a slope must be climbed, do not allow participants to climb while others are below, and
   (d) do not undercut unconsolidated cliffs that might slump.

7. CLIFF TOPS. Where possible, avoid cliff tops. Avoid potential overhanging cliff tops.

8. TIDAL SECTIONS. Take great care on sections that are cut off at high tide. If working in marginal tide conditions, only go with a person who has experience in the area. Do not work on a rising tide. A person will be delegated responsible to bring up the rear of the party and effectively watch for stragglers.

9. DO NOT CLIMB TILL CLIFFS unless it is essential to do so and you are experienced, then climb only if you have a friend present.

10. DO NOT CLIMB INTO TRENCHES, except where the trench is protected by a cage and supervised by the field trip leaders.

11. USE COMMON SENSE and when in doubt ask questions.

QUATERNARY GEOLOGY OF NOVA SCOTIA

GEOLOGICAL SETTING

Mainland Nova Scotia is characterized by geologically unique tectonic terranes in juxtaposition (Fig. 2). The Cobequid Fault system separates the Cobequid Terrane to the north from the Meguma Terrane to the south (Donohoe and Wallace, 1982). The Cobequid Terrane consists of a highland massif of Precambrian to Carboniferous volcanic, metasedimentary and igneous rocks flanked by basinal sedimentary rocks of Carboniferous and Triassic age. The Meguma Terrane consists largely of Cambro-Ordovician metasedimentary rocks intruded by Devono-Carboniferous granitoid rocks. These rocks form a large part of the Southern Uplands physiographic province (Goldthwait, 1924). The differing bedrock terranes provide useful indicator erratics for the reconstruction of ice flow events.

GEOMORPHIC EVOLUTION OF NOVA SCOTIA

A major landscape-forming event in Nova Scotia was the opening of the present day Atlantic Ocean in the Triassic. The Bay of Fundy is floored by Triassic rocks gently folded into a syncline (Fig. 2). The North Mountain marks the southern boundary of the Bay of Fundy. Cretaceous deposits have been found in lowlands adjacent to the Bay of Fundy (Stevenson, 1959). The evolution of Nova Scotia after the Triassic rifting episode has been debated over the last century. The arguments centre on two basic theories of landform development: the time-dependant, evolutionary concepts of the geomorphic cycle (Davis, 1922), and the steady-state landscape hypotheses (Hack, 1960). J. W. Goldthwait (1924) interpreted the geomorphic evolution of Nova Scotia in the light of

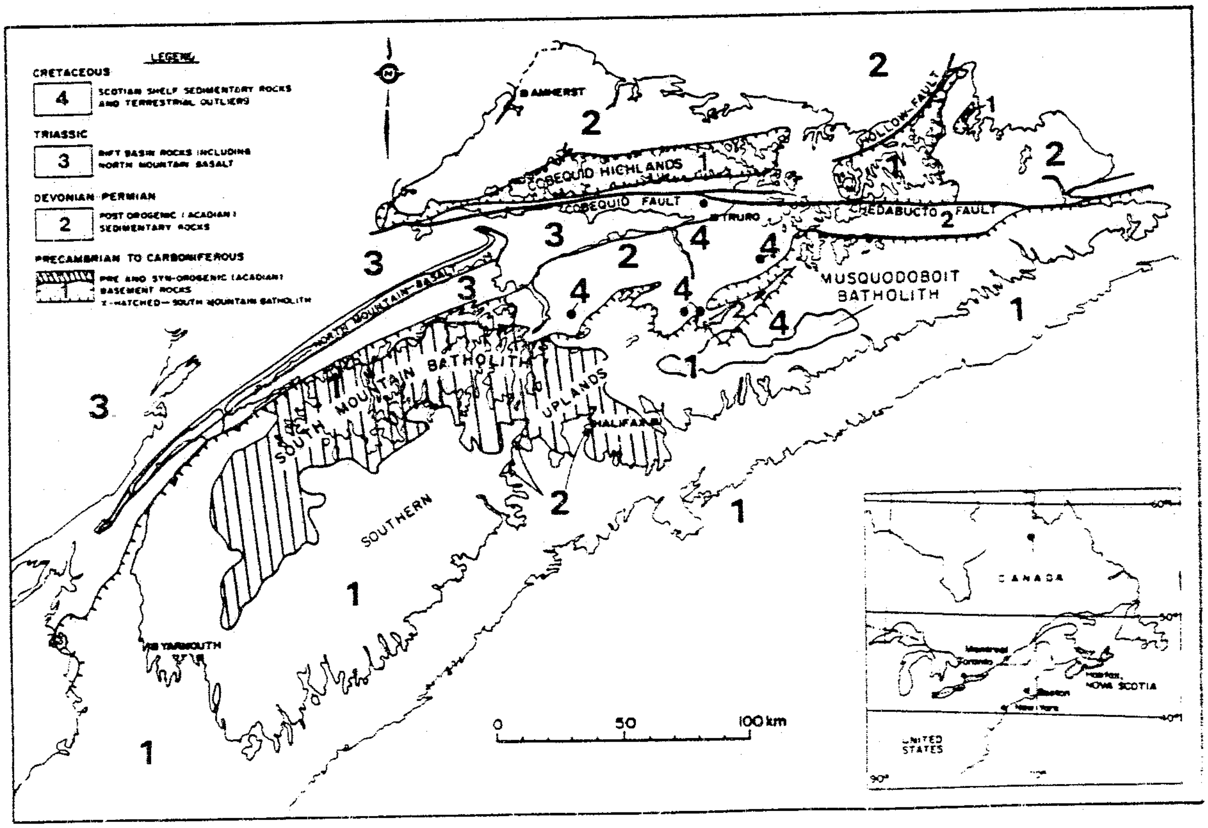

Figure 2. Generalized bedrock geology of mainland Nova Scotia.

Davisian concepts. He speculated that the entire region had been infilled after the Triassic and then planed off. Goldthwait believed that the age of this cycle predated deposition of kaolinitic clays and silica sands found in isolated sites in lowland regions of Nova Scotia. The age of these sediments was found to be Lower Cretaceous (Stevenson, 1959). Goldthwait suggested that the remnant of that peneplain is an erosional surface represented by the flat, accordant upland surfaces of the Caledonian Highlands in New Brunswick, and the North Mountain, Cobequid Highlands and Southern Uplands of Nova Scotia.

Southward-flowing rivers became superimposed on these older bedrock terranes. Evidence for the courses of these ancient rivers are wind gaps cut into the upland surfaces. These gaps, sometimes occupied by misfit streams, line up with modern rivers. Examples of this alignment are the Parrsboro and Folly Gaps with the Shubenacadie and Avon Rivers and the alignment of the St. John River and Digby Neck (Fig 3).

Uplift initiated a new cycle of erosion on the Cretaceous peneplain. The consequent southward-directed drainage patterns became pirated by subsequent streams developing in northeast-striking basins underlain by weaker Carboniferous and Triassic rocks.

Welsted (1971) argued that the trough of the Bay of Fundy had been in existence since the Triassic. There is little evidence for an extensive post-Triassic cover of rocks and superposition of streams. Welsted explained the flat uplands as exhumed surfaces of great antiquity. Rivers flowed down the synclinal limbs and down the axis of the Bay of Fundy. Wind gaps were created by tributary streams flowing along fault zones. The North Mountain cuesta was formed by longitudinal streams cutting through the basalt into the underlying softer sediments while tributaries denuded the basalt slopes.

King (1972) suggested that the lowlands of the Bay of Fundy may have been covered by extensive Cretaceous and Tertiary sediments and subsequently exhumed by uplift, although he questioned the existence of a peneplain. The presence of widespread unconformities in the offshore sequence implies erosion or rejuvenation of the landscape. Hacquebard (1984) calculated that the depth of burial of lignites in the Early Cretaceous outliers was between 700 and 1000 m. At this depth of burial Early Cretaceous or younger sediments would easily have overstepped the present-day Southern Uplands as well as the Cape Breton Highlands, which attain heights of 600 m. Apatite fission track data suggest an even greater post-Triassic cover, as much as 2 km (R. J. Ryan, pers. comm., 1990). Tertiary sediments have been found 40 km southwest of Yarmouth on the Meguma Group rocks of the Atlantic Uplands and are found in numerous outliers in the Gulf of Maine (King, 1972; G. B. Fader, pers. comm., 1986). Implicit in Welsted's (1971) theory of landscape evolution is the assumption that the present topographical relationships have been maintained; therefore, a kilometre or more of upland erosion is required to accommodate burial of the Cretaceous outliers. Extensive erosion of the Southern Uplands in post-Carboniferous time is unlikely given the preservation of thin skins of Carboniferous reef carbonates on the upland surface (Giles, 1981).

Recent work on the offshore Cretaceous sequence suggests that correlative beds to the onshore sediments were part of a huge drainage basin deriving material from the Canadian shield (Grist *et al.*, 1990). This is difficult to reconcile with a steady-state model where similar drainage patterns to today have been maintained since the Triassic.

DEVELOPMENT OF NOVA SCOTIA GLACIATION CONCEPTS

As early as the late 1800s, when the glacial theory was born, a controversy emerged about the nature of glaciation in Nova Scotia. Was the ice local, originating in upland areas and confined to the land masses, or was the ice part of a great continental mass that crossed the Bay of Fundy? The Reverend D. Honeyman, curator of the Provincial Museum in the late 1800s, discovered amygdaloidal basalt boulders along the Atlantic coast near Halifax (Honeyman, 1876). They had been transported a distance of 130 km. He used the observation to support the concept of a continental-based ice movement that crossed the Bay of Fundy.

Robert Chalmers (1895) of the Geological Survey of Canada mapped surficial deposits and glacial features in Eastern Canada. He carefully mapped glacial grooves and striations in Nova Scotia and interpreted a sequence of local ice movements. He proposed that northern Nova Scotia had been glaciated largely by local glaciers with floating ice a secondary agent in low-lying areas. In contrast to Honeyman, he did not believe that a glacier had crossed the Bay of Fundy. Chalmers (1895, p. 95m) stated:

"The depression of the Bay of Fundy was not

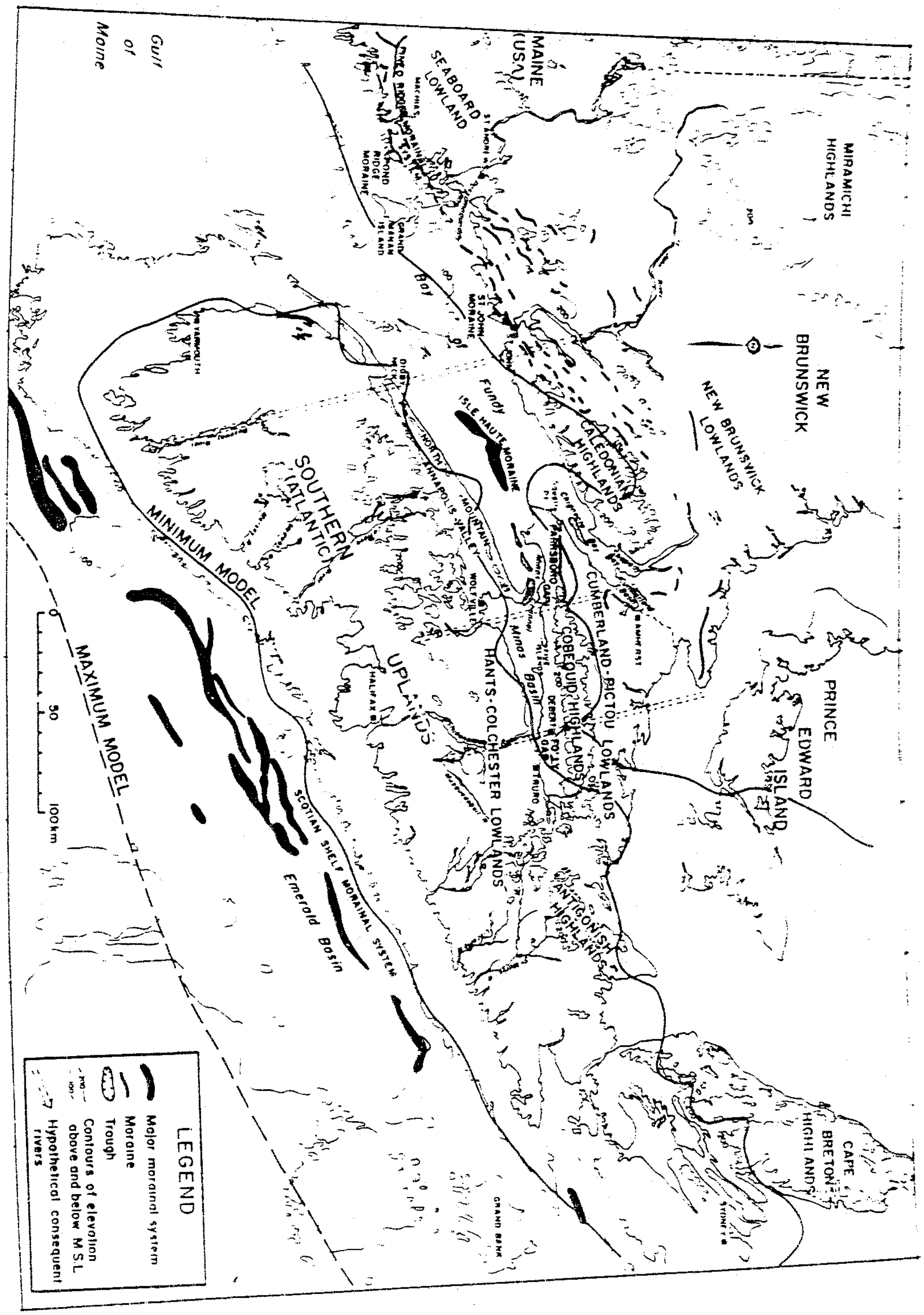

Figure 3. Physiography of the Maritime Provinces, with major morainal systems and former ice configuration models.

crossed by land ice from southern New Brunswick . . . Neither has Nova Scotia been glaciated by extra-peninsular ice from the north or northeast".

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

> "As in other parts of southwestern 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 centre of the movement, the latter now occurring in all directions".

J. W. Goldthwait (1924), in his treatise "The Physiography of Nova Scotia", dismissed all evidence regarding local glaciers in Nova Scotia. He envisioned a major ice mass moving southeastward, stemming from a Labrador source, and a subsequent southward-directed ice movement called the Acadian Bay Lobe, stemming from Laurentide ice in the Gulf of St. Lawrence. Goldthwait's (1924) ideas gained ascendancy in the following years, and much of the earlier work was discredited.

Pleistocene mapping in the Annapolis Valley initiated at Acadia University, Wolfville (MacNeill, 1951; Purdy, 1951; Swayne, 1952), revived the concepts of continental, then local glaciations (MacNeill and Purdy, 1951). Synchronously, Flint (1951) advanced the idea of highland centres of outflow based in part on the previous work of Chalmers (1895). Hickox (1962) confirmed that granitic erratics on the North Mountain were derived from the South Mountain Batholith in Nova Scotia and not from New Brunswick as Goldthwait (1924) had suggested.

During this period the radiocarbon dating method became established and processes of glacier mechanics were being elucidated. The timing of glaciations in Nova Scotia began to emerge as a subject of contention. Regional analyses of air photographs prompted Prest and Grant (1969) to postulate that there were several local centres of ice flow remnant from continental Laurentide ice that crossed the Bay of Fundy. They proposed that this local ice buildup was not due to climatic changes, but to drawdown of ice caused by incursion of the sea into the Bay of Fundy and Gulf of St. Lawrence.

The debate then focused on the timing of the Laurentide ice flood. King (1969) concluded that the Scotian Shelf morainal system represents the terminus of the Late Wisconsinan continental ice advance. Grant (1975) proposed a southward flood of ice prior to 39,000 years before present (yr B.P.) followed by a retreat about 38,000 years ago and re-expansion of Nova Scotia glaciers during the Late Wisconsinan. Grant (1977) further developed the hypothesis that Late Wisconsinan ice of Labradorean (Laurentide) origin never crossed the Bay of Fundy. This became known as the minimum model (Fig. 3). Grant (1977, p. 247) stated:

> "Evidence from scattered stratigraphic sections, from the relationship of a sequence of ice flow indicators to a raised interglacial marine platform, together with the limits of freshly glaciated terrain against weathered bedrock areas, indicates that late Wisconsinan glaciers spread weakly toward, and in many areas not beyond, the present coast. These were fed by a complex of small ice caps located on broad lowlands and uplands. The limiting factor was the deep submarine channels that transect the region. Thus, Laurentide ice was limited to the northern Gulf of St. Lawrence".

In the maximum model proposed by earlier workers the Late Wisconsinan Laurentide ice sheet filled the Bay of Fundy and overran Nova Scotia, terminating offshore (Flint, 1971) (Fig. 3). In a variation of the maximum model, Denton and Hughes (1981) proposed that the Bay of Fundy was occupied by an ice stream that merged with a stream flowing into the Gulf of Maine. Recent work in the offshore suggests that a major ice flood extended to the edge of the shelf (King and Fader, 1986). Accelerator radiocarbon dates on shells in deglacial sediments give a Late Wisconsinan age for this ice advance (Gipp and Piper, 1989).

Since 1977 the Nova Scotia Department of Mines and Energy has conducted regional surficial mapping and till geochemistry programs. Much of the data in this field guide results from these mapping programs, funded in part by regional development agreements between the governments of Nova Scotia and Canada. These programs also involved the systematic stratigraphic and lithological analyses of till sections. The mapping and till provenance data from over 3000 samples have enabled us to construct a picture of ice flow events that is different from either of the two models favoured since the 1800s.

STRATIGRAPHY AND CHRONOLOGY OF QUATERNARY EVENTS

In Nova Scotia, a chronology has developed through the correlation of glacial deposits in stratigraphic section with mapped erosional and depositional landforms (Stea, 1984; Stea *et al.*, 1985, 1987). Difficulties with this approach arise in correlation across the varied bedrock terranes and physiographic regions of Nova Scotia (Fig. 2). A till deposited during one glacial flow will show facies changes across bedrock contact zones, as well as depositional variations due to ice dynamics. Striation trends associated with one flow may display variations due to topographic deflection. Several flows of differing ages may have the same trends.

These problems are overcome by the integration of detailed mapping on the surface with facies and provenance analysis of till sheets in stratigraphic section. Till sheets are linked to mapped ice flows through erratic dispersal studies, fabric, and their relationships to underlying striated bedrock and boulder pavements.

EROSIONAL STRATIGRAPHY

An ice flow sequence has emerged in Nova Scotia through the careful mapping of striations. Ice flow trends can be traced over broad regions of the Province using striations (Stea and Finck, 1984). In addition, crosscutting relations and the preservation of older, weathered striations on lee-side surfaces have allowed for the development of an erosional stratigraphy. The sequence of events defined by discrete, regionally mappable trends of striations are termed ice flow phases. The patterns of ice flow mapped by striations are verified by the orientation of glacier landforms such as eskers and drumlins, till fabric and dispersal studies (Stea, 1984). The sequence of ice flow phases has been discerned from superimposed striation sites, and through correlation with stacked till sheets (Stea, 1984). Each of the ice flow phases produced at least one recognizable till sheet, with lodgment and melt-out facies.

Ice Flow Phases

Phase 1

Striation patterns, distinctive erratics, till fabric, and striated boulder pavements suggest that the earliest and most extensive ice flow in Nova Scotia was eastward then southeastward (Fig. 4). Several widely spaced striation sites reveal a distinct eastward flow, preserved in lee-side hollows and depressions, later overrun by southeastward-trending striations. In fact, the eastward ice flow may represent a separate, older phase of glaciation. Erratic trains of igneous rocks from the Cobequid Highlands and trains of basaltic rocks from the North Mountain are oriented southeastward and can be traced to the Atlantic Coast, up to 120 km down-ice (Fig. 5; Grant, 1963; Nielsen, 1976). This phase may represent a Laurentide ice flow. Evidence of its passage across New Brunswick, however, is equivocal (Rampton *et al.*, 1984). Anorthosite boulders in western Prince Edward Island, of a presumed Canadian Shield source, suggest that Laurentide ice did cross the region at some time (Prest and Nielsen, 1987).

Phase 2

The second major ice flow was southward and southwestward from the Escuminac Ice Centre in the Prince Edward Island region (Rampton *et al.*, 1984; Fig. 4). This flow phase is analogous to the Acadian Bay Lobe of Goldthwait (1924) and the "Fundian" glacier of Shepard (1930). Goldthwait envisioned southward flow from a Laurentide source across the Gulf of St. Lawrence. Ice flow trends in Prince Edward Island (Prest, 1973) and adjacent New Brunswick (Rampton *et al.*, 1984) do not reflect a pervasive southward flow, but suggest radial flow from a local centre. This event is recorded by southward-trending striae crossing earlier southeastward-trending striae at many localities on the upland regions of Nova Scotia and New Brunswick. Material from the vast area of redbeds in northern mainland Nova Scotia and Carboniferous basins in the Prince Edward Island region was transported southward onto the metamorphic and igneous Cobequid and Meguma Terranes of mainland Nova Scotia. Southward dispersal of distinctive Cobequid Highland erratics occurred with the dispersal of the red material (Grant, 1963). Evidence of southward dispersal of red clastic material from the Bay of Fundy has also been noted in sediment cores in the Gulf of Maine (Schnikter, 1987)

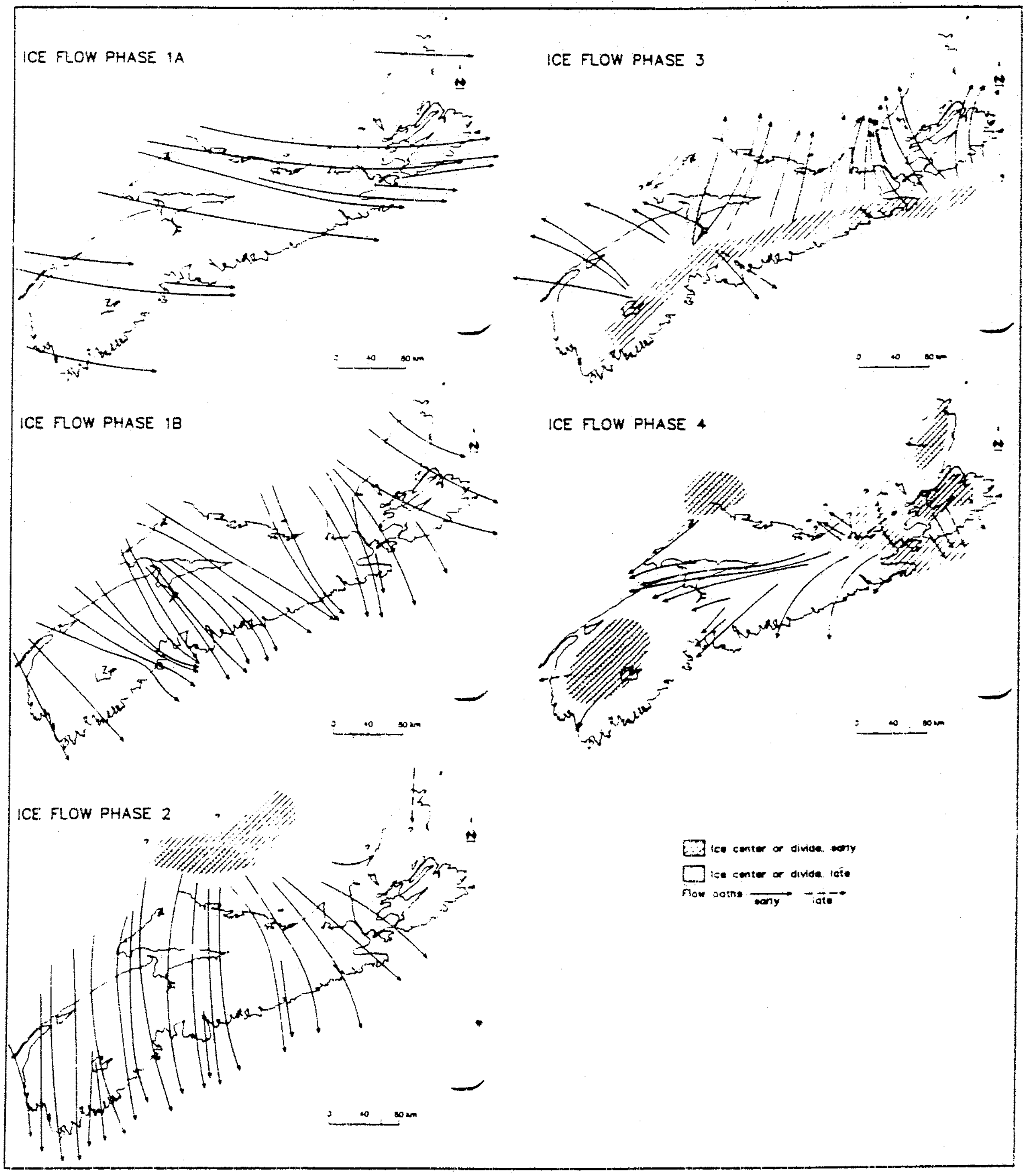

Figure 4. Ice flow phases and ice centres in Nova Scotia during the Wisconsinan Stage.

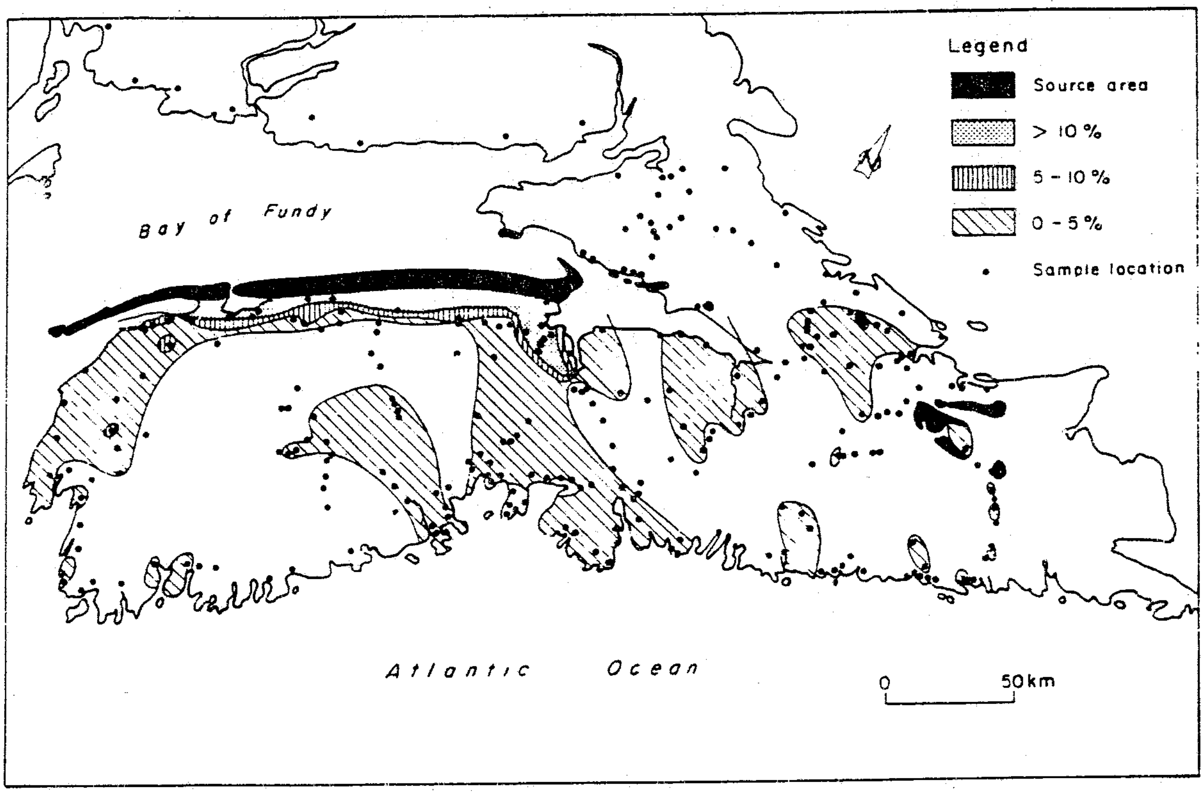

Figure 5. Distribution of basaltic erratics in Nova Scotia (after Nielsen, 1976).

Phase 3

During ice flow phase 3 granites from the South Mountain Batholith were transported northward onto the North Mountain basalt cuesta (Figs. 3, 4; Hickox, 1962). Erratics from the Cobequid Highlands can be found throughout the Carboniferous lowlands to the north (Fig. 6; Stea and Finck, 1984). Northward-trending striations can be traced across the northern mainland of Nova Scotia (Fig. 4). This well-documented northward ice flow occurred in response to the development of an ice divide in southern Nova Scotia (Fig. 4). Ice flow was northward and southward from this divide across the axis of the Nova Scotia peninsula. This divide may have formed as a result of marine incursion into the Bay of Fundy (Prest and Grant, 1969) or a climatic event (MacNeill and Purdy, 1951; Hickox, 1962). Northward ice flow from the Scotian Ice Divide was probably synchronous with the ice dome off Cape Breton Island proposed by Grant (1977).

Phase 4

During this final phase remnant ice caps developed from the Scotian Ice Divide (Fig. 4). Eskers and striations cut across features formed by earlier ice flows. Ice caps or glaciers that formed over the Chignecto Peninsula and southern Nova Scotia had margins on land marked by moraines, ablation till, glaciofluvial deposits, and the pinch-out of till sheets. Ice flow during this last phase was strongly funnelled westward into the marine basins. Erosional features and deposits relating to these late-glacial ice caps are restricted to low-lying areas. Figure 7 illustrates an area with striations relating to all four ice flow phases. In this region, appinites (distinctive mafic pegmatites) were transported up to 40 km southwest along flow lines formed during ice flow phase 4.

DEPOSITIONAL STRATIGRAPHY

Tills that relate to the four phases of ice flow

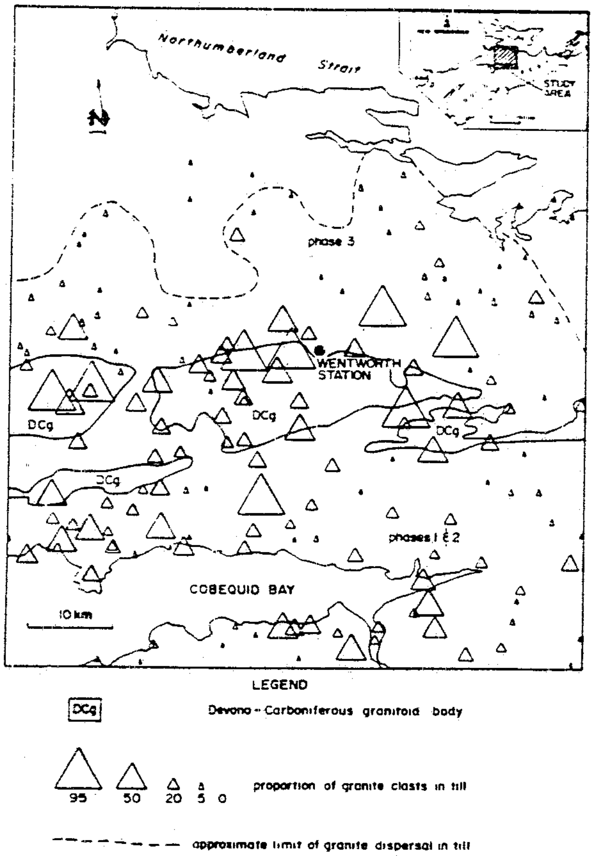

Figure 6. Distribution of granitic erratics in northern Nova Scotia.

not only overlie striated bedrock surfaces but at many locations overlie older, nonglacial and glacial deposits. In some cases the sediment package rests on a bedrock bench cut by a high stand of sea level.

The oldest sediments overlying bedrock are the Bridgewater and Mabou Conglomerates, deposits of deeply weathered and iron-cemented drift and outwash that have been assigned by various authors as being of Tertiary to early Pleistocene age (Prest *et al.*, 1972). A pre-Illinoian interglacial interval (200,000-300,000 yr B.P.) is indicated by amino acid racemization dates from a shelly diamicton in southwest Nova Scotia (Wehmiller *et al.*, 1988). Some organic deposits that have been assigned to the last interglaciation rest on tills (Addington Forks site is one that will be seen on this excursion) that likely relate to Illinoian glaciation (Mott and Grant, 1985).

Deposits assigned to the last interglacial interval are widespread throughout Nova Scotia (Fig. 3). This last interglacial interval, the Sangamonian Stage (75,000-128,000 yr B.P.), as defined by Fulton (1984), can be correlated with Stage 5 of the deep sea oxygen isotope record (Grant and King, 1984). During this interval, the climate fluctuated considerably. An early climatic optimum (substage 5e) was followed by less temperate cycles (substages 5d to 5a) that culminated in glaciation (Stages 4 to 2). Figure 9 is a summary of the stratigraphic and temporal relationships of Nova Scotia Quaternary deposits compared to the oxygen isotope record.

The early climatic optimum of substage 5e 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 that were depositional sites for nonglacial and glacial sediments.

Numerous organic deposits, none of which completely span this lengthy nonglacial interval, have been studied. In their recent synthesis Mott and Grant (1985) and Vernal *et al.* (1986) 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 Stea (1982) and Grant and King (1984).

These forest beds were deposited between the Illinoian and Wisconsinan glaciations. Radiocarbon dates seem to confirm a Sangamon age for most of the beds (Table 1). Any finite dates obtained have been suspect because the associated pollen spectra show the age to be untenable, or the material dated is considered unreliable (Fig. 10). Thorium/uranium disequilibrium method age determinations on wood from several deposits (Causse and Hillaire-Marcel, 1986; Fig. 11) have shed some light on the chronology. The dates seem to confirm a Sangamon age for most of the beds, but the youngest unit (Unit 3) produced some Mid-Wisconsinan ages. Vernal *et al.* (1986a) proposed an extended interglacial interval with fluctuating climate and no glacial interruptions until the Mid-Wisconsinan. The Mid-Wisconsinan U/Th dates, however, are considered minimum ages because of the strong possibility of post-depositional U migration (Mott and Grant, 1985).

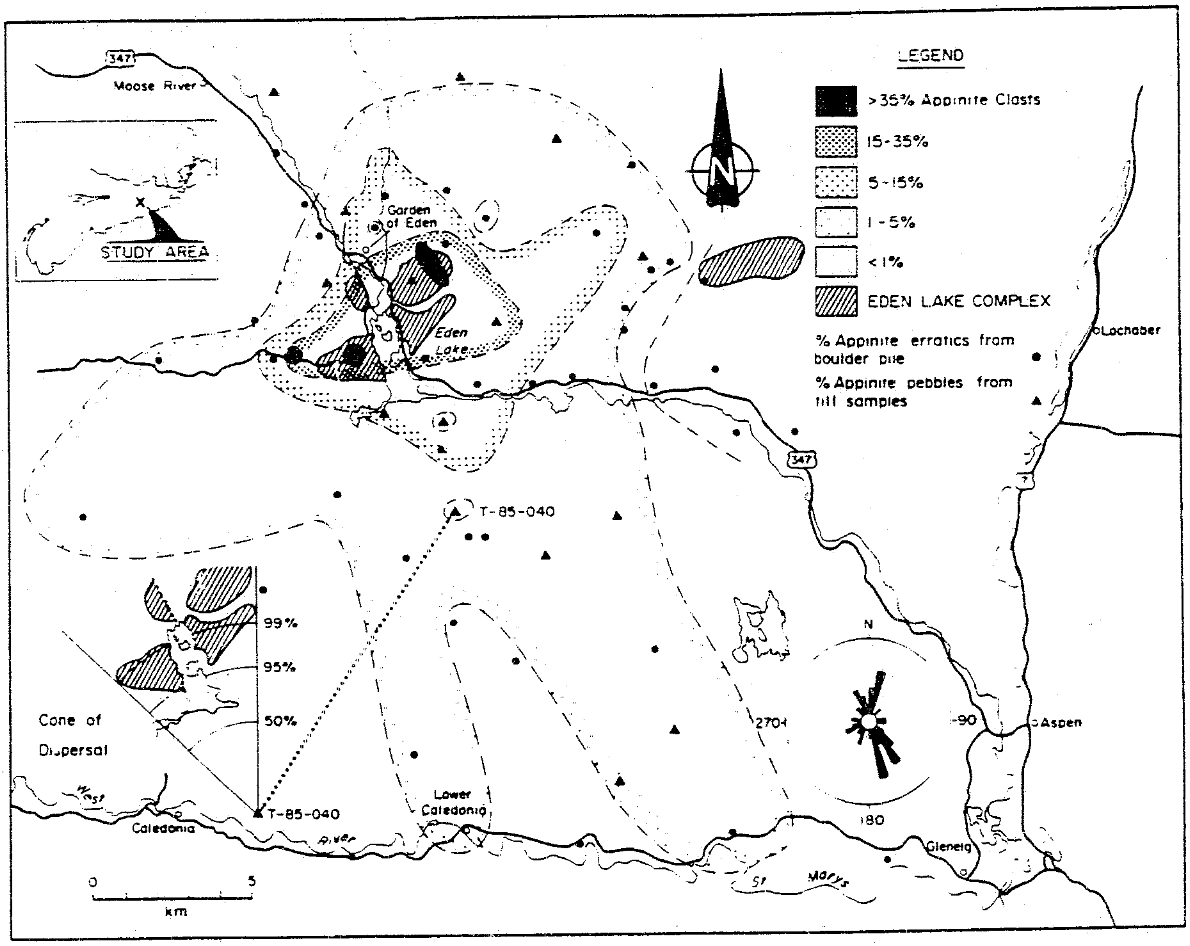

Figure 7. Dispersal of distinctive erratics from an appinite stock in northern Nova Scotia. A rose diagram of striation trends in the region is included on the diagram.

It is not certain when glaciers first developed in the Wisconsinan. In northern Cape Breton Island and possibly some other areas, ice free conditions may have persisted until 50,000 yr B.P.; offshore in the Gulf of St. Lawrence perhaps later. An alternative scenario is that glaciation began during oxygen isotope Stage 4 about 70,000 yr B.P. in most of Nova Scotia. The multiple tills relating to all four ice flow phases overlying the Sangamon deposits lack intervening nonglacial horizons implying continuous ice cover throughout most of the Wisconsinan Stage. At some sections, however, soil-like alteration and lacustrine deposits intervening between tills suggest a brief recession. On this trip we will visit some of the reference sections for the named Quaternary units in Figure 9.

DEGLACIATION

14,500 yr B.P.

The distribution of late-glacial features during ice flow phase 4 (Fig. 4) suggests that the last glaciers flowed radially from centres of remnant ice (Fig. 12). Earliest deglaciation occurred adjacent to the Bay of Fundy. Ice flowed southwestward out of Chignecto Bay and westward out of Minas Basin, probably as a response to marine incursion into the isostatically depressed Bay of Fundy. The timing and extent of this ice flow phase is uncertain. An ice margin is defined by a glaciomarine delta at Spencers Island (Wightman, 1980) dating from 14,300 - 12,600 yr B.P. It

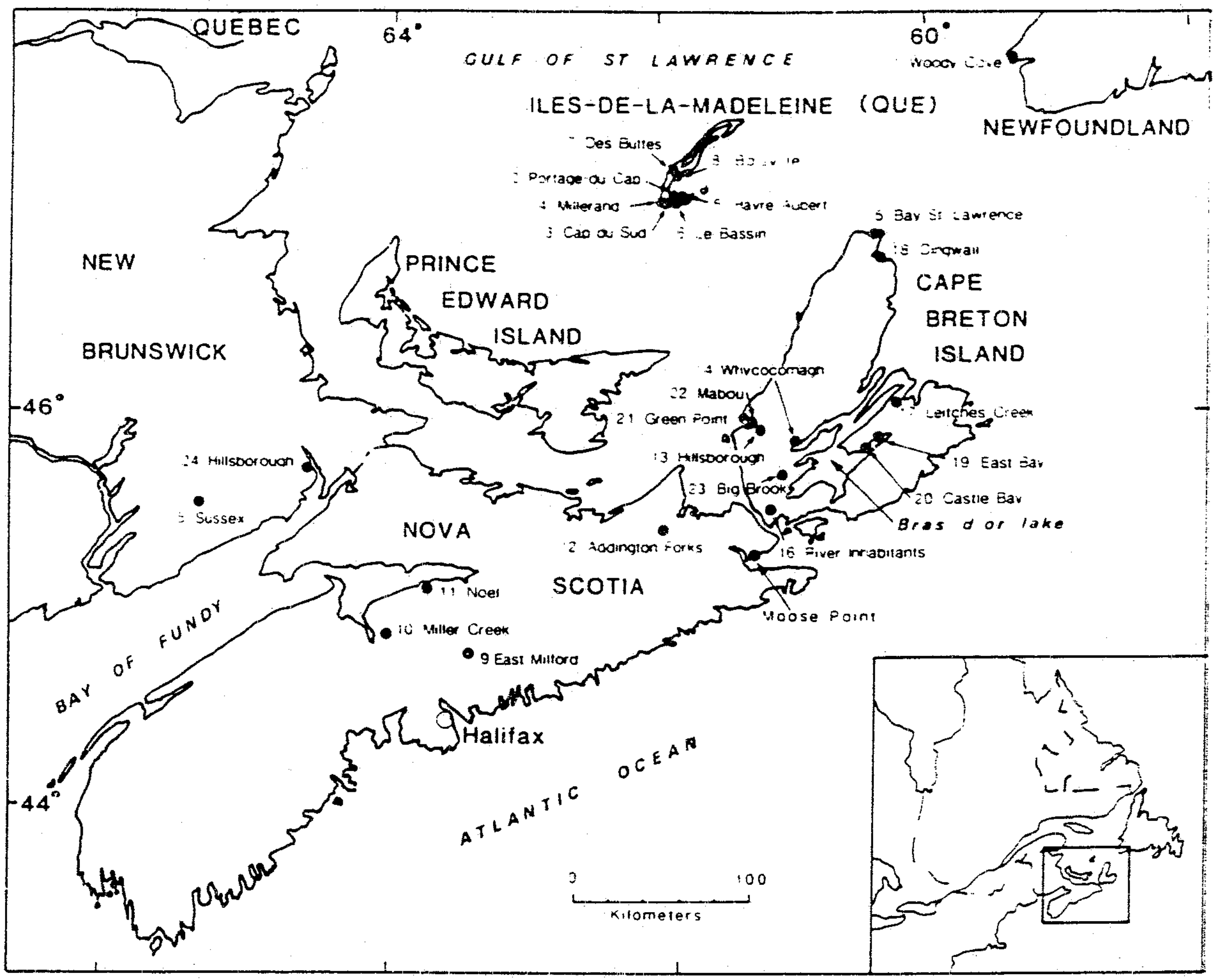

Figure 8. Location of buried pre-Wisconsinan organic beds in Maritime Canada.

represents the retreat of the southwestward-flowing glacier centred north of the mainland during ice flow phase 4. The Spencers Island dates, and similar dates on glaciomarine deltas in New Brunswick, suggest that the Bay of Fundy was ice free up to the Chignecto Peninsula by 14,000 yr B.P. (Fig. 13). A raised beach at +13 m in the Digby area near the Bay of Fundy in southwestern Nova Scotia indicates ice free conditions on land at this time (Grant, 1980). Ice extended off the Atlantic coast to an unknown distance. Sea level off the Atlantic coast at this time is believed to have been -110 m (King and Fader, 1986).

13,000 yr B.P.

Separation of local centres of outflow in northern Nova Scotia and in the South Mountain region may have occurred between 14,500 and 13,000 yr B.P. Retreat continued up the Minas Basin, and ice front retreat along the Atlantic coast may have reached the vicinity of the present coast. Ice may still have covered Northumberland Strait and Prince Edward Island. Basal radiocarbon dates from lakes in southwestern Nova Scotia suggest that by 13,000 yr B.P. the land areas adjacent to the Bay of Fundy were ice free.

12,000 yr B.P.

The impetus for ice removal continued to increase and glaciers retreated farther toward the local centres by 12,000 yr B.P. (Fig. 12). Chedabucto Bay area had already been ice free for several hundred years, and dates from sites in eastern Cape Breton Island suggest that deglaciation had occurred there as well (Mott *et al.*, 1986b). The

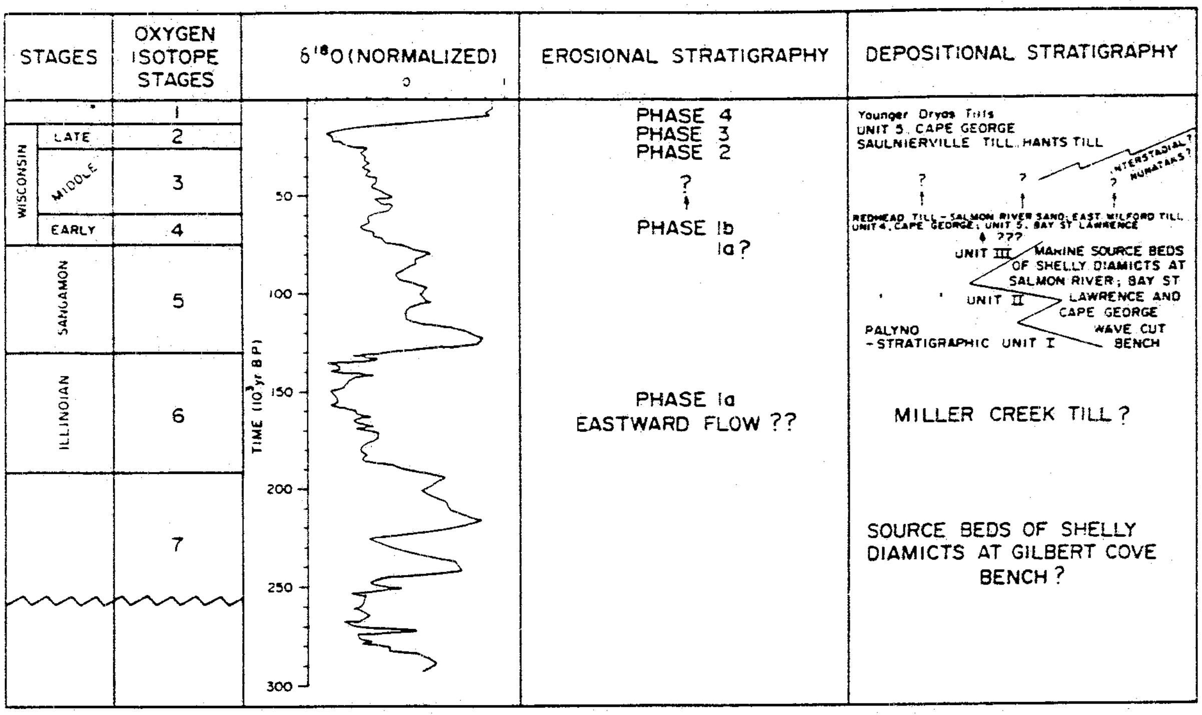

Figure 9. Stratigraphic summary for Nova Scotia Quaternary deposits and proposed chronology. Offshore chronology after Martinson et al., 1987.

Truro area at the head of Minas Basin was available for organic accumulation by this time. Ice must still have blocked the lower Shubenacadie River valley as thick, rhythmically-bedded clay deposits, some with dropstones, are found throughout the valley with ice proximal deposits nearest the Minas Basin. This lake probably formed sometime between 12,500 and 12,000 yr B.P. and lasted for a few hundred years with a southern outlet to the Atlantic Ocean (Fig. 14).

Glaciers must have remained in the Cobequid and Antigonish Highlands, most of western Cape Breton Island, possibly part of Prince Edward Island and on the South Mountain to the west. Lake cores from southern Nova Scotia (Fig. 12) reveal a mineral layer near the base of the cores which separates two periods of organic accumulation. The lower period of organic accumulation started about 13,000 to 12,000 yr B.P.

Low shrub and herbaceous vegetation occupied suitable sites on the landscape, usually low, wet areas that offered some protection from the weather or where soils were more suitable for plants to colonize. Trees had not yet migrated into the region.

11,000 yr B.P.

Deglaciation of Nova Scotia may have been virtually complete by 11,000 yr B.P. Small remnant or stagnant ice masses may have persisted in the Antigonish Highlands, South Mountain, the north side of the Cobequid Highlands and the Cape Breton Highlands. Some lake cores in northern Nova Scotia do not display the inorganic hiatus near the base of the core dated between 10,000 and 11,000 yr B.P. (Fig. 12). The minimum dates for deglaciation indicated by these lakes is 10,000 yr B.P. The absence of organic sites on Prince Edward Island dating from more than 10,000 yr B.P. may be due to a residual ice mass over part of the island and Northumberland Strait about this time.

It has been previously assumed that the decline in elevation of raised marine features from west to east in Prince Edward Island implies a greater ice load to the west and the absence of ice in the eastern region (Grant, 1977; Quinlan and Beaumont, 1982). This concept was also applied to the

Table 1. Radiocarbon dates on pre-Late Wisconsinan buried organic deposits in Atlantic Canada (see Fig. 18 for locations).

Site	Lab. No.	Age (Yr B.P.)	Material dated
1. Woody Cove	I-10203	>40,000	wood
2. Portage-du-Cap	BGS-259	>35,000	plant
	GSC-2313	>38,000	wood
3. Cap du Sud	GSC-3413	>38,000	peat
4. Millerand	GSC-3631	>46,000	wood
5. Havre-Aubert	GSC-3633	>47,000	wood
6. Le Bassin	GSC-3623	>46,000	wood
9. East Milford	GSC-33	>33,800	wood
	GSC-1642	>50,000	wood
10. Miller Creek	I-3237	33,000±2000	wood
	GSC-2694	>52,000	wood
12. Addington Forks	I-3236	33,700±2300/1800	wood
	GSC-1598	>42,000	wood
	GSC-3848	36,000±520	wood
13. Hillsboro	W-157	>40,000	wood
	GSC-370	>51,000	wood
14. Whycocomagh	GSC-290	>44,000	wood
15. Bay St. Lawrence	GSC-283	>38,300	wood
	GSC-3636	44,200±820	wood
	GSC-3864	>46,000	wood
16. River Inhabitants	GSC-1406-2	>49,000	wood
17. Leitches Creek	GSC-2678	>52,000	peat
18. Dingwall	GSC-3381	32,700±560	wood
	GSC-3417	>39,000	wood
	GSC-3541	>48,000	wood
19. East Bay	GSC-3861	>50,000	wood
	GSC-3871	>49,000	wood
	GSC-3878	>50,000	wood
20. Castle Bay	GSC-1577	>42,000	silt
	GSC-1619	>52,000	wood
21. Green Point	GSC-3220	>53,000	wood
22. Mabou	GSC-3317	>53,000	wood
23. Big Brook	GSC-3289	>49,000	wood
	GSC-3206	36,200±1280	silt
	GSC-3880	>52,000	wood
24. Hillsborough	GSC-1222	13,600±200	bone
	GSC-1310	37,300±1310	peat
	GSC-1680	>43,000	coprolite
25. Sussex	BGS-806	>35,000	wood

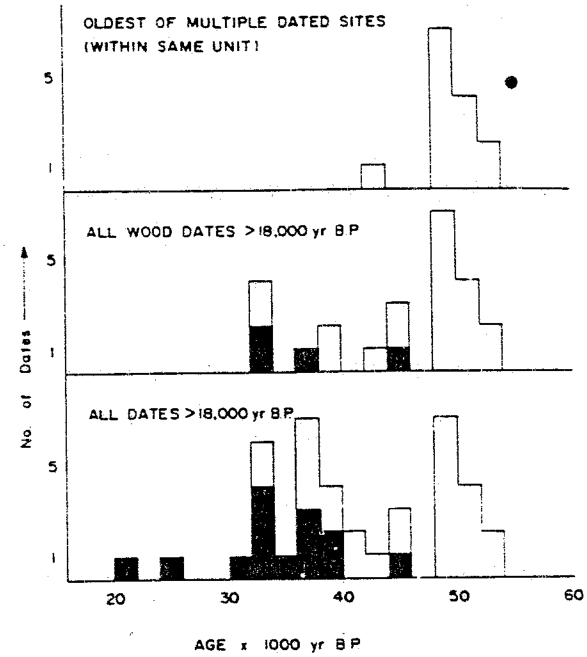

Figure 10. Histograms comparing all radiocarbon dates in Nova Scotia >18,000 yr B.P. with those on wood only, and sites where the unit with a finite date has been redated. Solid bar represents finite date.

Chignecto Bay area where the elevation of marine emergence features decreases northeastward and eastward up the Minas Basin and Cobequid Bay from 40 m to zero (Stea, 1983). Mapping results, however, show that ice flow was toward the areas of highest raised marine features (Stea *et al.*, 1985). Therefore, the regions showing no emergence were under ice for a longer period of time, either emergent under a carapace of residual ice or covered by a local re-advance (Stea, 1988).

Trees began to colonize the area after 11,500 yr B.P. Spruce, sometimes preceded by poplar/aspen trees in the form of open woodlands, occupied suitable sites south of Truro and westward along the Minas Basin coast. The tree line lay somewhere in northern Nova Scotia but had not reached Cape Breton Island.

Caribou and other large animals may have migrated to the Minas Basin area by this time through the emergent Bay of Fundy. Sea level may have been near its lowest point (Fig. 12). Remains of mastodon and mammoth dated around 11,000 yr B.P. have been uncovered from bank areas adjacent to the Bay of Fundy (Oldale *et al.*, 1988). Occupation of the Debert site

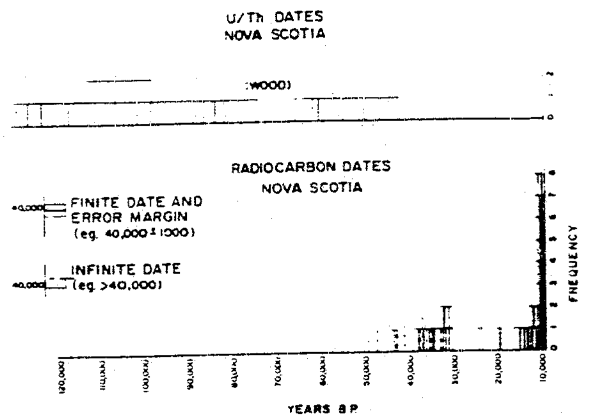

Figure 11. Histogram of U/Th and radiocarbon dates in Nova Scotia.

(Stop 2-2, Fig. 1) by paleo-Indian hunters may have begun by 11,000 yr B.P. if the oldest date is valid (MacDonald, 1968).

10,500 yr B.P.

Around 11,000 yr B.P. an abrupt and pronounced climatic deterioration occurred that strongly affected the landscape and its vegetation cover. Fluvial and lacustrine deposition buried organic sediments at many sites in the Maritime Provinces. The sedimentology of the units and evidence for base level lowering at this time (Scott in Stea *et al.*, 1987) strongly suggest that deposition was glacially controlled, either by stagnant ice (snowfields?) or active ice. Lake sites indicate mass wasting and that solifluction processes were active (Mott *et al.*, 1986a; Fig. 15). Diamictons overlying organics at many sites are believed to be of glacial origin.

Glacial and periglacial deposition of sediments overlying the organic sediments may be explained by re-advances of small remnant ice masses or build-up of ice caps that advanced into lowland regions (Fig. 16). These glaciers could have advanced and retreated relatively rapidly over till deposits in the lowland areas (Boulton and Jones, 1979). Any marine-based centres may have been extreme examples of low-profile ice masses riding over water-saturated marine mud. These glaciers built up in areas that today have the thickest snow cover in late winter (Fig. 17). There is a marked similarity between the distribution of residual snow and our postulated ice caps (Fig. 16). During most of the late-glacial period the region around Prince Edward Island would have been covered with ice or dry land. This would increase the likelihood of

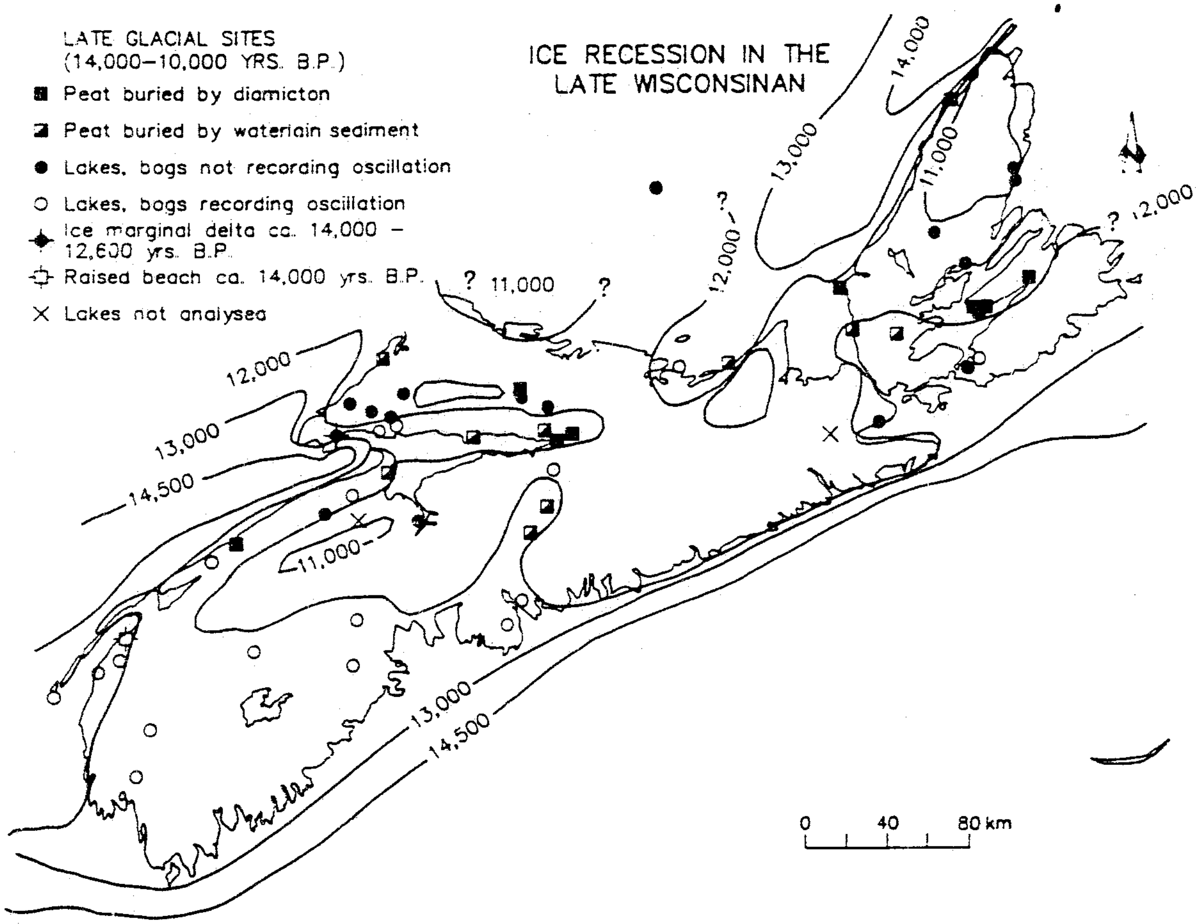

Figure 12. Deglaciation of Nova Scotia.

ice buildup in Prince Edward Island, which is today moderated by the ocean.

The extent of these glaciers in other areas is uncertain. G. B. Fader (pers. comm., 1988) suggests that ice reformed in the offshore shelf areas from 10,000-10,500 yr B.P. on the basis of a dated erosional unconformity overlain by till. Some of this ice may have been nurtured by flow from mainland ice caps such as the one over the Antigonish Highlands. The lack of ice marginal features in much of mainland Nova Scotia suggests that these ice masses either were largely stagnant or they were more extensive and terminated offshore in some areas.

Frozen-ground features in the deposits at Lantz and those described by Borns (1965) on outwash sand of the Five Islands Formation indicate that a periglacial environment existed in areas not covered by ice. Eolian sand has been found in some of these areas, including the Indian occupation site at Debert (Stop 2-2; Fig. 1). MacDonald (1968) suggested that the sudden abandonment of the site implied a climatic deterioration. Lake basins that were not completely inundated by mineral sediment show a period of increased mineral deposition suggestive of increased slope wash and solifluction processes that affected productivity in the lakes.

A profound change shortly after 11,000 yr B.P. is also recorded in the vegetation. Spruce trees were greatly affected and were decimated if not completely eliminated at some sites. Shrub birch communities reverted to willow dominance, and herbaceous taxa increased or dominated at some sites. Figure 15 is a summary of the late-glacial environmental changes in the Maritimes indicated by lake sites. This late-glacial climatic oscillation has been equated with the Allerod/Younger Dryas event of Europe and the north Atlantic (Mott *et al.*, 1986; Stea and Mott, 1989).

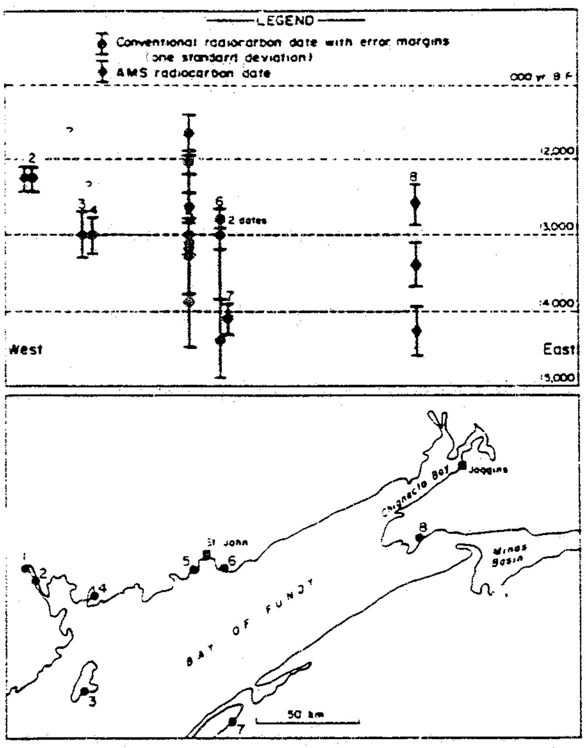

Figure 13. Radiocarbon dates on glaciomarine deltas along the Bay of Fundy. Dates compiled from Rampton et al., 1984; Nicks, 1989; Stea and Wightman, 1988.

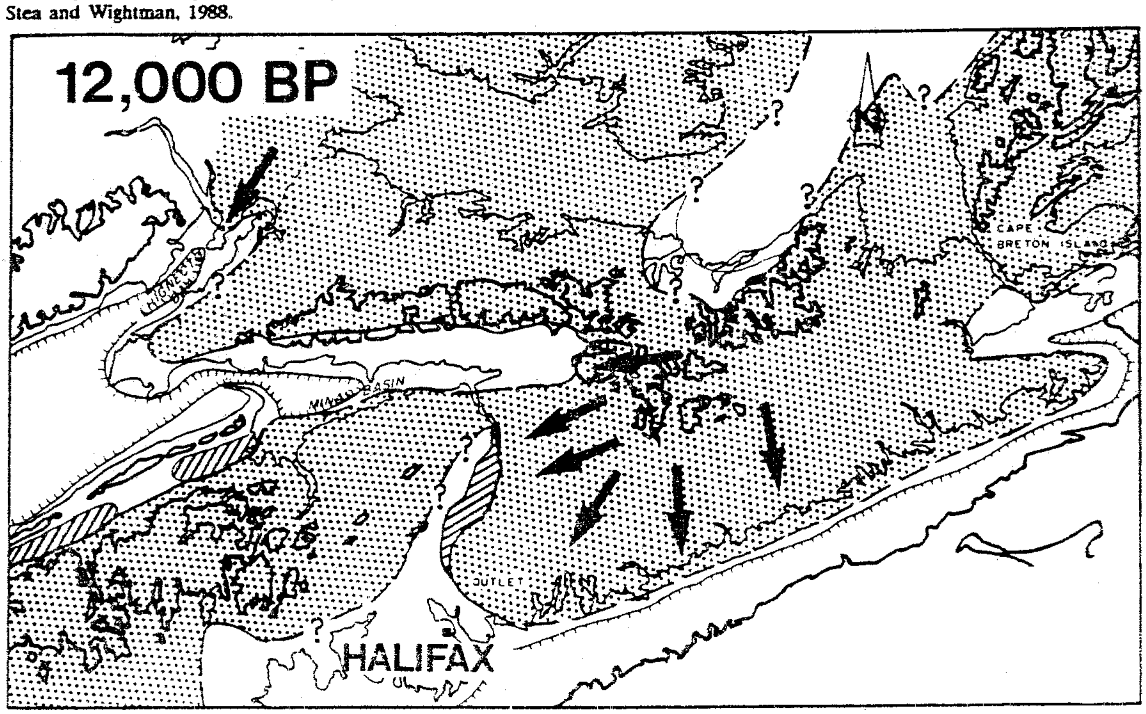

Figure 14. Ice configuration at 12,000 yr B.P. and formation of a glacial lake in the Shubenacadie Valley (diagonal lines).

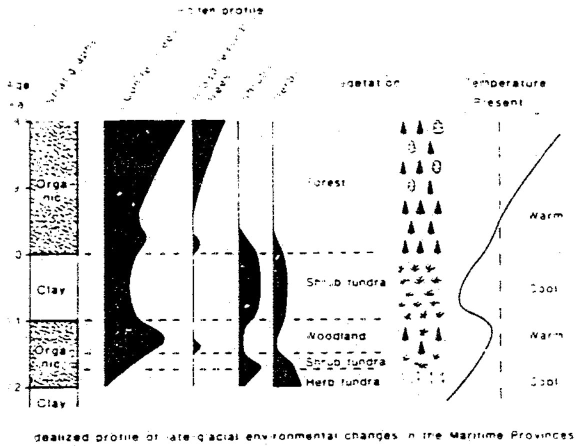

Figure 15. Typical stratigraphy and environmental significance of a lake site in Maritime Canada revealing the late-glacial climatic oscillation at its base.

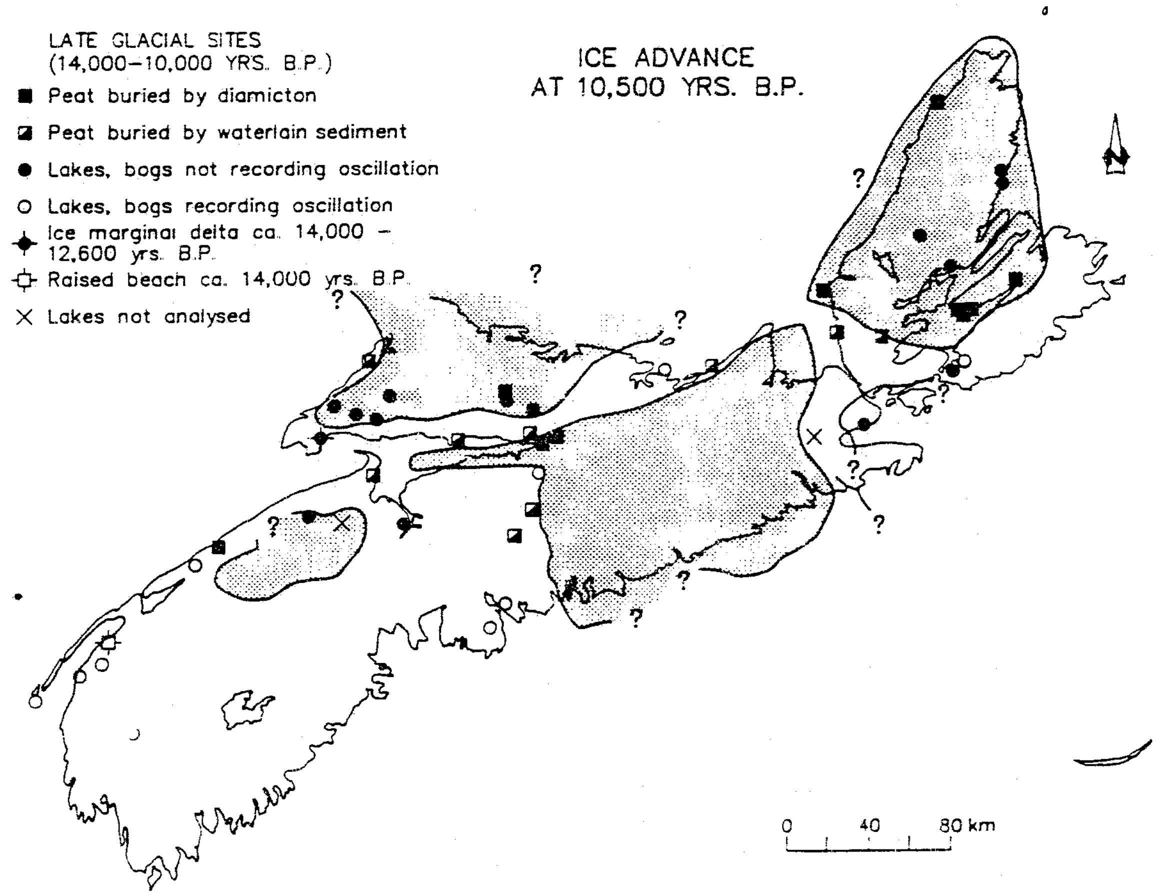

Figure 16. Ice re-advance at 10,500 yr B.P

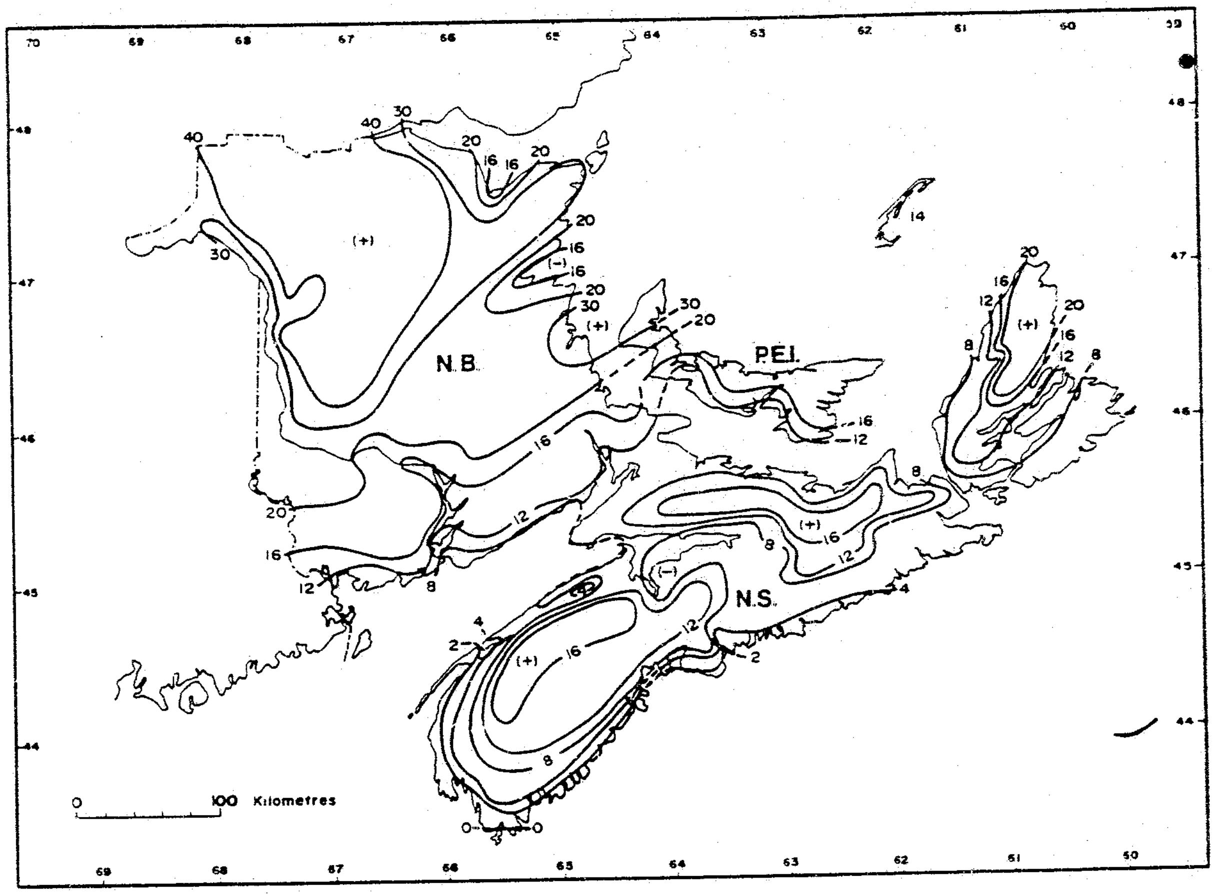

Fig ~e 17. Residual snow cover in late winter over the Maritime Provinces. The contours represent inches in excess of the median value for the region at that time of year.

DAY 1

STOP 1-1: WEST LAWRENCETOWN

Leader: R. R. Stea

Purpose: To examine a reference drumlin section with two distinct tills and a nearby striated outcrop. Discuss till genesis, drumlin formation and glaciation models for the area.

Route: Depart Shirreff Hall, Dalhousie University, Halifax, at 8:00 AM. Turn left and proceed on South Street to Robie Street. Turn left on Robie Street and follow to A. M. McKay Bridge. Cross to Dartmouth and proceed along Highway 111 through the MicMac rotary 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 until you spot a 'duck mailbox'. Turn in the driveway and park at the end of the road. Arrive 8:45 AM.

EN ROUTE TO STOP

Halifax is the capital of Nova Scotia. It was founded in 1749 by British Colonel Edward Cornwallis and 3000 settlers to act as a foil for the French fortress of Louisbourg in Cape Breton Island. As we travel to our first stop we will pass several areas of interest. From the A. M. McKay bridge several drumlins are visible in the harbour to the south, including Georges and McNabs Islands (Fig. 18). These islands formed part of the defense system of Halifax in the early 1800s. 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. 1600 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 5000 years ago (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 rotary we pass from this rough bedrock terrain onto a rolling terrain characterized by a series of southeastward-trending lakes and drumlins.

INTRODUCTION

The West Lawrencetown site (Fig. 18) is a wave-eroded part of a large southeast-trending drumlin. Bedrock outcrops on the flank of the drumlin. The outcrop reveals an older, parallel set of wide grooves and striations that trend 155°, cut by divergent, finer striations trending 180-190°.

STRATIGRAPHY

The southern, tapered end of the drumlin has been truncated by the sea. Exposures show two distinct diamicton units, the Hartlen and Lawrencetown Tills (Fig. 19). The lower, Hartlen Till is greyish, silty and very compact. It contains abundant metagreywacke boulders of the underlying Meguma Group and a few Devono-Carboniferous granitic clasts. These are derived from the Musquodoboit Batholith 30 km to the north and northeast (Fig. 20). Many bullet-shaped, stoss-lee boulders are embedded in the Hartlen Till. The trend of the long axes of these large boulders reflects the direction of ice flow that emplaced the till (Hicock and Dreimanis, 1985). The shape implies lodgment and subsequent moulding by overriding ice (Krüger, 1984; Clark and Hansel, 1989). Striations on the upper surface of the boulders parallel the long axis trends (Fig. 21). These are also parallel to the older set of grooves and striations on the bedrock surface and subparallel to the drumlin axis. A second set of striations, trending 180-190°, was also seen on these boulders, some of which were rotated. This is parallel to the second bedrock striation set. The Hartlen Till has a strong, unidirectional clast fabric with an eigenvector of 155°, parallel to the older set of striae in the rock core (Fig. 22).

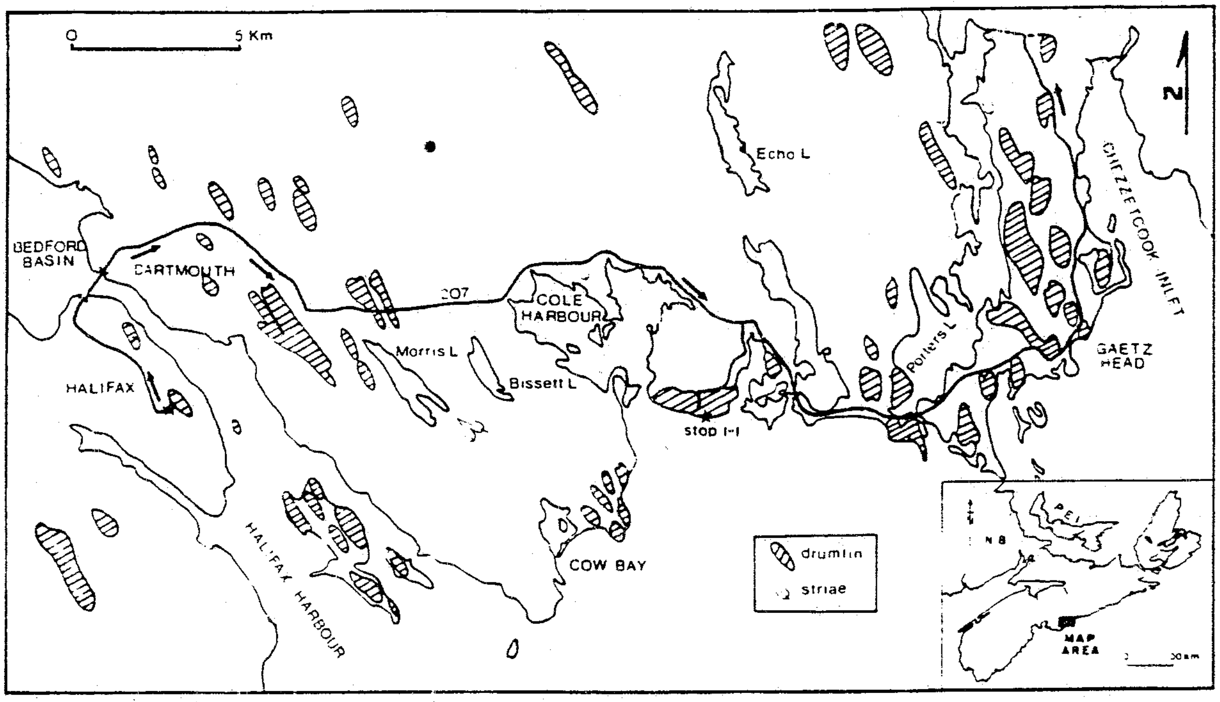

Figure 18. Drumlins and glacial features of the Halifax area and location of Stop 1-1.

A reddish, muddy diamicton unit (Lawrencetown Till) overlies the Hartlen Till with a knife-sharp contact. It is also silty and compact but is distinguished from the lower unit by its red colour and erratic content. Distinctive hornblende-bearing syenogranites, foliated granodiorites, and volcanic erratics from the Lawrencetown Till have source areas in the Cobequid Highlands, 80 km north of the section. The most probable range of flow paths from which the glacier would have sampled all these erratics is between 350° and 010° (H. V. Donohoe, Jr., pers. comm., 1988) (Fig. 20). Fine grained, micrographic syenogranite erratics and foliated syenogranites and granodiorites within the Lawrencetown Till are thought to have the most restricted ranges of the erratics. Their source areas (source areas C, D and G; Fig. 17) are the southeast part of the Hart Lake-Byers Lake pluton, due north of the section, and parts of the Salmon River Pluton adjacent to the Cobequid Fault (H. V. Donohoe, Jr., pers. comm., 1988). Using the ranges of these erratics the flow range is confined to 357-010°. The red colour of the Lawrencetown Till is likely derived from glacial crushing of red mudstones from areas north of the Cobequid Highlands or in the Minas Basin area. Pebble lithology of a bulk sample of the Lawrencetown Till shows that 37% of its clasts are from the underlying Meguma Group, indicating 0-10 km transport along the flow path designated in Figure 20. It also contains 12.2% Cobequid Highland lithologies, which represent 70-90 km transport distance. The anomalous red matrix of the Lawrencetown Till and the high erratic content suggest englacial deposition.

Stoss-lee boulders in the Lawrencetown Till (Fig. 21) show a more northerly trend than in the Hartlen Till. Striation trends on their surfaces are more divergent, with a peak parallel to the younger striae on the bedrock outcrop. Most boulders and surface striae are oriented parallel to the Hartlen Till fabric. Till fabric in the Lawrencetown Till shows a minor southeast peak and an eigenvector of 176° (Fig. 22).

The older striations and grooves on the striated outcrop are parallel to the fabric of the Hartlen Till and the younger striae to the fabric of the Lawrencetown Till. This suggests that the two units were formed by discrete ice flows, first southeastward then southward. The abundance of southeastward-trending boulders in the Lawrencetown Till is a paradox. Did the southward ice flow drape the Lawrencetown Till over the drumlin form without modifying its trend? Were the southeastward-trending boulders in the Lawrencetown Till inherited from the Hartlen Till in their original orientation? An englacial origin

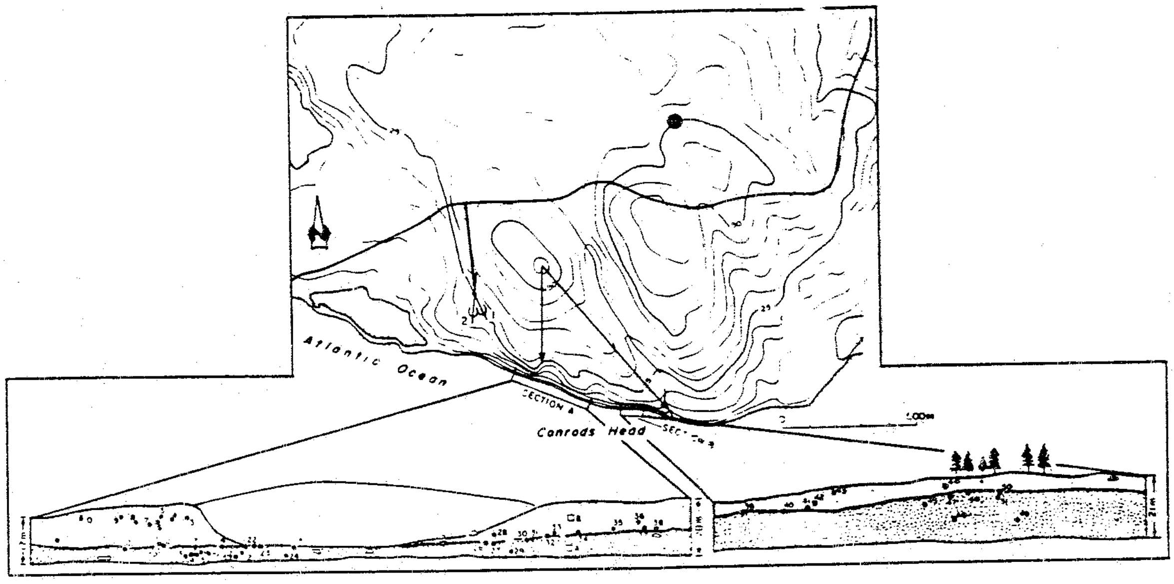

Figure 19. Map of the drumlins at West Lawrencetown and stratigraphic sections. Locations of measured boulders.

for the Lawrencetown Till would mitigate against this hypothesis. It is possible that the basal part of the Lawrencetown Till was formed by southeastward flow. The modal value of Lawrencetown Till boulder trends parallels the drumlin axis. More work on the section would help to resolve this problem.

The West Lawrencetown drumlin has well defined stoss and lee form. This stoss end faces northwest. The lee side gently slopes away towards 145° (arrow, Fig. 19). The Hartlen Till forms most of Section B (Fig. 19) at the lee end of this drumlin. Section A is located on a lobe-like extension of the drumlin oriented 193° (arrow, Fig. 19). In this section, the Lawrencetown Till forms much of the drumlins constructional topography. This suggests that the lobe-like extension of the drumlin, dominated by Lawrencetown Till, was accreted on the earlier tapered, southeast-trending drumlin formed of Hartlen Till.

Regional data suggest that the ice centres responsible for the flows changed abruptly, rather than shifted gradually. Two striation sets are commonly found on adjacent rock outcrops. This drumlin, as well as most other drumlins along the Atlantic coast, is characterized by two distinct tills. The drumlin axis can be parallel to the inferred flow direction for either the lower or upper till. In this case the main drumlin axis is 141°, subparallel to the earlier striation set on the rock core (155°) and lower till fabric. The secondary 'lobate' axis is 193°, parallel to the younger striation set and subparallel to the upper till fabric.

IMPLICATIONS FOR DRUMLIN GENESIS

A strong correspondence exists between striations and drumlin long axis trends in Nova Scotia. A plot of the frequency of striation trends versus drumlin trends illustrates this (Fig. 23). The variation in striation trends is largely due to successive ice flow events which inscribed many outcrops with sets of superimposed striations that can be correlated over broad areas of Nova Scotia. One type of drumlin in Nova Scotia drumlin fields is the 'lobate' or 'reoriented' drumlin which is a form with two principal axes parallel to ice movements established by striations and dispersal of distinctive erratics (Fig. 24). The west Lawrencetown drumlin is an example of a 'lobate' form. Some other examples of these lobate drumlins are seen in the Chezzetcook drumlin field (Fig. 25) which we pass through after Stop 1-1. The internal structure of Nova Scotia drumlins is generally a layer cake of till sheets. The most common situation is a grey 'core' till (Hartlen Till) formed by a southeastward ice flow across grey metasedimentary bedrock and a distinctly red 'mantle' till formed by phases of southward to southwestward ice flow (Lawrencetown Till). Drumlins in the Yarmouth field of southern Nova Scotia exhibit the greatest

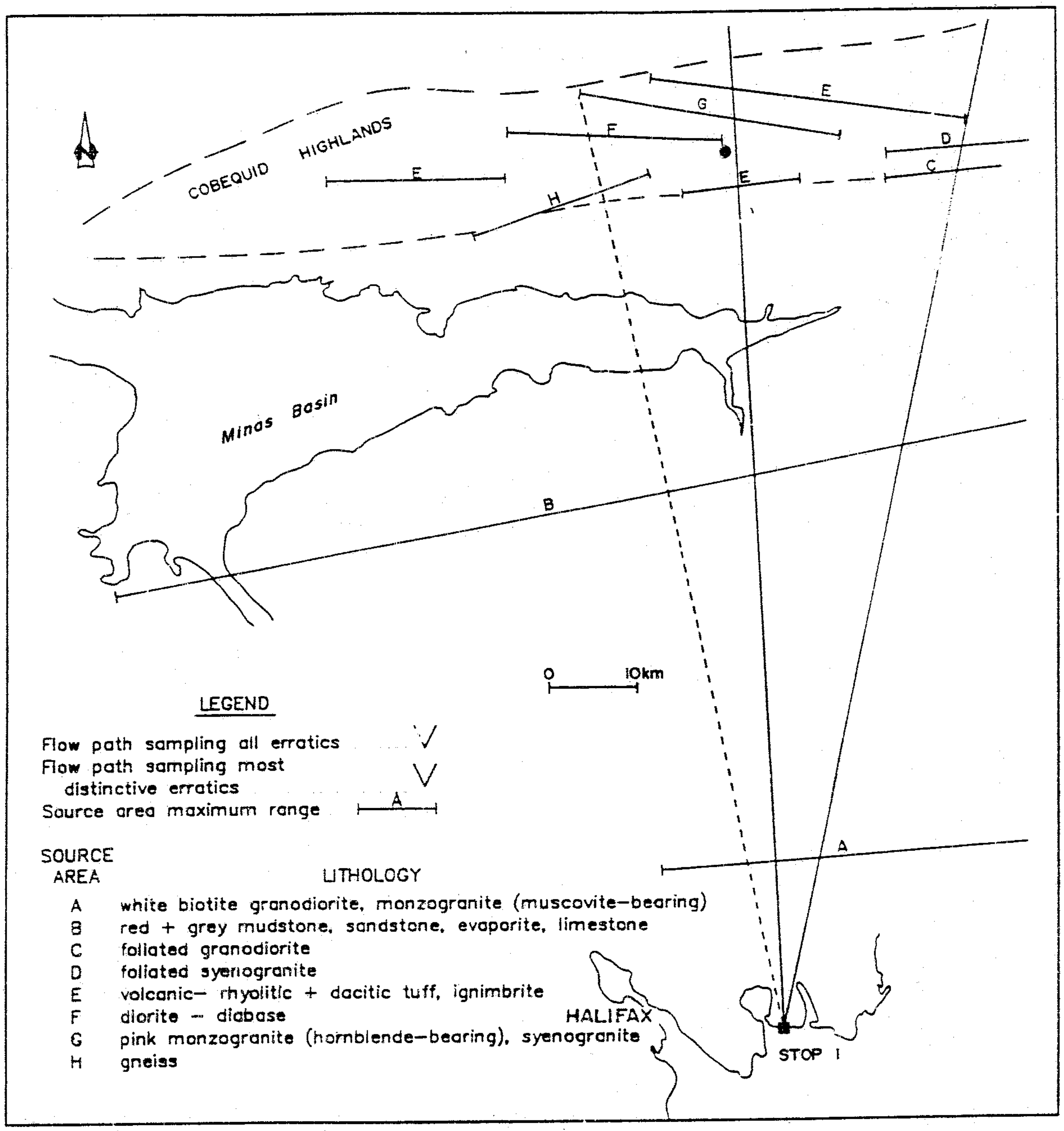

Figure 20. Source areas of boulders in the West Lawrencetown section.

number of stacked units capped by a course textured till that represents deposition by a late-stage ice mass centred inland. This late-stage ice mass also moulded some drumlins, with their stoss end indicating flow from southern Nova Scotia.

Nova Scotia drumlins appear to have formed during three major ice flow phases over the Province. The direction of the first and major drumlin-forming ice flow phase (phase 1) was southeastward. Nucleating agents for initiation of these drumlins may have been pre-existing tills or waterlain deposits. Individual and lobate drumlin axes imply a later, southward drumlin-forming ice flow (phase 2) parallel to the inferred ice flow trend for the mantle till units, based on clast provenance. Some of these drumlins nucleated from pre-existing southeastward-trending forms.

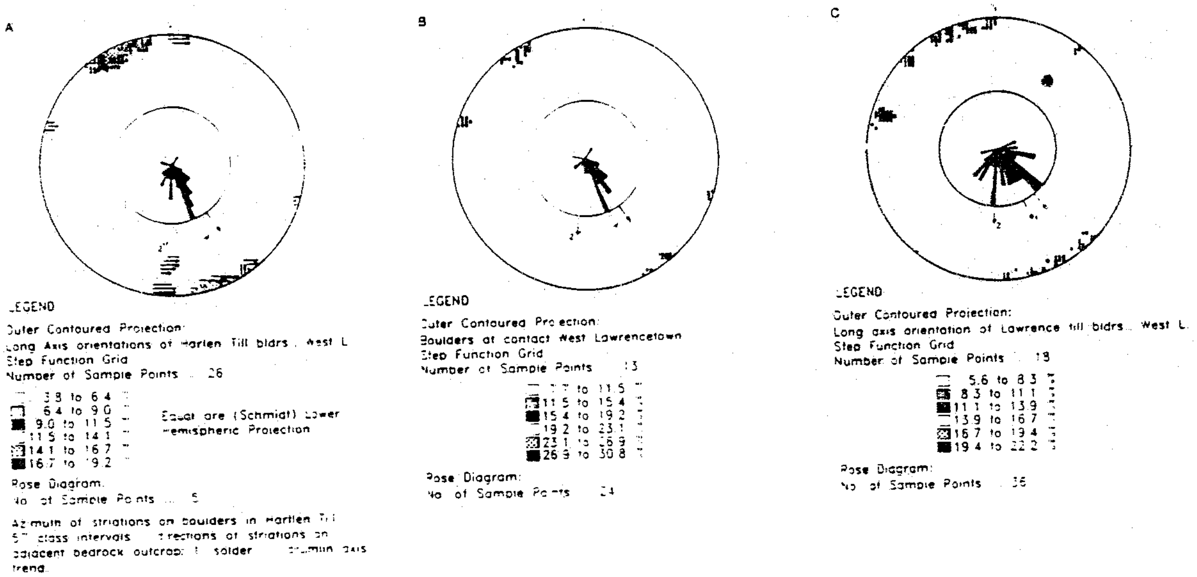

Figure 21. Lower hemispheric, equal area stereographic plot of the trend and plunge of the long axes of stoss-lee boulders at West Lawrencetown. Inner rose diagram of upper surface striation trends. (A) Hartlen Till (B) Boulders at the contact (C) Lawrencetown Till.

Flow from the local Nova Scotia ice caps (phase 4) also formed some drumlins (Fig. 24). These drumlins can have irregular shapes, but some are oriented with their stoss end facing inland.

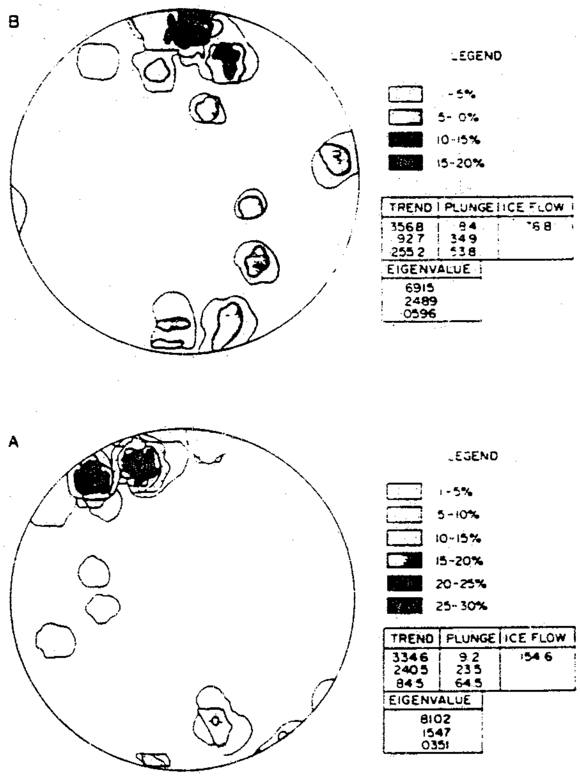

Figure 22. Lower hemispheric equal area projections of the plunge of the A-axes of elongate (3:1) stones in the (A) Hartlen and (B) Lawrencetown Tills. Contours represent percentages of 50 points in 1% area of projection.

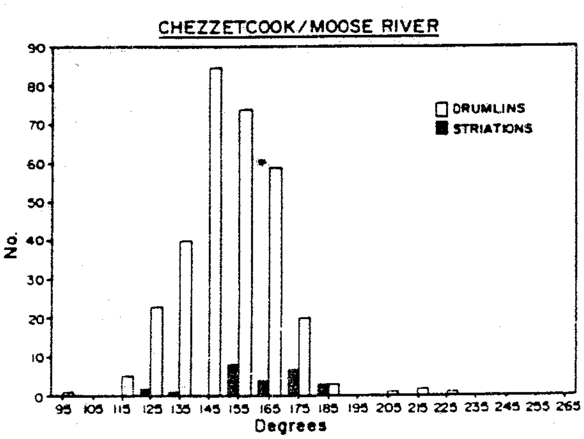

Figure 23. Histogram of striation and drumlin trends in the Chezzetcook drumlin field.

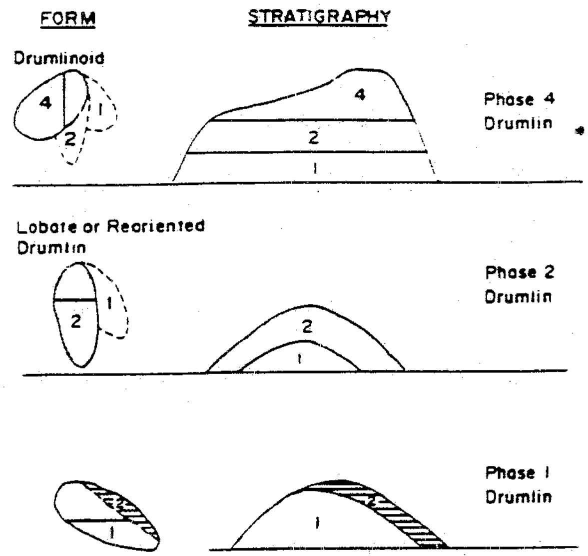

Figure 24. Relationship between form and stratigraphy in Nova Scotia drumlins. The dotted lines represent possible precursor shapes. Dashed lines represent passive deposition on the drumlin without form modification.

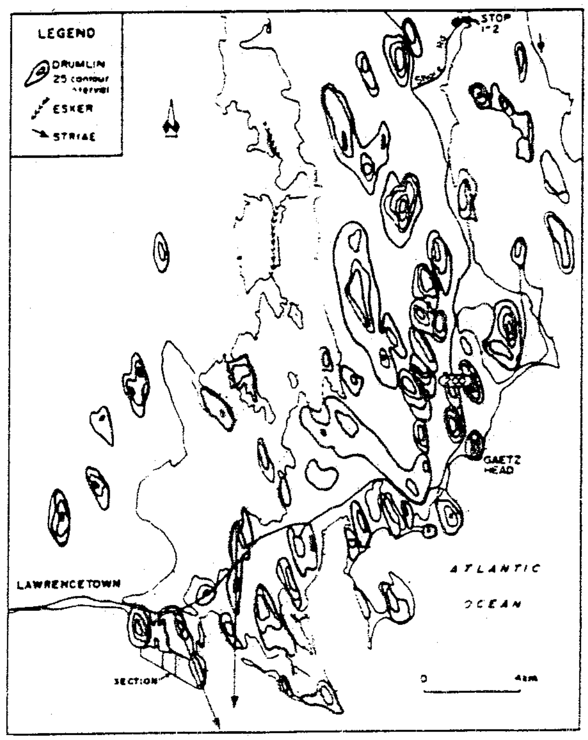

Figure 25. Chezzetcook drumlin field and location of Stop 1-2. Note lobate drumlins at the north end of the field.

STOP 1-2: CHEZZETCOOK

Leader: D. Scott and P. W. Finck

Purpose: To discuss sea level rise in the Maritimes and look at an outcrop of the youngest till unit in the region (phase 3).

Route: Leave Stop 1-1 at 9:45 AM. Proceed back to Highway 207. Follow it westward 14 km. Turn right along shore road past West Chezzetcook. Arrive 10:15 AM.

EN ROUTE TO STOP

We will drive through the Chezzetcook drumlin field (Fig. 25). On the route is an exposure of a beautifully symmetrical drumlin at Gaetz Head. This drumlin was the 'star' of a recent video called Ice Ages produced by the Atlantic Geoscience Society.

SITE DESCRIPTION

The Chezzetcook salt marsh is part of an estuary that formed as a result of sea level rise after 6600 yr B.P. 12 m of mud have aggraded since that time. Before that it was a freshwater bog. The submergence rate is 10 cm per century. The salt marsh is a recent phenomenon, a result of sedimentation due to human occupation. The few centimetres of sediment caused by land clearance upset the balance of the estuary, which was formerly a mud flat.

A road cut across from the salt marsh reveals a distinctive till unit which represents the most widespread surficial deposit on the Eastern Shore. It has been called local till, stony till, loose, sandy till, loose-textured till sheet, gravel-rich till, and granite, slate and metagreywacke till. All of these terms adequately describe this unusual till facies which largely reflects the nature of the underlying bedrock unit. Contrast this unit with the matrix-rich Hartlen Till seen at Stop 1-1 produced during ice flow phase 1 over the same bedrock units. Clearly, geology and the changing dynamics of the glaciers during the ice flow phases were major factors influencing the variability of local ground moraine. Graves and Finck (1988) estimated a renewal distance (distance down-ice in which 50% of clasts from the underlying bedrock are incorporated into the till; Peltoniemi, 1985) of tens to hundreds of metres compared to kilometres for the older till units. The autochthonous nature suggests formation near the base of a glacier; clast angularity, lack of striations, variable fabrics and lack of compactness suggest a melt-out origin (Dreimanis and Schluchter, 1985).

STOP 1-3: SHUBENACADIE WILDLIFE PARK

Leaders: R. J. Mott and R. R. Stea

Purpose: To visit a site where late-glacial peat is buried by sand believed to represent outwash from a late-stage glacier.

Route: Leave Stop 1-2 at 10:45 AM and proceed northward along shore road to Highway 7. Follow Highway 7 eastward 12 km to Musquodoboit Harbour. Turn left on Highway 357 to Meaghers Grant. Travel northward 26 km to Elderbank and turn left on the road to Cooks Book. Proceed 8 km. Turn right on Highway 224. Travel northward 14 km and turn right on Highway 2. The Wildlife Park is 1 km past the intersection on the right. Arrive 11:30 AM.

EN ROUTE TO STOP

Musquodoboit (pronounced Muskadawbut) is a Micmac Indian word meaning "rolling out in foam". Shubenacadie means "place where wild potatoes grow".

Near Wyse Corner the Musquodoboit River changes from an eastward-flowing, meandering stream to a straight, relatively rapidly-flowing river cutting through a narrow gorge through the Atlantic Uplands. Ice stagnated in this region forming ablation till and ice contact stratified drift. Lin (1970) suggested that this morainal dam caused the former tributaries of the north-flowing Shubenacadie River to reoccupy the former consequent valley by the Musquodoboit River.

We get a good view of a single southeastward-trending drumlin as we turn off to Cooks Brook.

INTRODUCTION

The Shubenacadie site is located in the Provincial Wildlife Park near the town of Shubenacadie (Fig. 26). The surrounding area is a gently rolling till plain which attains elevations of 66 m above mean sea level.

STRATIGRAPHY AND CHRONOLOGY

A report by Hennigar (1970) brought the existence of buried peat beds to our attention, and backhoe excavations were conducted to obtain material for analysis. The stratigraphic column from this site (Fig. 26) is a composite profile gleaned from borehole records and backhoe excavations.

The lowest unit in the boreholes is a plastic clay that extends to a depth of 25 m or more (Fig. 26). The till that generally covers bedrock in the area underlies the clay. Overlying the clay to the surface is a yellowish-grey, well-sorted, coarse to fine sand. Layers of a brownish, fibrous sedge-moss peat were found within the sand unit near the base at an average depth of 2 m. Two peat beds were encountered in some drillholes, one lying directly over the clay. Road exposures of the sand unit show horizontal beds of coarse and fine sand with normal grading of some beds.

The backhoe excavations encountered about 2 m of yellowish-grey sand overlying a 15 cm thick peat bed. About 25 cm of grey to grey-brown sand were seen beneath the peat bed overlying about 20 cm of reddish-brown clay and silty clay with minor pebbles. The excavation bottomed in sand with coarse pebbles and boulders. Thin organic seams and disseminated organics occur in the sand immediately above and below the peat bed. The excavations did not reach the lower clay encountered in boreholes.

A radiocarbon date of 10,800 ± 100 yr B.P. (GSC-3981) was obtained on the top 0.5 cm portion of the peat bed sampled by backhoe. A date obtained from a 1-2 cm thick basal increment of the peat was 11,400 ± 100 yr B.P. (GSC-4337).

PALYNOLOGY

Palynomorphs were recovered from the peat and from organic sand immediately below and above it (Fig. 27). Basal assemblages are characterized by large values for Cyperaceae (sedge) and abundant herb pollen and spores of *Lycopodium* (club moss), Pteridophyta (ferns), and *Selaginella selaginoides*. *Pinus* (pine) pollen may be overrepresented, due to its capability for long distance transport. *Picea* (spruce) values increase above the base of the peat and exceed 40%. *Betula* (birch) values also increase

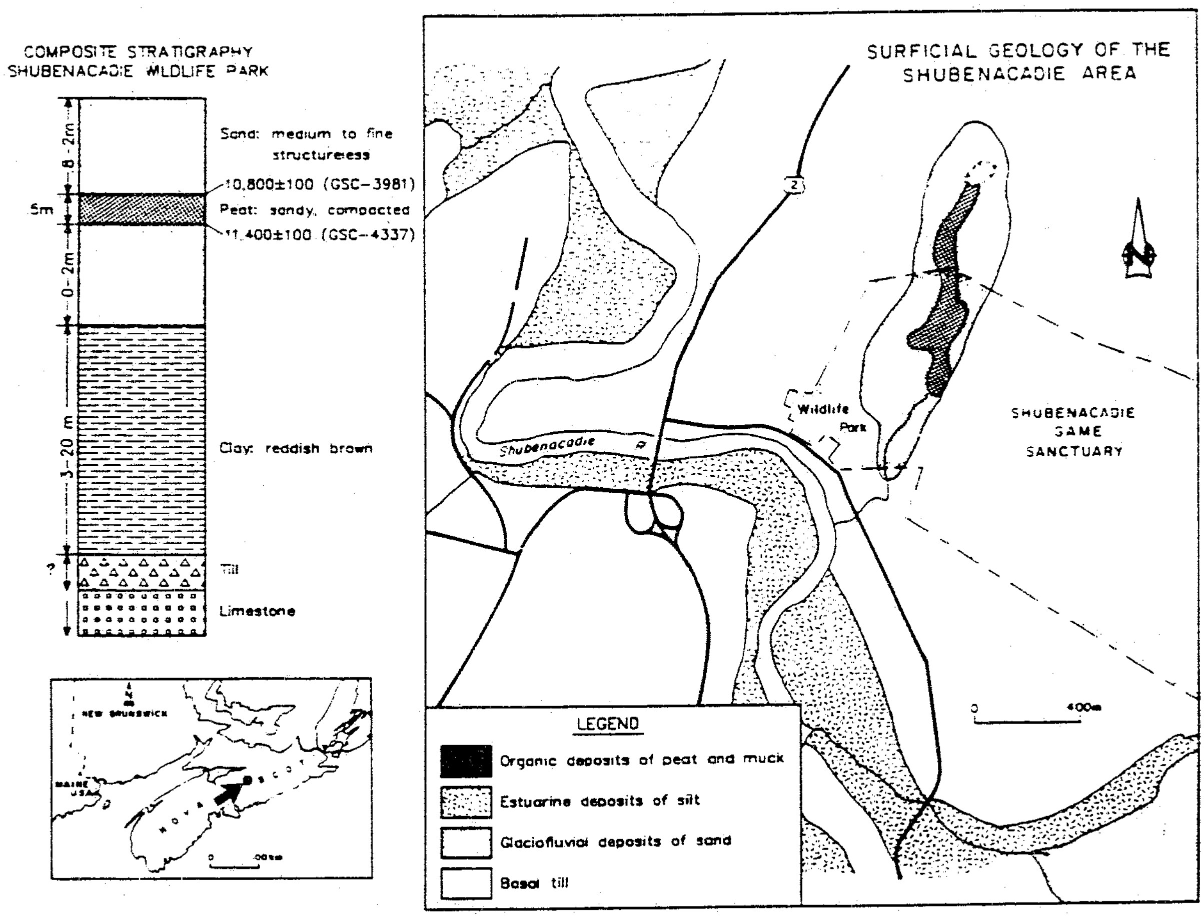

Figure 26. Location, surficial geology and stratigraphy of the Shubenacadie Wildlife Park site.

slightly up-section along with other shrub taxa, whereas Cyperaceae and herb and spore percentages decline. *Sphagnum* abundances reach high values near the top of the peat. In the sand unit above the peat bed, *Picea* declines abruptly, *Betula* (probably shrub birch) increases somewhat and Cyperaceae increases greatly.

INTERPRETATION

The basal clay encountered in boreholes at Shubenacadie is interpreted to be glaciolacustrine. Shallow ponds and wet sedge meadows characterized the Shubenacadie site when the glacial lake drained and the river occupied the valley prior to 11,400 yr B.P. Ferns, club mosses, *Selaginella* and a variety of herbs became locally abundant. Birch, willow and other shrubs bordered the wet areas. Spruce invaded the area early and open woodlands occupied suitable sites on the landscape surrounding the wetlands until about 11,000 yr B.P. Following this time, spruce declined in abundance and birch, grasses and herbs increased as sand deposition inundated the area.

The Shubenacadie sand deposit is one of the sand bodies flanking the Shubenacadie River that was mapped and interpreted as glaciofluvial by Hughes (1957; Fig. 28). The Shubenacadie deposit lies above the floodplain of the Shubenacadie River and extends to near the crest of the adjacent slopes to an elevation of 20 m. Most sand bodies lie on the eastern side of the Shubenacadie River at elevations up to 30 m. Some of them are distal to esker systems and have a deltaic structure with north-westward-dipping foreset beds (Hughes, 1957).

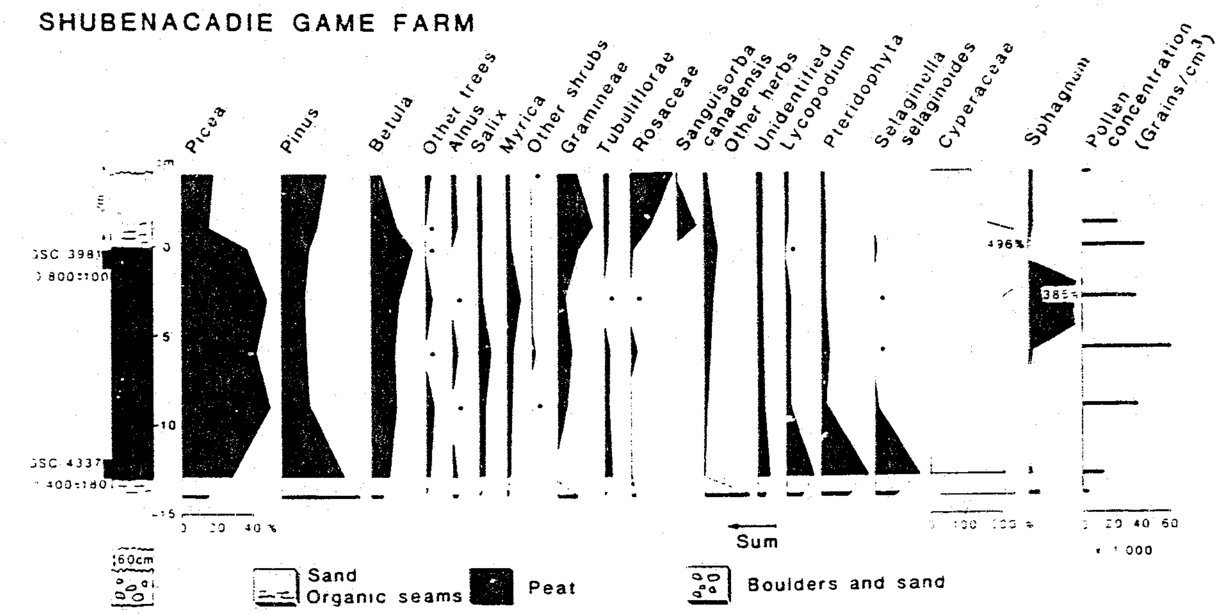

Figure 27. Pollen diagram of the Shubenacadie site.

The grain size and grading of the Shubenacadie sand deposit suggest a fluvial origin. However, a local fluvial origin is unlikely because of the lack of a substantial drainage area to produce this volume of sediment. The clean, well-sorted nature of the sand and its yellowish-grey colour are also difficult to explain if the sand was derived from adjacent red clay till slopes. The circular pond, marking the upper reaches of the sand body, may be a kettle. Therefore, although glaciers did not cover the site, they may have been present in the region east of the Shubenacadie Valley. Outwash from these glaciers was distributed down the east-trending tributary valleys and covered the Shubenacadie peat deposit.

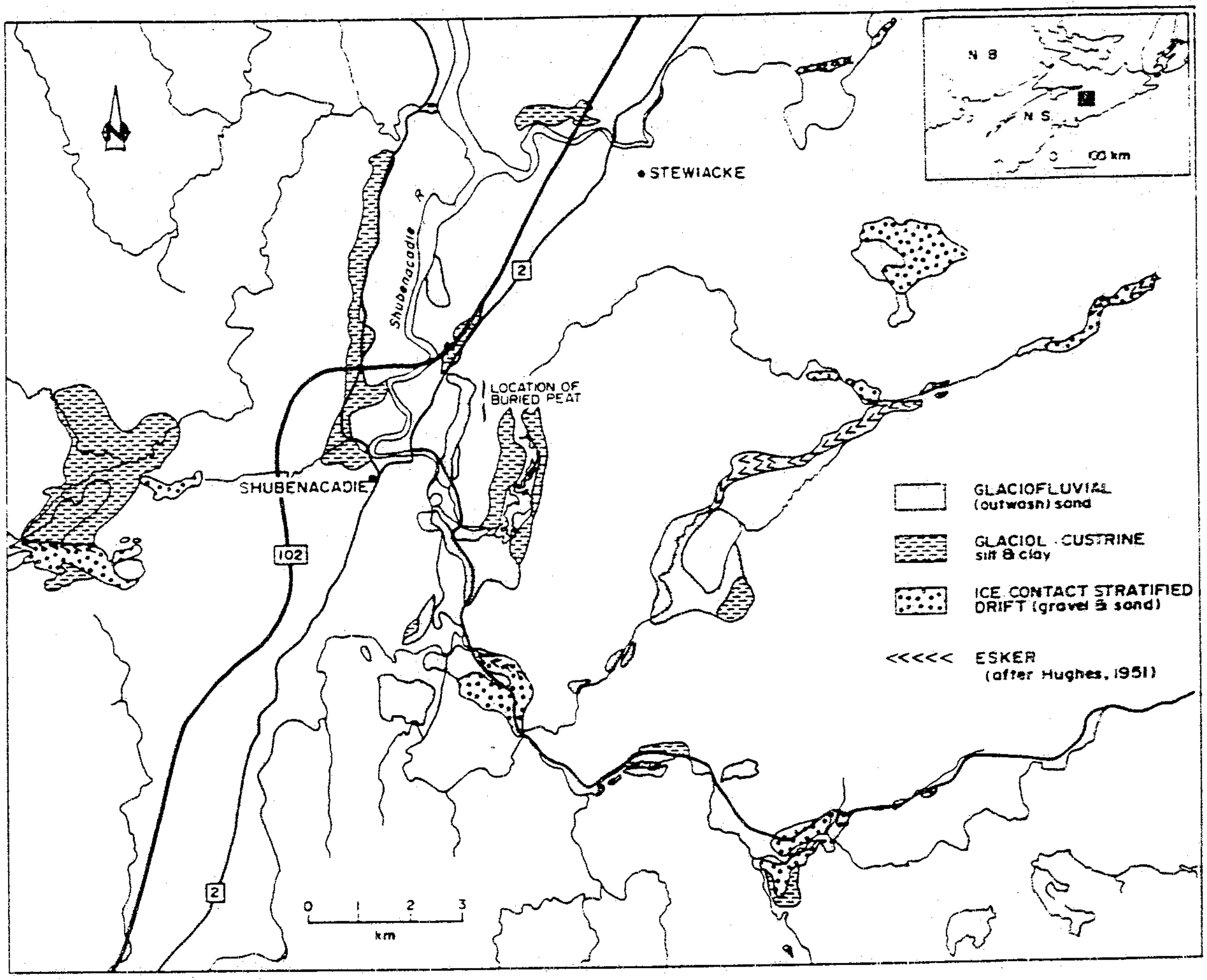

Figure 28. Surficial geology of the Shubenacadie area (after Hughes, 1957).

STOP 1-4: LANTZ

Leaders: R. R. Stea and R. J. Mott

Purpose: To visit another late-glacial peat buried by a diamicton of uncertain genesis.

Route: Leave Stop 1-3 at 1:00 PM. Proceed 15 km south along Highway 2. Turn into Shaw Brickyard on left. Arrive 1:15 PM.

EN ROUTE TO STOP

South of the village of Shubenacadie lie prominent north-trending till ridges (Fig. 29). These are believed to be ice-moulded drumlins. Sand and gravel deposits overlie the drumlin till. Eskers north of the till ridges trend east-west, almost at right angles to the inferred ice flow that produced the ridges. These were produced during the last stage of Wisconsinan ice flow from local centres (Fig. 4; phase 4).

Interglacial/interstadial forest beds have been unearthed under 20 m of till at the National Gypsum Company Quarry at East Milford (Fig. 29). 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,

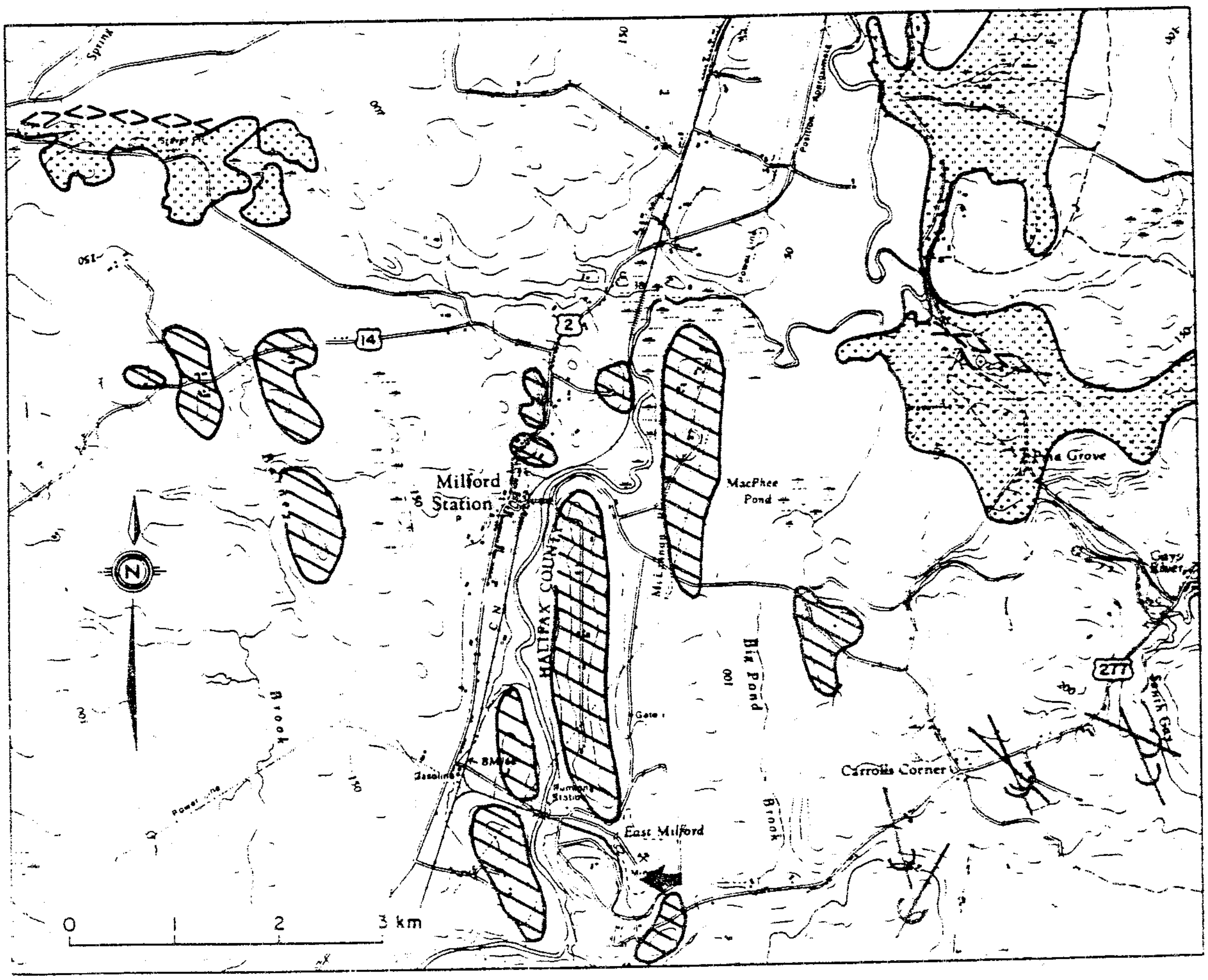

Figure 29. Surficial geology of the route between Stops 1-3 and 1-4 and location of the East Milford gypsum quarry (arrow).

(1975), Prest (1977), Mott *et al.* (1982) and Stea (1982). Numerous sections have been unearthed at the site since 1954 when the pit first opened.

Mott *et al.* (1982) studied a complex organic sequence beneath tills at the site. The stratigraphy in a trench dug by the company was described and samples were collected by R. Grantham and E. Nielsen (Fig. 30). The exposure was subsequently destroyed during overburden removal. The trench revealed gypsum bedrock covered by gravelly grey clay rubble overlain by a sequence of red, grey and black organic clays. Compressed peat covered the clays and the whole sequence was buried by 20 m of till.

Pollen analysis of the black clay and peat provided a profile that was divided into four zones, as seen in Figure 31. The results show a change from basal spectra with abundant ferns and relatively thermophilous hardwood taxa to conifer and alder dominance. A trend from climate similar to present in the area to cool boreal conditions is indicated. A nonfinite radiocarbon date of >50,000 yr B.P. (GSC-1642) was obtained from wood from the peat layer (Table 1). Uranium series ages of 84,200 (UQT-186) and 84,900 (UQT-185) yr B.P. were also obtained from the wood (Table 2) (Mott and Grant, 1985). The deposit may relate to the cooling phase following the thermal maximum of oxygen isotope substage 5e, or if the Th/U dates are valid, to another relatively warm interval, probably substage 5a (Vernal *et al.*, 1986).

INTRODUCTION

The clay quarry is operated by L. E. Shaw Ltd. in Lantz (Fig. 32). Pleistocene clay is mined for the manufacture of brick. Kaolinitic clays of Cretaceous age are also used and we will see the stockpile of these different clay types as we enter the plant. The different clays are blended to produce brick of varying colour and quality.

The site is part of the Hants - Colchester Lowland region which is underlain by Carboniferous and Triassic rocks (Figs. 2, 3). The site is underlain by Early Viséan Windsor Group rocks including gypsum and limestone. The clay quarry area is in a gently undulating plain that lies above the floodplain of the Shubenacadie River. The tidal surge reaches up river as far as the village of Shubenacadie.

STRATIGRAPHY AND CHRONOLOGY

The stratigraphy of the site has been revealed in pits exposed during the last five years. Three sections described below are located on Figure 32. Section A is part of a small rise or low mound. sections B and C are located on the floodplain of the Shubenacadie River (Fig. 32).

Unit I varies from a massive brown clay to rhythmically laminated greyish-brown clay-silt. Calcareous concretions are commonly found weathering out of exposed clay blocks. One was dated at 11,310 yr B.P. (Q.C.-1411; V. K. Prest, pers. comm., 1985). Rootlets were found in the upper 1 m of the clay unit but these did not continue upward into the overlying sand (Unit II). This is the unit being quarried. It underlies most of the area of the brick plant to a depth of 10 m. Auger holes in the area have revealed sand and gravel underneath the clay to thicknesses of 5 m.

Unit II is a grey to brownish sand which becomes oxidized toward the contact with Unit III. An iron-cemented zone is developed near the top of Unit II in some areas.

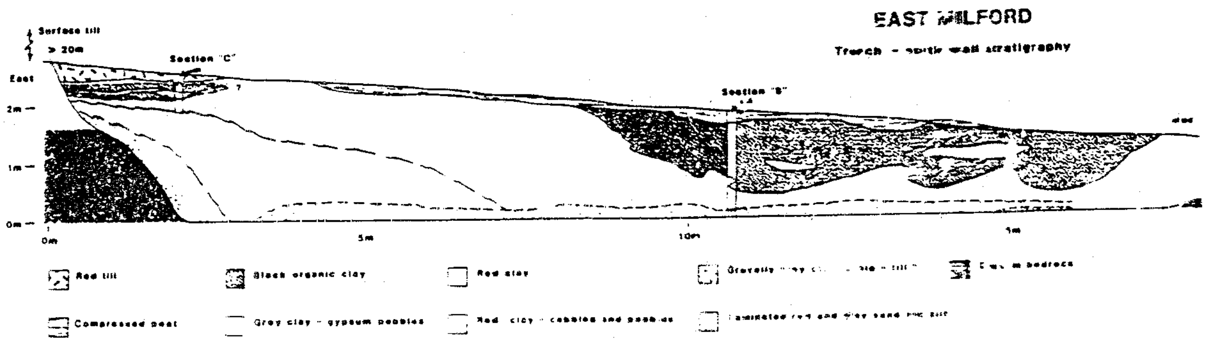

Figure 30. Stratigraphy of the south stripping wall, East Milford quarry.

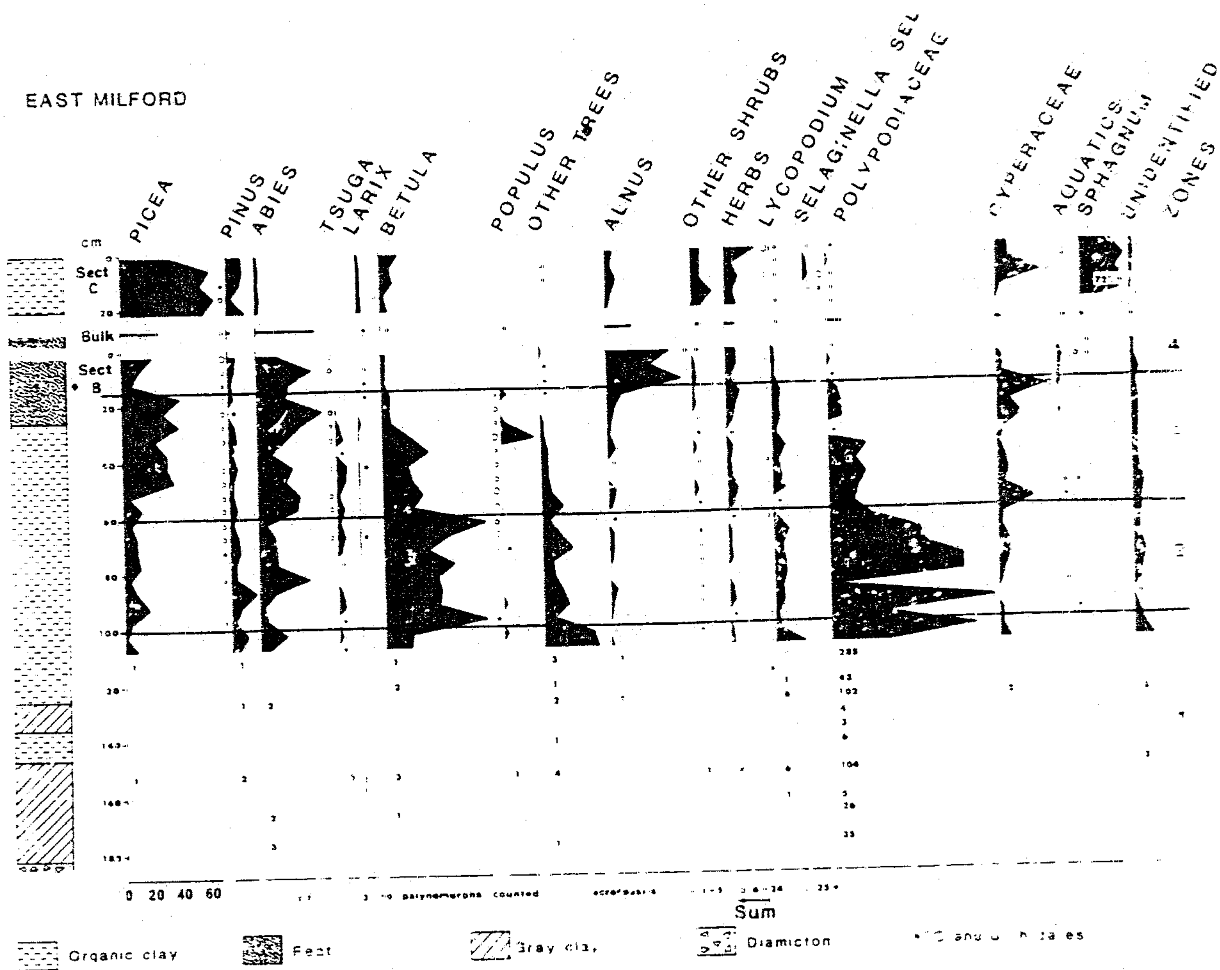

Figure 31. Pollen diagrams for the organic sections at East Milford.

Unit III consists of a peat bed on top of a buff-grey, reduced, silty-clay layer. The silty-clay beneath the peat is up to 15 cm thick and is sometimes gravelly at the base with a discontinuous greenish grey clay zone at the peat contact. The peat layer is irregular and varies in thickness from 0-30 cm. Sand up to 15 cm thick, similar in appearance to that immediately beneath the peat layer, overlies the peat bed. The peat horizon (Unit III) pinches out at the edge of the rise seen in section A. Faulting is evident throughout the unit. A bulk sample of the peat was dated at 11,100 ± 100 yr B.P. (GSC - 3116). One centimetre increments from the base and top of the peat horizon were dated at 11,700 ± 100 yr B.P. (GSC-3774) and 10,900 ± 90 yr B.P. (GSC-3771), respectively.

Unit IV is a reddish, blocky clay-rich diamicton that becomes sandier toward the top. Sand lenses and thin silty partings occur sporadically at the base of this unit along with contorted thin beds of sand and clay. A pocket of steeply-dipping, bedded clay and sand was found within the diamicton. The diamicton forms most of the rise represented by section A. The reddish diamicton is correlated with a stonier, sandier sediment, exposed at section B. that overlies Unit I directly and intrudes into it.

Unit V pinches out over Unit IV at the edge of the rise and lies directly on top of it at the southern end of the quarry nearest the Shubenacadie River. It is a buff, gravelly sand with reddish clay layers. At section B (Fig. 32) the gravelly sand is capped by a reddish clay layer. Organic clay lenses and wood fragments were found in arcuate, channel-fill gravelly-sand beds in Unit V (section C; Fig. 32). A fragment of poplar/aspen (*Populus* sp.) wood from one of these organic layers was dated at 969[illegible] ± 110 yr B.P. (GSC-[illegible]42). Unit VI is a thin [illegible] clay layer (section B, Fig. 23).

Table 2. Thorium/uranium disequillibrium age determinations (see Fig. 8 for locations).

Site	Lab. No.	Age (yr B.P.)
19. East Bay	UQT-175	126,400±15000/12800
	UQT-176	123,400±30000/23400
	UQT-108	106,600±9600/8600
	UQT-179	60,800±5100/5000
	UQT-109	86,900±6000/5700
	UQT-177	62,100±5000/4600
	UQT-227	98,700±10500
	UQT-188	50,200±5000
21. Green Point	UQT-181	117,400±10000/8800
6. Le Bassin	UQT-183	10[illegible],400±8400/8000
	UQT-182	101,700±17000/14[illegible]00
	UQT-184	99,400±8000/71[illegible]0
9. East Milford	UQT-185	84,900±[illegible]500/6100
	UQT-186	84,200±11300/10100
15. Bay St. Lawrence	UQT-178	47,000±4700/4300

PALYNOLOGY AND MACROFOSSIL ANALYSIS

The pollen profile (Fig. 33) covers Units II, III and IV at section A (Fig. 32). *Picea* (spruce) becomes abundant above the base of the peat and remains so to the top. *Betula* values decline slightly along with values for other shrubs and herbs. Cyperaceae also declines somewhat but is still relatively abundant. *Pinus* pollen, although present throughout the profile, is attributed to long distance transport. In the overlying diamicton (Unit IV), *Picea* pollen declines abruptly with a concomitant increase in *Betula* and various herbaceous taxa. Size measurements and morphology of the *Betula* pollen suggest that it probably represents a shrub type. Needles of *Picea* were recovered from the peat along with seeds of Cyperaceae and *Menyanthes trifoliata* (J. V. Matthews, Jr., pers. comm., 1982).

INTERPRETATION

The banded or rhythmic bedding and lack of fossils in the basal clay (Unit I) at Lantz suggest a glaciolacustrine origin. It is postulated that a lake formed in the Shubenacadie River lowlands during regional deglaciation when ice still blocked northward drainage. This lake may either have drained subglacially or southward into Bedford Basin (Fig. 14).

Fluvial deposition (Unit II) followed breaching of the ice dam and drainage of the lake. Organic deposits (Unit III) began to form about 11,700 yr B.P. in shallow ponds and mires that developed in depressions on the former lake plain. Pollen and macrofossil studies indicate that shrub birch and a variety of other shrubs, herbs and ferns bordered wet areas where sedges were abundant. Spruce trees invaded the surrounding area early to form open woodlands with interspersed low shrubs and herbs. A distinct, abrupt environmental change occurred about 10,900 yr B.P. and caused the decline in spruce trees and increase in shrub birch and other shrubs and herbs coincident with the change to inorganic deposition (Unit IV).

Lamination and the pollen content of Unit IV imply that it formed in a lacustrine environment. Some tectonic features imply either direct glacial or periglacial activity. Evidence for sea-level lowering at this time (Stea *et al.*, 1987) mitigates against lake formation; however, lake re-formation in the Shubenacadie Valley could have occurred when ice briefly covered the outlet. Continuous lake sediment records spanning this time interval south at Penhorn Lake (Ogden, 1987) and at Brookfield (Mott *et al.*, 1986) relegate any ice build-up to areas

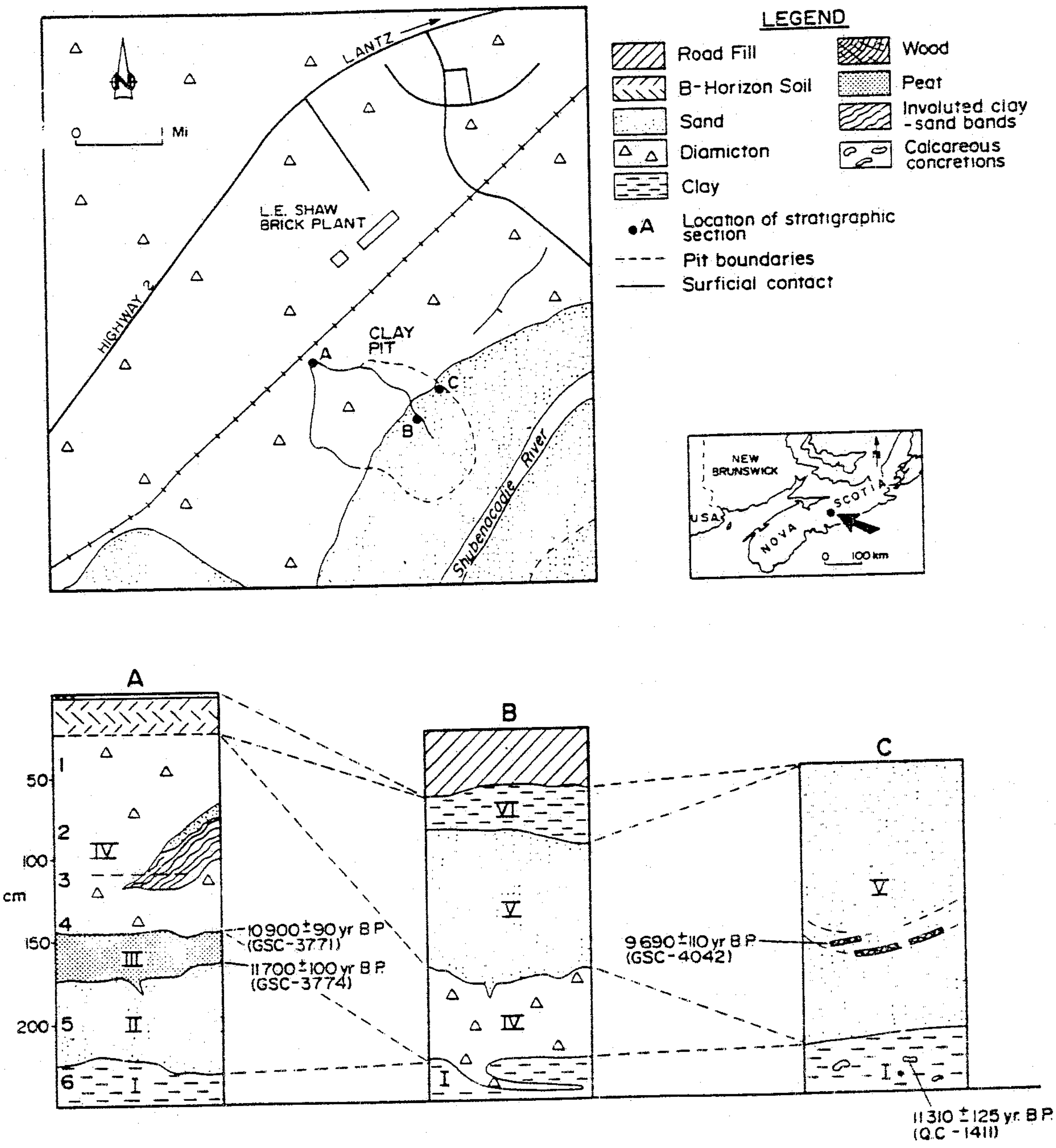

Figure 32. Location and stratigraphy of the Lantz site.

east or west of Lantz.

Units V and VI, the gravelly-sand and silty-clay units overlying the diamicton, were deposited after 9500 yr B.P. These units probably represent a Holocene flood deposit of the Shubenacadie River.

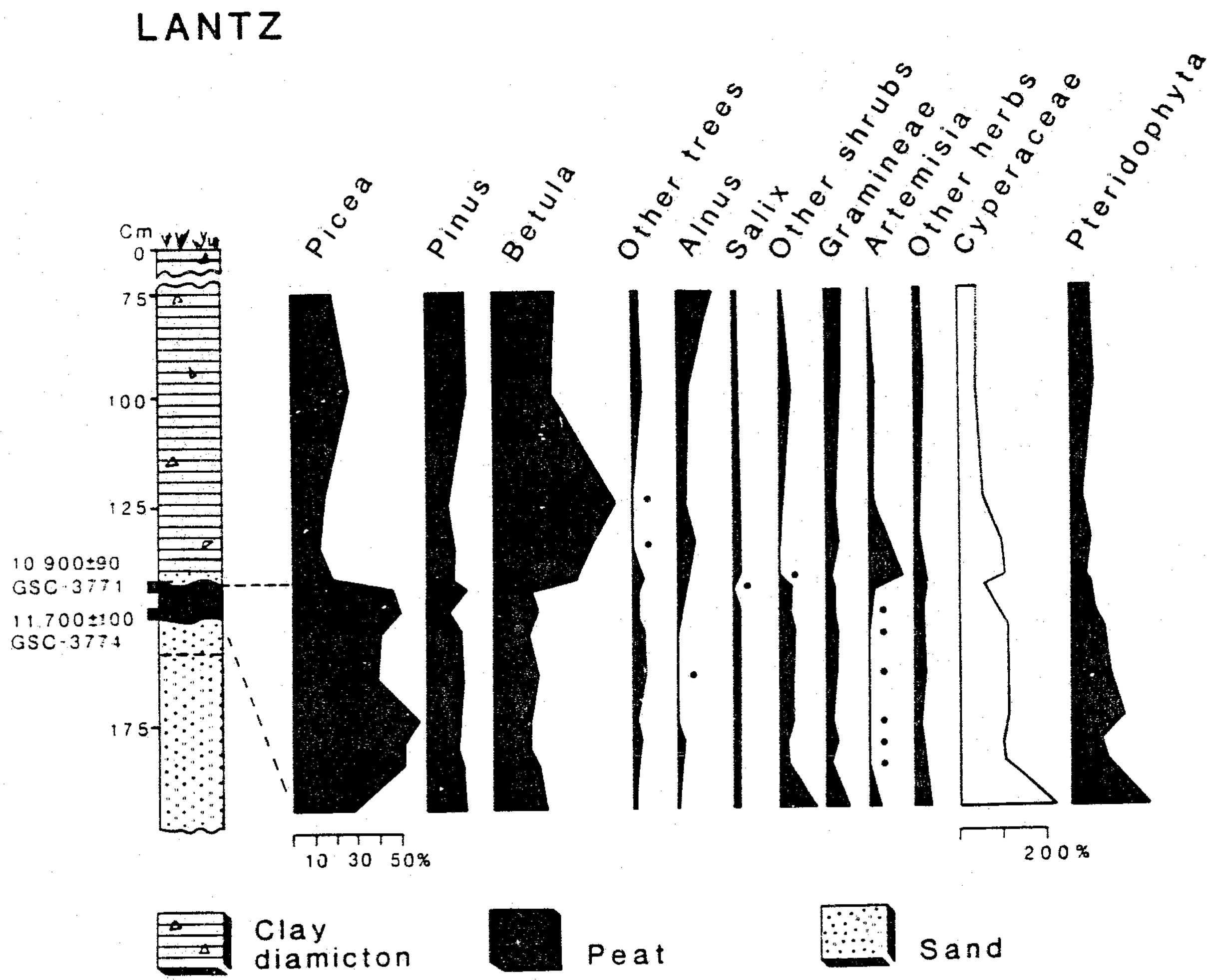

Figure 33. Pollen diagram for the Lantz site.

STOP 1-5: BAILEY QUARRY, MILLER CREEK

Leaders: R. R. Stea, R. J. Mott, R. Miller and R. G. Grantham

Purpose: To visit a gypsum quarry where interglacial/interstadial forest and lake beds are exposed underneath tills and glaciofluvial deposits. Also to visit the site of a recently discovered mastodon tusk.

Route: Leave Lantz 2:15 PM. Follow Highway 22 km to Highway 214. Turn right and drive 6 km. Turn left onto Highway 14. Follow this route 46 km. Turn right on Highway 215. Cross the Herbert River and proceed 1 km. Turn left on #236. Drive 4 km and turn into quarry site. Arrive 3:00 PM.

EN ROUTE TO STOP

Near Nine Mile River (Fig. 34), Nova Scotia Sand and Gravel Ltd. operates a quarry in a 25 km long esker system. Watson (1984) defined six main lithofacies in the esker, ranging from laminated silt to coarse gravelly-sand. He interpreted the ridges as beads deposited at the mouth of a subglacial tunnel. Westward paleoflow was determined, parallel to bedrock striations representing the last phase of ice flow westward from local centres (phase 4, Fig. 4). Unfortunately, most of the landform has been mined.

The village of Rawdon Gold Mines arose as a result of a gold rush in the late 1800s in Nova Scotia. The Rawdon Hills are an outlier of Cambro-Ordovician Meguma Group slate and metagreywacke. The slate hosts quartz veins with sporadic, rich accumulations of gold. In the 1980s a mini-rush brought on by high gold prices and tax incentives culminated in the re-opening of several mines in Nova Scotia.

INTRODUCTION

Bailey Quarry, operated by the Fundy Gypsum Company, is located 10 km northeast of the town of Windsor, Hants County (Fig. 34). The exposure is a stripping wall about 30 m high. The underlying bedrock lithologies are gypsum, limestone and dolostone of the Carboniferous Windsor Group. The area around the section is one of gently rolling topography, primarily drift controlled. Drift thicknesses up to 25 m are common.

STRATIGRAPHY AND CHRONOLOGY

A section exposed in 1978 is shown in Figure 35 (Stea and Hemsworth, 1979). The section was located above the gypsum surface. A greyish clay-rich diamicton with organic debris (Miller Creek Till) overlies the gypsum surface. This is sometimes overlain by a yellowish, oxidized diamicton with dolostone fragments. A discrete peat bed was found overlying the Miller Creek Till with a conformable contact. Evidence of soil formation was seen at the top of the till. Wood fragments from the peat bed were dated at >52,000 yr B.P. (GSC-2694). A finite date of 33,200 ± 2000 had been obtained previously from wood at the site (MacNeill, 1969), but this date is considered spurious. Associated with the peat bed were rhythmically-bedded silt and sand. The section was truncated by cross-bedded sand grading upwards into gravelly sand. A greyish-red, calcareous, massive, silty diamicton with few stones (East Milford Till) overlay the sand with a knife-sharp contact. It was formed by a regional southeastward flow (Phase 1b; Fig. 4). The East Milford Till intruded part of the lower section at many parts of the stripping wall. The section was topped by a reddish, sandy, silt diamicton (Hants Till).

More recently (1987), a section showing black organic clay-silt with large logs over gravelly sand with organic pods between the Miller Creek and East Milford Tills was exposed. Although situated stratigraphically between the same two tills, the exact relationship of the two organic units to each other is not known.

PALYNOLOGY AND MACROFOSSILS

Two distinct pollen spectra characterize the organic beds. The unit exposed in 1978, Facies (A), is dominated by *Pinus banksiana/resinosa* (jack/red pine) pollen and small amounts of *Picea* (spruce) and hardwood taxa (Fig. 36). *Sphagnum* spores are abundant and several bog indicators are present as low values. By contrast, the organic unit exposed in 1987, Facies (B), contains large percentages of

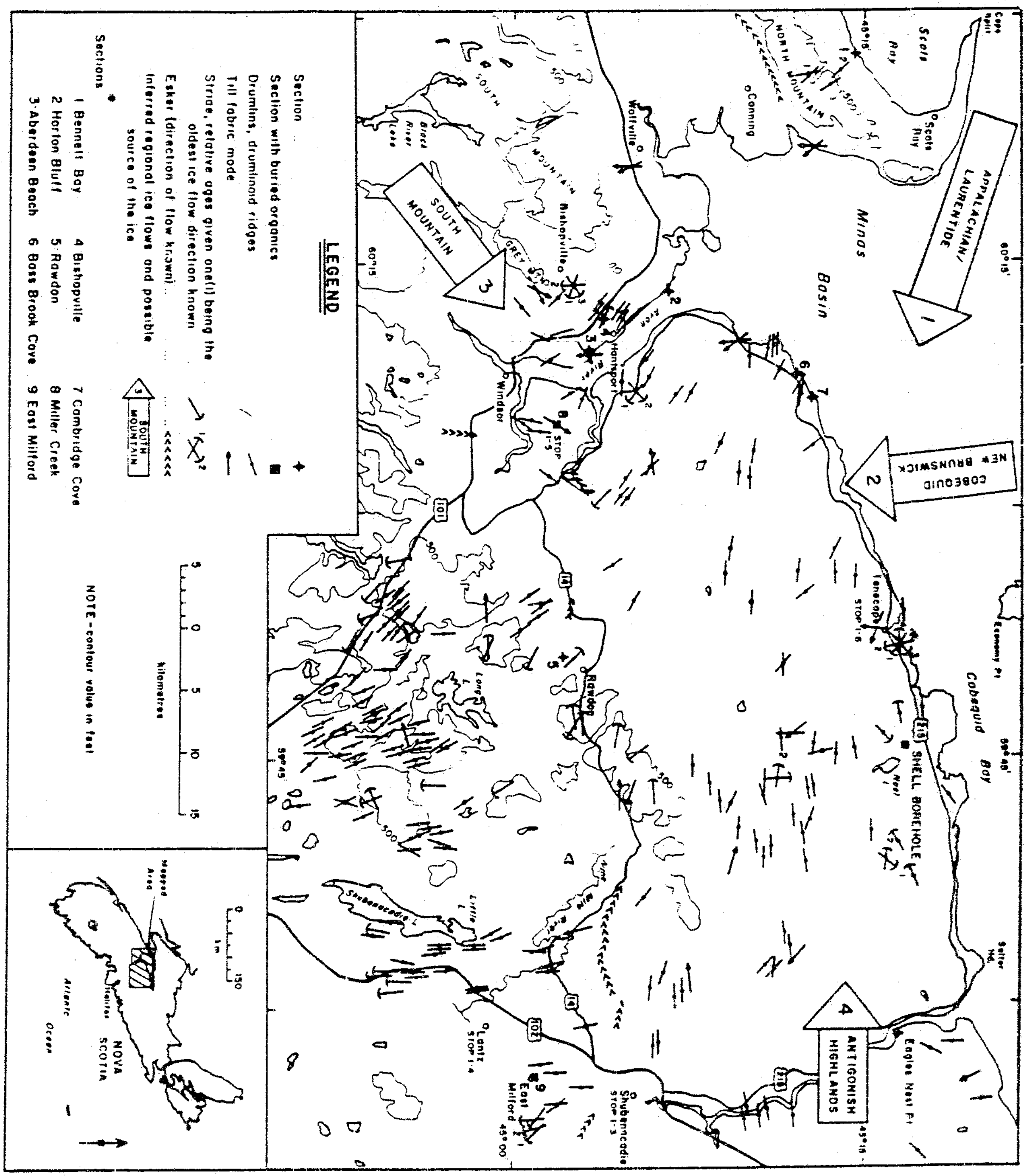

Figure 34. Ice flow features of central mainland Nova Scotia and locations of Stops 1-5 and 1-6.

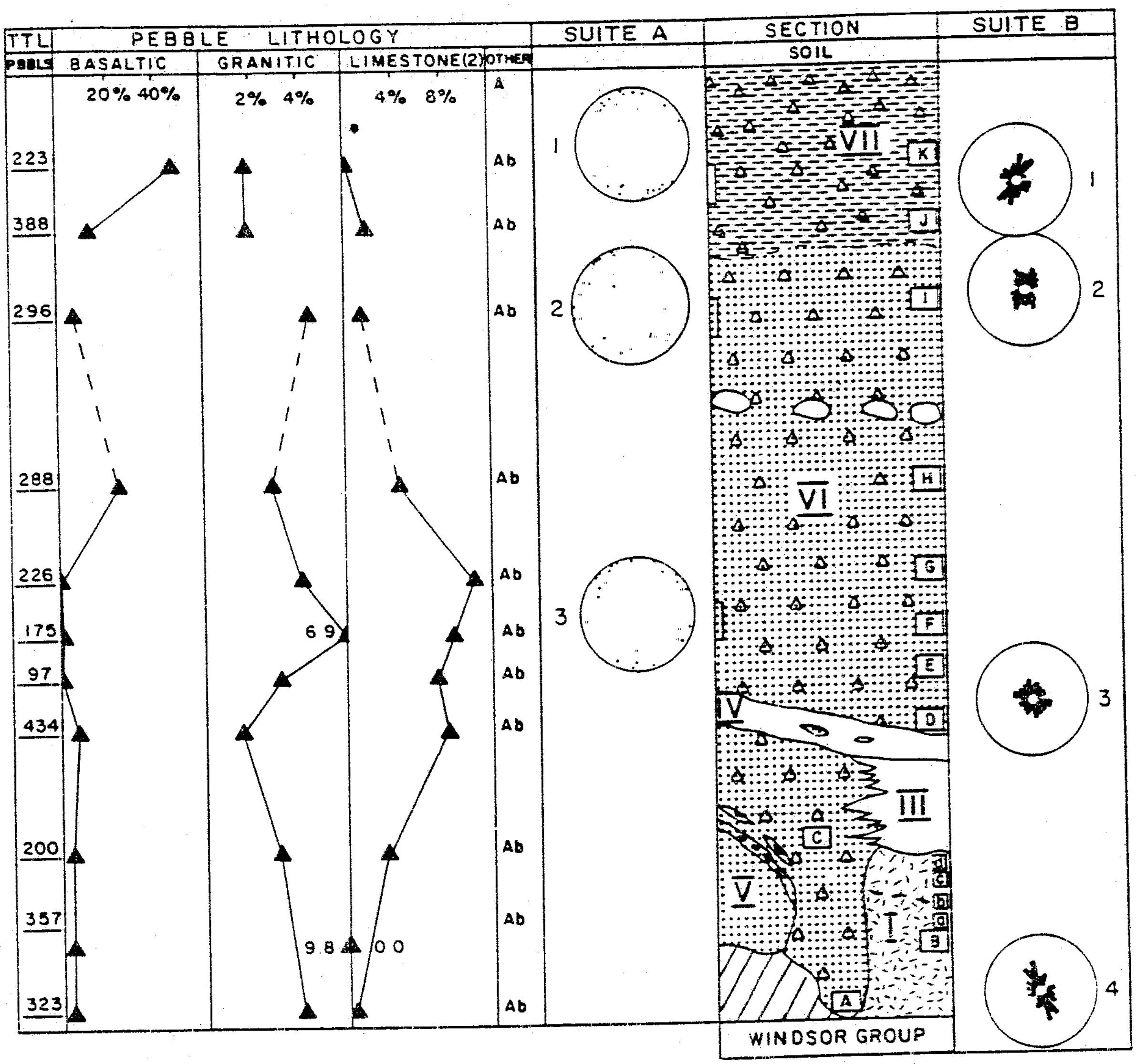

Figure 35. Stratigraphy, fabric, and lithology of the Miller Creek section. Units described in text. Other lithology refers to water-worn quartz pebbles derived from Triassic bedrock, Ab=abundant. Basaltic pebbles are derived from the North Mountain to the northwest (see Fig. 2).

Picea and small values for *Pinus* pollen (Fig. 37). Other tree taxa are minimal as are shrub and herb taxa, although Cyperaceae pollen exceeds 30% at the top of the profile.

Wood from Facies (A) was identified as jack pine, whereas wood of Facies (B) was spruce.

FOSSIL ARTHROPODS

The fossil beetle assemblage recovered from peat associated with the Fundy Gypsum quarry mastodon is typical of a boreal fauna. Only a very small number of specimens were isolated and few could be identified to the species level. However, those specimens that have been identified are indicative of the present boreal zone. The assemblage contained a mix of specimens including ground beetles, staphylinid beetles, aquatic beetles and bark beetles. In general the species can be said to inhabit mossy or sandy, riparian habitats. The two bark beetles live on *Abies* spp. or *Picea* spp. *Carabus maeander*, *Olophrum rotundicolle*, *Phloeotribus*

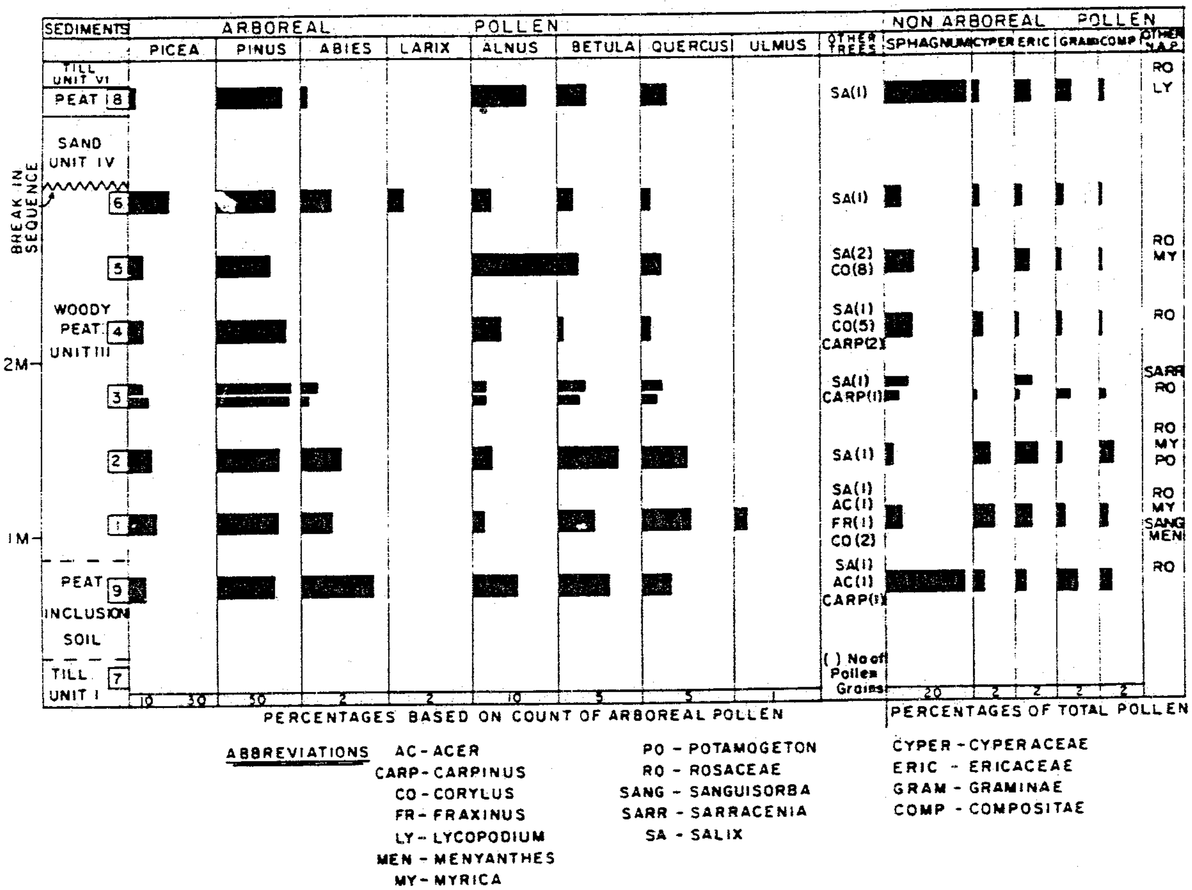

Figure 36. Pollen diagram of organic facies A. Miller Creek (after Stea, 1982).

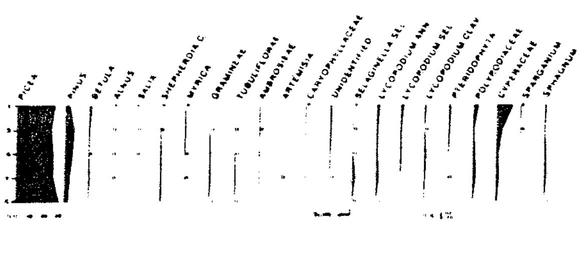

Figure 37. Pollen diagram for organic facies B. Miller Creek.

piceae and *Cryphalus ruficollis* all have transcontinental, boreal distributions (Fig. 38).

INTERPRETATION

Spectra from both organic facies indicate that boreal coniferous forests were present, the *Pinus*-dominated spectra with small amounts of *Picea* and some hardwood taxa of Facies A suggest somewhat warmer conditions than the *Picea*-dominated spectra of Facies B. Both profiles show little variation in their spectra indicating only short time intervals are represented. Without more distinctive trends, interpretations and correlations are difficult.

In their recent syntheses Mott and Grant (1985) and Vernal *et al.* (1986) differentiate three periods of organic accumulation termed Palynostratigraphic Units 1-3 that span a lengthy interglacial period with only Unit 1 attaining true 'interglacial' warmth in the palynological sense. Facies B at Miller Creek

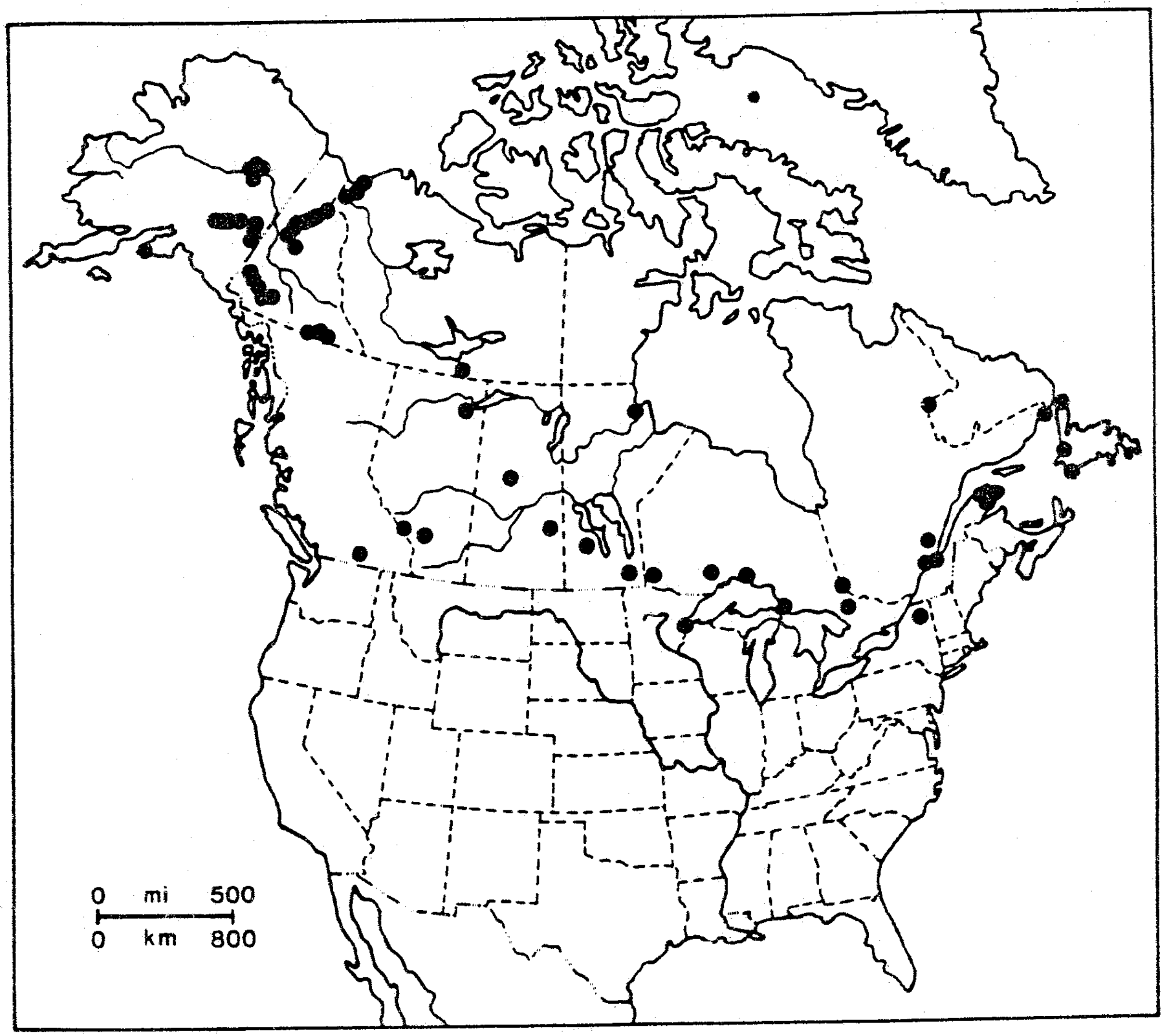

Figure 38. Distribution of the beetle Olophrum rotundicolle in North America.

and Section C at East Milford (Fig. 31) may correlate with Unit 3 because their spectra are boreal in character, that is, dominated by *Picea* and *Pinus* pollen with little or no *Abies balsamea* (balsam fir). The lower part of the profile at East Milford is correlated with Palynostratigraphic Unit 2 which is characterized by the presence of *Abies* with *Fagus* (beech) and *Quercus* (oak). Facies A at Miller Creek may also correlate with Palynostratigraphic Unit 2. The Miller Creek spectra do not indicate conditions comparable to the climatic optimum of the last interglacial (oxygen isotope substage 5e) although the waning phase of 5e is a possibility. It is more likely that the deposits relate to a younger, cooler interval such as substage 5a.

STOP 1-6: TENNYCAPE QUARRY (OPTIONAL)

Leaders: R. R. Stea and R. G. Turner

Purpose: To examine striated outcrops and tills formed by several ice flows.

Route: Leave Miller Creek 4:00 PM. Drive back to Highway 224. Turn left and follow the road along the shore for 56 km. Turn left into driveway of white house past the bridge over the Tennycape River. Arrive 4:45 PM.

EN ROUTE TO STOP

We will be driving along the Minas Basin if time permits, which is at the northwestern end of the Bay of Fundy. The tides of the Bay of Fundy are the highest in the world, reaching 16 m in the Basin. The unusual height of the tides is due to a funnelling effect and to the resonance period of the Fundy Basin. The 12.4 hour period of the semidiurnal lunar tides is close to the natural oscillation period of the Bay of Fundy basin. The effect is accentuated when the moon is at perigee. There is a delay of three hours between high tides in the Bay of Fundy and the Atlantic coast because the tidal water surges up the Bay as a moving mass. This tidal surge tends to pile up on one side of the Bay due to the Coriolis effect. The range of the tides are up to 1.5 m higher in the Minas Basin compared with the Cumberland Basin. The highest tide recorded in historical times was a tide associated with the Saxby Gale in 1869. A combination of high winds up the Bay, abnormally low pressure and a rare alignment of the earth, sun and moon caused the tidal range to reach 21.6 m at the head of the Bay of Fundy.

SITE DESCRIPTION

The site at Tennycape Quarry (Fig. 34) reveals most of the ice flows that affected the area during the Wisconsinan Stage.

Exposed at the quarry is a westward-sloping bedrock surface of Carboniferous siltstone. The bedrock is buried under 8 m of till(s). The lower surface reveals striations trending 120-140°. Bedrock surfaces in the upper part of the quarry are inscribed with two sets of striations, one trending 020-025° truncated by a set of trending 275-285°. Pebble shadows or mini crag and tail features on conglomerates at nearby localities also show the same two ice flows. A boulder of porphyritic granite embedded in the upper part of the till section at the quarry was transported by a flow that crossed batholith areas to the southwest.

Three units can be discerned in the till banks at this site. The lower unit is a silty, greyish-red till with abundant Cobequid Highland erratics. It covers the older, southeastward-trending striations lower in the pit. Above this unit is a stonier till unit, reddish-brown in colour, with a fabric parallel to the northward-trending striation set. The uppermost unit is a thin, oxidized, bouldery till with a sandy matrix.

DAY 2

STOP 2-1: TRURO

Leaders: R. R. Stea and R. J. Mott

Purpose: To examine a site where a late-glacial peat bed is overlain by a glacial? diamicton.

Route: Leave Agricultural college 8:00 AM. Follow Lower Onslow road to Highway 102. Arrive 8:10 AM.

INTRODUCTION

The Truro site is located on the west side of Highway 102, 300 m north of the Onslow bridge at Truro (Fig. 39). It was first exposed during highway construction.

STRATIGRAPHY AND CHRONOLOGY

Description of this site was hampered by the timing of excavation and subsequent regrading (Fig. 40). A peat layer (Unit II) up to 50 cm thick and continuous for 50 m is overlain by 1-2 m of a reddish-brown diamicton with rare cobble-sized stones (Unit III) and underlain by a reddish stony diamicton. The lower diamicton is differentiated from the upper diamicton by an abundance of cobble-sized stones. The fibrous, compacted peat pinches out into the upper diamicton 1 m below surface along the face exposed by excavation. Thin silty clay and sand layers only a few centimetres thick separate the base of the peat bed from the underlying till, whereas at the top of the peat unit thin sand layers alternate with thin peat seams. Peat seams also occur within the upper diamicton.

A date of 12,000 ± 120 yr B.P. (GSC-4297) was obtained from the basal 1 cm of the peat layer (Fig. 40). The top 1.5 cm of the uppermost dateable peat stringer produced an age of 11,400 ± 150 yr B.P. (GSC-4265).

PALYNOLOGY

As in the other sites discussed, the pollen profile at this site (Fig. 41) is dominated by Cyperaceae (sedge) pollen. It is especially abundant at the base where it occurs with smaller but significant amounts of *Salix* (willow) and *Shepherdia canadensis* (shepherdia). Higher in the peat, *Salix* values decline considerably, *Betula* increases to near 20%, and Cyperaceae increase along with Pteridophyta (fern) spores. Toward the top of the peat, where sand layers become abundant, *Betula* declines, *Salix* declines even further, and *Picea*, Gramineae (grass) and Cyperaceae increase. *Salix* and *Juniperus-Thuja* type (probably juniper-cedar) pollen percentages rise in the organic sand horizon as do Pteridophyta spore values.

INTERPRETATION

As at the previous sites sedge meadows occupied the low wet areas when the landscape became available for organic accumulation at about 12,000 yr B.P. Willow and *Shepherdia* were early invaders of the site along with various herbs. Willow soon dominated locally, judging by the extremely large pollen representation, but declined as birch (probably shrub birch) invaded the area. By 11,400 yr B.P. spruce may have just moved into the area in small numbers. It had invaded the Lantz area a few hundred years earlier, it was at the Shubenacadie site by 11,400 yr B.P., and high pollen values and macrofossils confirm its presence at the Brookside site (Mott *et al.* 1986a; Site 9, Fig. 39) a few kilometres east of the Truro site by 11,100 yr B.P.

The lack of high *Picea* values at the top of the sequence, as well as the age, suggests that part of the record seen at the other sites is missing. The diamicton-peat contact may be erosional since peat accumulation continued unabated at the sites to the south between 11,400 and 11,000 yr B.P.

The overlying diamicton is lithologically and structurally similar to local till deposits and is compacted. The sandy nature of this diamicton (Fig. 42) is probably inherited from the underlying Triassic sandstone bedrock, which contributed clasts to the unit. It also contains erratic clasts derived from the Cobequid Highlands Massif to the north. Some of the upper organic horizons appear to have been sheared up into the diamicton unit, perhaps

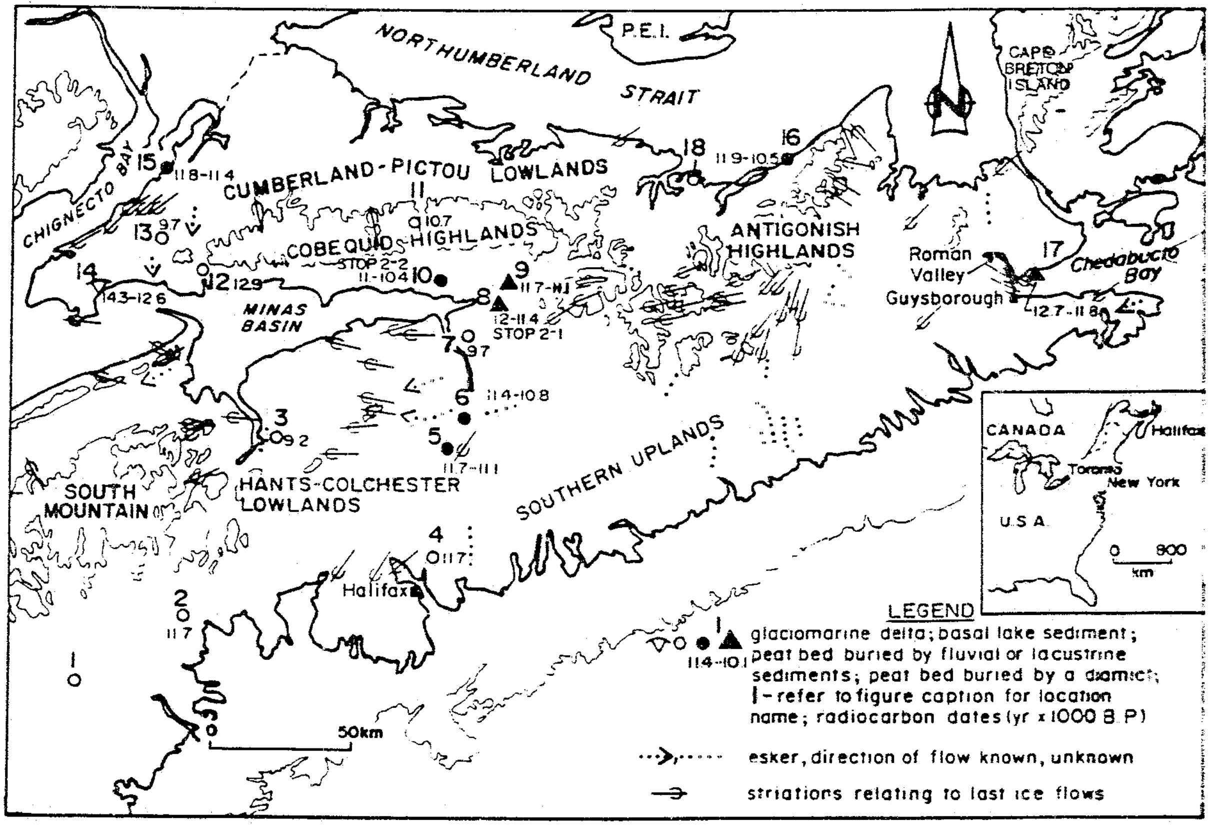

Figure 39. Location of Stops 2-1 and 2-2. Other late-glacial sites in mainland Nova Scotia: 1. Oak Hill Lake (Railton, 1972) 2. Canoran Lake (Mott, unpublished 1987) 3. Shaws Bog (Hadden, 1975) 4. Penhorn Lake (Ogden, 1987) 5. Lantz 6. Shubenacadie Game Farm 7. Brookfield (Mott et al., 1986b) 8. Truro 9. Brookside (Mott et al., 1986a) 10. Debert (Stuckenrath et al., 1966) 11. Folly Lake Bog (Livingstone, 1968) 12. Leak Lake (Stea and Wightman, 1987) 13. Weldon Lake (Mott, unpublished 1987) 14. Spencers Island delta (Stea and Wightman, 1987) 15. Lower Cove (Stea et al., 1985) 16. Lismore 17. Collins Pond 18. Chance Harbour Lake (Jetté and Mott, 1989).

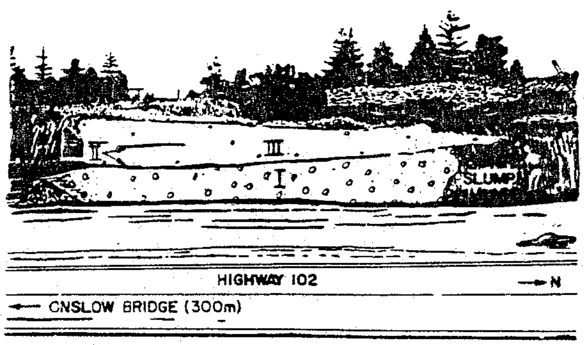

Figure 40. Stratigraphy of the Truro section. Units described in text.

accounting for the possible lack of a recorded interval between 11,400 and 11,000 yr. B.P., present at the other sites. A fabric measurement was not taken in the unit due to the lack of pebble sized stones. A colluvial origin for the diamicton is unlikely as it is near the crest of a drumlinoid hill and thickens uphill.

TRURO SECTION

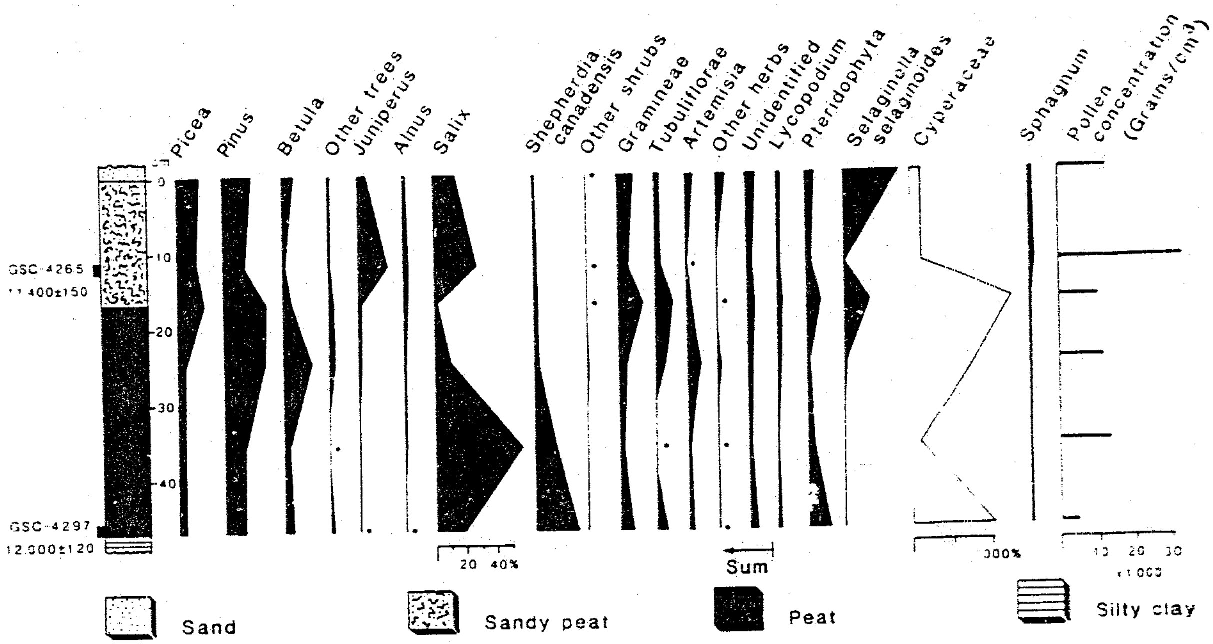

Figure 41. Pollen diagram of the Truro section.

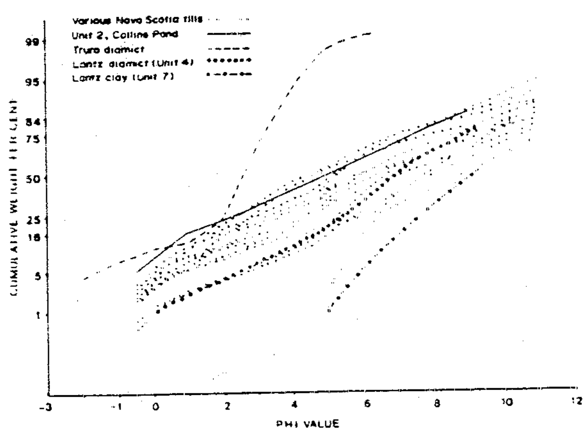

Figure 42. Grain size distributions of late-glacial diamictons in mainland Nova Scotia.

STOP 2-2: DEBERT

Leader: S. Davis

Purpose: To visit the paleo-Indian site at Debert and a newly discovered site at Belmont.

Route: Leave Truro site 8:40 AM. Travel east 8 km along Onslow Road to Game Sanctuary. Turn right on McElmons Road. Proceed north to Plains Road. Turn off on Lancaster Cr. Pull into tree breeding centre at 9:10 AM.

THE OLDEST FRIENDS OF THE PLEISTOCENE

INTRODUCTION

The Debert/Belmont paleo-Indian complex is the oldest evidence for human habitation in the Maritime Provinces. The sites have provided data toward the interpretation of an occupation known as the Clovis Period for the far northeast. The artifactual material represents a continuum of a broad, continent-wide adaptation to the hunting of late Pleistocene fauna. The people involved in this adaptive strategy were the descendants of the immigrants who crossed Beringia and settled in the heart of North America. From this core area they spread west to California, east to New England and northeast into the Maritimes. The carbon dates from the 1963-64 excavations at Debert gave an average age of 10,600 $\pm$ 47 yr B.P. for the occupation of this site. The two new sites at Belmont have yet to be excavated, having only been discovered late in the fall of 1989. Therefore the summary of the data presently known from Nova Scotia is taken from MacDonald (1968).

SUBSURFACE GEOLOGY

MacDonald (1968) identified two aspects of the subsurface deposits at Debert that he deemed to be of archaeological significance: structural horizons and soil-weathering horizons. Their determination is complicated by the fact that they are derived from a single parent matrix, the underlying site bedrock of red Wolfville Formation sandstone. The stratigraphy was poorly defined as interfaces based on changes in colour or texture were difficult to determine. Further, over the past ten millennia factors such as ice, water, wind, organic agents and insects have reworked each stratum. These same agents have also contributed to the vertical displacement of the artifactual materials. A typical section of a Debert deposit is provided as Figure 43, a detailed discussion of the various horizons can be found in MacDonald's site report.

CULTURAL EVIDENCE

Although various natural agents caused vertical displacement of artifacts the horizontal positioning was not significantly altered. MacDonald (1968) used the distribution of artifacts and other cultural features to address a number of problems concerning the occupation of the site. The excavations recorded a total of 4,471 lithic specimens associated with 25 distinct features. These were used by MacDonald to define eleven discrete activity areas with eight being concentrated across a ridge in the central portion of the site. This same area produced the major percentage of artifactual material (92%).

The lithics represented in the collections from Debert are dominated by two varieties of chalcedony. The most common (40.4%) is an opaque, brecciated variety followed by a translucent chalcedony (34.2%). The fact that few cortex flakes of either variety were found suggests that the initial manufacturing of tools occurred somewhere other than on the site. An attempt to locate the source of these chalcedonies proved unsuccessful; however, a best guess suggested a location offshore near the present community of Parrsboro. The remaining lithics are represented by silicified siltstone of unknown origin and porphyritic rhyolite. The rhyolite is locally available in the underlying till.

MacDonald recognized eleven different types of tool manufactured from the various lithics. The most diagnostic was a fluted projectile point that closely resembles the Clovis type (Fig. 44). The highest number was represented by end scrapers (34.68%) followed by retouched flakes (19.79%). The percentage of scrapers would suggest that animal hide processing was a dominant activity at the Debert site.

It is anticipated, on the basis of the preliminary

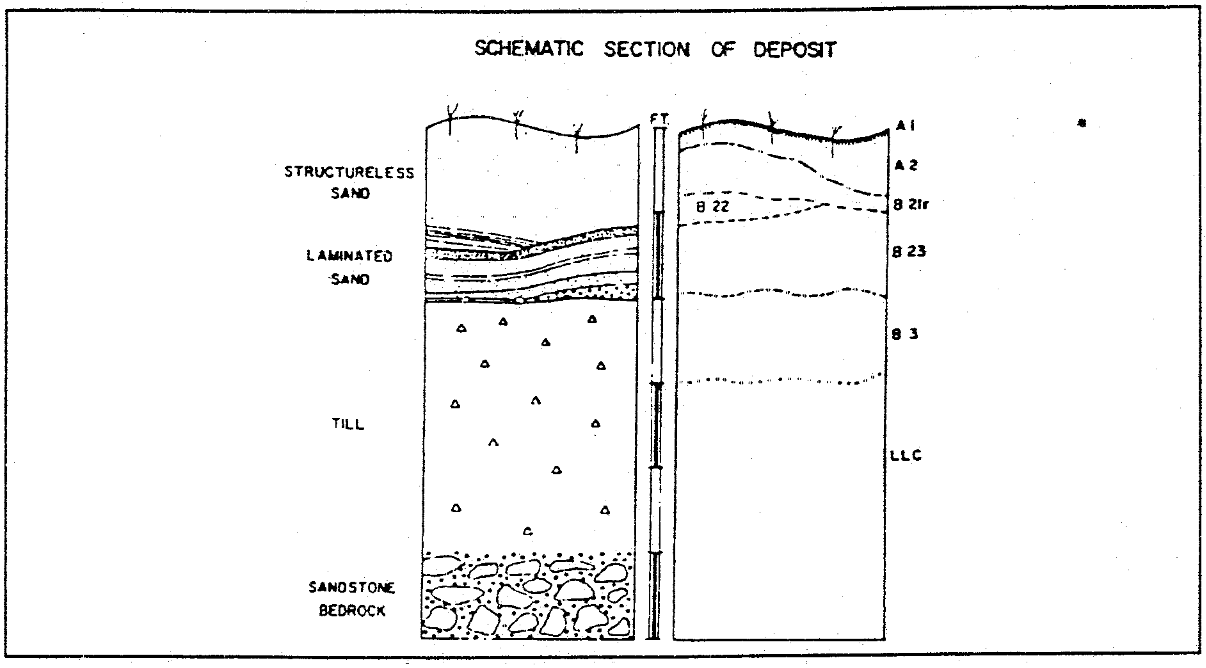

Figure 43. Schematic section of Debert deposit (after MacDonald, 1968).

investigations, that the new sites will duplicate the results achieved by MacDonald. However, it is now clear that a much more complex culture history exists within the Debert/Belmont area. With additional archaeological, palynological, geological and soils analyses we are in a position to gain a better understanding of the events surrounding the occupation of Nova Scotia by the "oldest Friends of the Pleistocene".

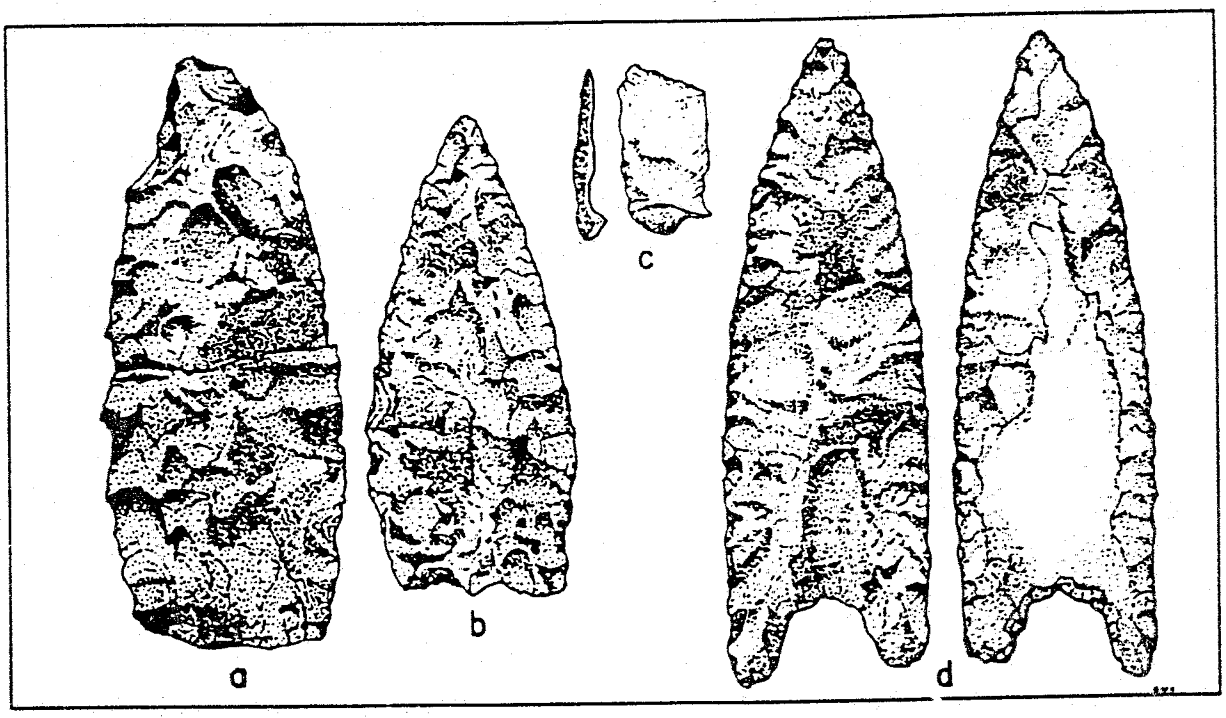

Figure 44. (A and B) Projectile point preforms, (C) Channel "flute" flake, (D) Clovis type fluted projectile point (after MacDonald, 1968).

STOP 2-3 CHANCE HARBOUR LAKE

Leader: R. J. Mott, P. W. Finck and R. R. Stea

Purpose: View ice marginal hummocky moraine and stratigraphy and palynology of a lake within the moraine.

Route: Leave stop 2-2 10:00 AM. Travel 6 km southward on McElmons Road. Drive eastward on the Trans-Canada 86 km. Turn left at the junction with Highway 348 just past the East River. Follow northward 5 km. Take dirt road north 12 km to Boat Harbour. Turn left on Highway 346. Pull into Lewis Road and into gravel pit. Arrive 11:00 AM.

EN ROUTE TO STOP

From Stop 2-" we will be travelling along the Trans-Canada Highway over Mt. Thom. This high ground is called the Cobequid Highlands. It is made up of a suite of Precambrian to Carboniferous igneous, volcanic and sedimentary rocks. Many of these rocks are distinctive from rocks of the Meguma terrane to the south, which is separated from the Cobequid terrane by a major suture, the Cobequid-Chedabucto fault system (Donohoe and Wallace, 1982). This unique bedrock terrane provides indicator erratics that have been used to reconstruct ice movements in the Province.

INTRODUCTION

The Chance Harbour area is characterized by arcuate, linear ridges (Fig. 45) that are composed of a clast-dominated, sandy diamicton termed the Toney River Till. The Toney River Till is almost entirely made up of local red and grey Carboniferous sandstone and the clasts are generally angular. Chance Harbour Lake lies within the morainal topography. It was cored and analyzed by Jetté and Mott (1989).

STRATIGRAPHY AND CHRONOLOGY

A core 850 cm in length was recovered from Chance Harbour Lake and revealed 755 cm of gyttja overlying 60 cm of pink clay and 18 cm of organic silt with marl layers above the basal pink clay (Fig. 46). Organic content increases above the basal clay and into the organic silt, declines abruptly in the overlying clay layer and increases again in the gyttja (Fig. 47). This stratigraphic sequence is apparent in numerous lake profiles throughout Atlantic Canada (Mott *et al.*, 1986b). Seven radiocarbon dates were obtained at critical depths in the core and are shown adjacent to the stratigraphic column in Figure 47.

PALYNOLOGY

The basal herb/shrub zone with abundant Cyperaceae (sedge) and *Salix* (willow) within the basal clay gives way to a *Betula/Populus/Juniperus* (birch/poplar-aspen/juniper) zone in the organic silt (Fig. 48). *Picea* rises abruptly near the top of the organic silt but declines abruptly in the clay where herb and shrubs dominate once again In the overlying gyttja, a *Betula/Picea* zone is followed successively by *Pinus* (Pinus), *Tsuga* (hemlock), *Betula*, *Tsuga/Fagus* (beech)/*Betula* and *Picea* pollen zones.

INTERPRETATION

The morainal topography appears to relate to an ice marginal stand, but the glacier's source is uncertain. The last major ice flow event in this region was a strong, northeastward ice flow during ice flow phase 3 (Fig. 4) that was directed along the Northumberland Strait and eventually out along the Cape Breton Channel, merging with an ice stream coming out of Georges Bay (Fig. 4). Numerous striation sites and northward-transported Cobequid Highland erratics are local evidence for this flow. Inland, this northeastward flow was followed by a westward flow from a centre in the Antigonish Highlands (ice flow phase 4; Fig. 4). Evidence for a corresponding event in this area, however, is equivocal. This morainal area should, therefore, represent a marginal stand of a southwestward retreating glacier. The shape of the morainal ridges (convex to the northeast), breaching of the ridges by meltwater channels, and ice contact faces suggest that ice flow was landward. The lack of Cobequid erratics in the morainal till corroborates this interpretation. The age of this event would have to predate 12,000 yr B.P., before organic accumulation at Chance Harbour Lake.

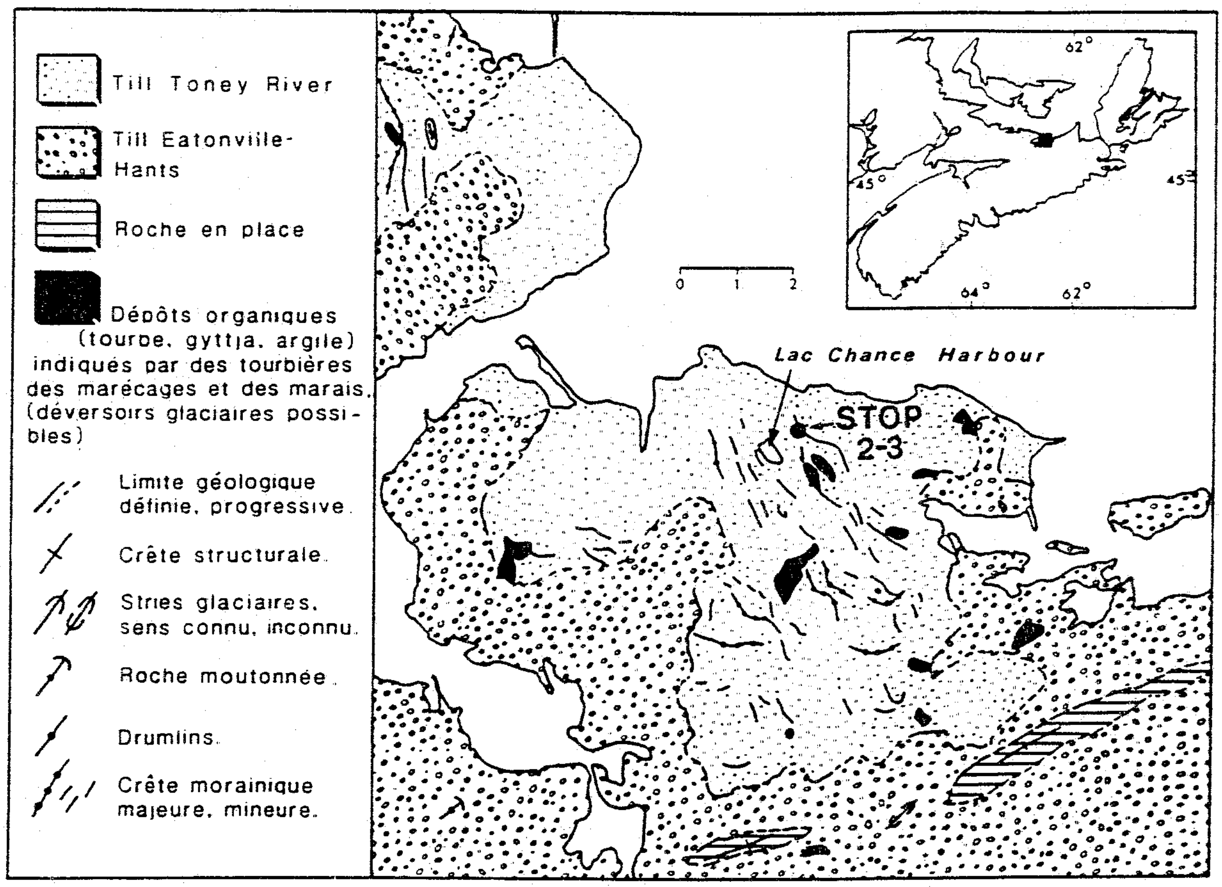

Figure 45. Location and surficial geology of Stop 2-3, Chance Harbour Lake.

The basal four dates are considered anomalous due to hard water error and an alternative chronology was proposed by Jetté and Mott (1989) based on comparison with similar stratigraphy at several other sites where the dates are considered reliable (Fig. 47). A basal age of 11,900 yr B.P., an age of 11,000 yr B.P. for the top of the organic silt and 10,000 yr B.P. for the base of the gyttja are probably more accurate estimates, although these should not be considered exact.

The pollen stratigraphy indicates that early herb/shrub communities of sedge and willow were invaded by aspen/poplar, shrub birch and juniper and then by spruce trees, probably as open woodlands, as the climate warmed after about 12,000 yr B.P. A climatic cooling about 11,000 yr B.P. then caused an abrupt decline in spruce and a return to shrub and herb dominated communities. Continuation of the warming trend about 10,000 yr B.P. promoted the return of spruce and birch followed by successive changes as pine, hemlock and various temperate hardwood taxa migrated into the area. The climatic oscillation near the base of the profile has been equated to the Allerod/Younger Dryas event of Europe and the North Atlantic Ocean (Mott *et al.*, 1986, Jetté and Mott, 1989).

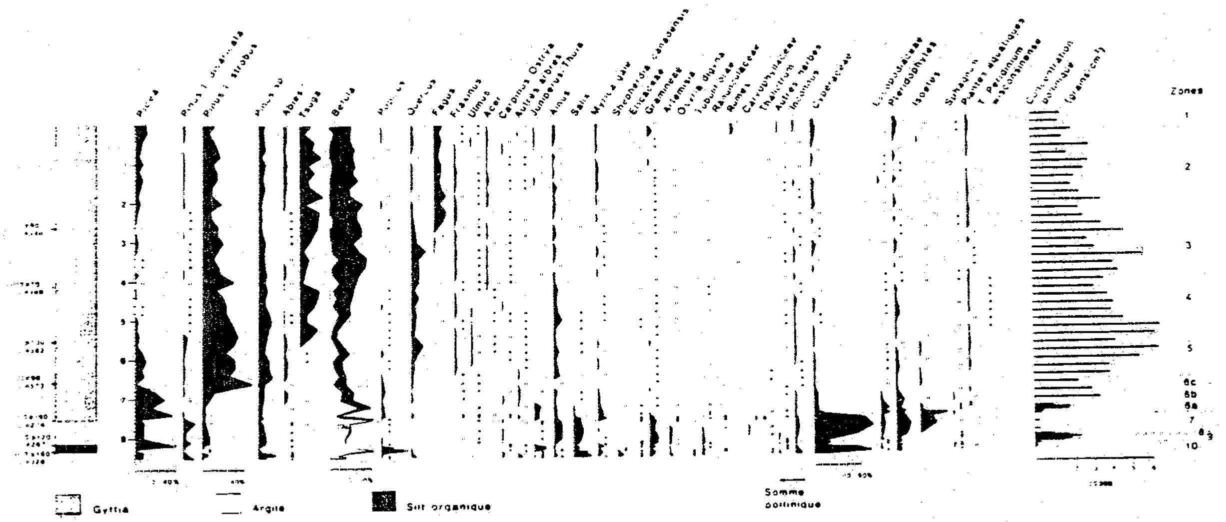

Figure 46. Complete percentage pollen diagram for Chance Harbour Lake site showing stratigraphic column and radiocarbon dates (after Jetté and Mott, 1989).

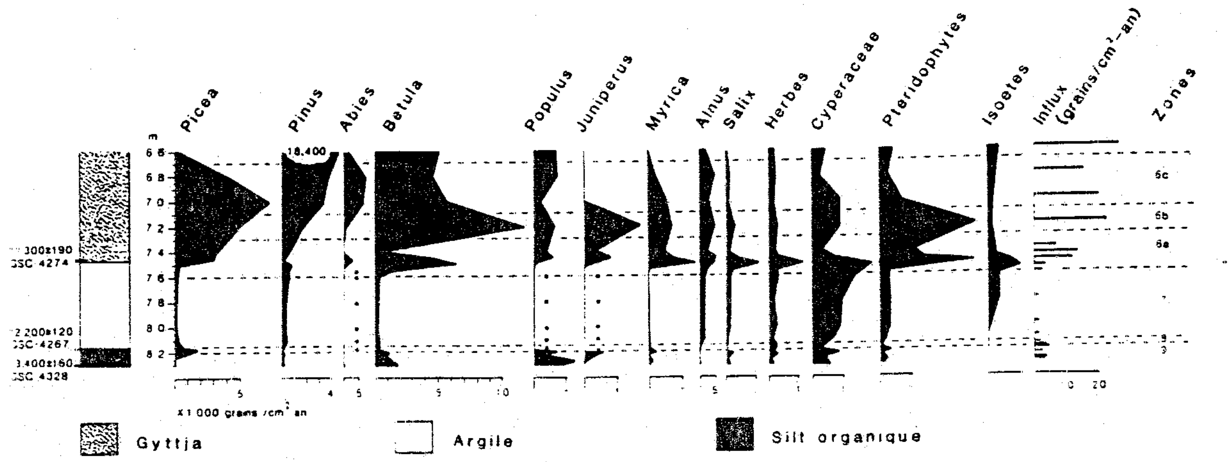

Figure 47. Abbreviated pollen influx diagram for base of Chance Harbour Lake core (after Jetté and Mott, 1989).

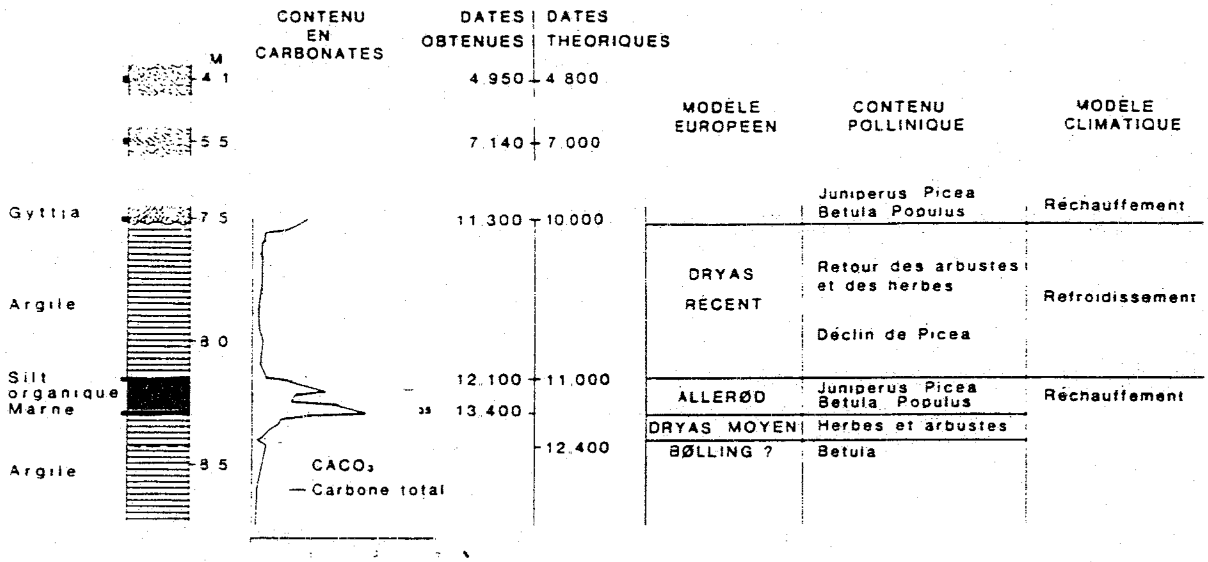

Figure 48. Diagram showing detailed stratigraphy for base of Chance Harbour Lake core, total carbon and carbonate content of sediments, radiocarbon dates obtained compared to postulated ages and a comparison of the climatic interpretation to the late-glacial European model (after Jetté and Mott, 1989).

STOP 2-4: LISMORE

Leaders: R. J. Mott, G. Brewster, R. Miller and R. R. Stea

Purpose: To visit one of the reference sections for buried late-glacial peats in the Province. Longest late-glacial record on the mainland of Nova Scotia spanning 11,800 yr B.P. to 10,500 yr B.P.

Route: Leave lunch site 12:30 PM. Travel along Highway 348 7 km to Woodburn turnoff. Turn right up the hill at the Woodburn sign. Follow to junction with Highway 4. Turn left towards Sutherlands River. Merge with Trans-Canada. Drive 2 km east. Turn left at junction with Highway 245 (Merigomish). Drive northward for 21.2 km to Lismore site on dirt road past second Bailey Brook sign. Turn left onto mobile home driveway before G. D. Miller sign. Arrive 1:15 PM.

INTRODUCTION

The Lismore site is located along the Northumberland Strait shore 10 km southwest of Arisaig near the village of Lismore (Fig. 49). The section is located in an outwash delta complex that formed when northward-flowing ice began to retreat toward the Antigonish Highlands. At some localities the outwash contains 35% igneous and volcanic clasts from the Antigonish Highlands.

STRATIGRAPHY AND CHRONOLOGY

The section consists of Silurian sandstone overlain by 2 m of red medium sand to fine gravel, up to 40 cm of fibrous peat, the upper half of which is very sandy, and 50-150 cm of red, coarsely laminated or massive fine sand with red silty clay seams and pods (Fig. 49). The base of the peat is about 3 m above sea level. Sand in contact with the organic sediments is bleached grey. The modern soil with its leached zone has developed in the upper sand. Southwest along the shoreline, cliffs reveal a thicker section of waterlain sediments, with coarse to fine red laminated sand dipping seaward at the base, and coarse gravelly sand on top. The stratigraphic relationship of the sand covering the peat bed and the gravelly sand in the adjacent section to the southwest is uncertain.

Increments 2 cm thick from the base and top of the peat bed gave radiocarbon dates of 11,900 ± 100 yr B.P. (GSC-4153) and 10,500 ± 120 yr B.P. (GSC-4156), respectively. A sample from the contact of the sandy peat and fibrous peat dated 10, 600 ± 100 yr B.P. (GSC-4762).

PALYNOLOGY

Pollen results from the Lismore site are shown in Figure 50. As at the other sites, Cyperaceae (sedge) is the dominant pollen taxon throughout the sequence but it declines from high values at the base of the compact peat, rises again in the sandy peat but gradually declines toward the top of the organic deposit. Within the lower peat, *Betula* (probably shrub birch) and *Picea* show increases upward to a maximum at the contact with the sandy peat. The *Pinus* (pine) pollen profile is probably attributable to long distance transport because its abundance is inversely proportional to pollen concentrations, which rise from low values at the base to a maximum at the top of the compact peat, and low values above. In the sandy peat, *Betula* and *Picea* decline abruptly, and Cyperaceae, *Salix*, *Alnus* (alder), Gramineae, and *Artemisia* increase.

FOSSIL ARTHROPODS

The richest samples were found near the bottom of the core in the samples that were the most peaty. Based on the available radiocarbon chronology, these samples are probably older than 11,000 yr B.P. Species of staphylinid beetles were most abundant at Lismore. Total numbers of individuals were also higher for this group of beetles. The fossils identified to the species level include *Acidota* cf. *A. quadrata*, *Olophrum consimile*, *Olophrum rotundicolle* and *Tachinus* cf. *T. tachyporoides*. Other staphylinid beetles included *Stenus* spp., and *Eucnecosum* sp. The most common ground beetles were *Dyschirius* spp. and beetles from other families were also present. All these beetles are typical of riparian habitats, either living in *Carex* or *Sphagnum* bogs, under leaf litter or on sandy shores of standing or slow moving water. All the species identified above are typical of the boreal forest and their presence suggests conditions cooler than at present (Fig. 39).

Near the top of the core, beetle fossils are less

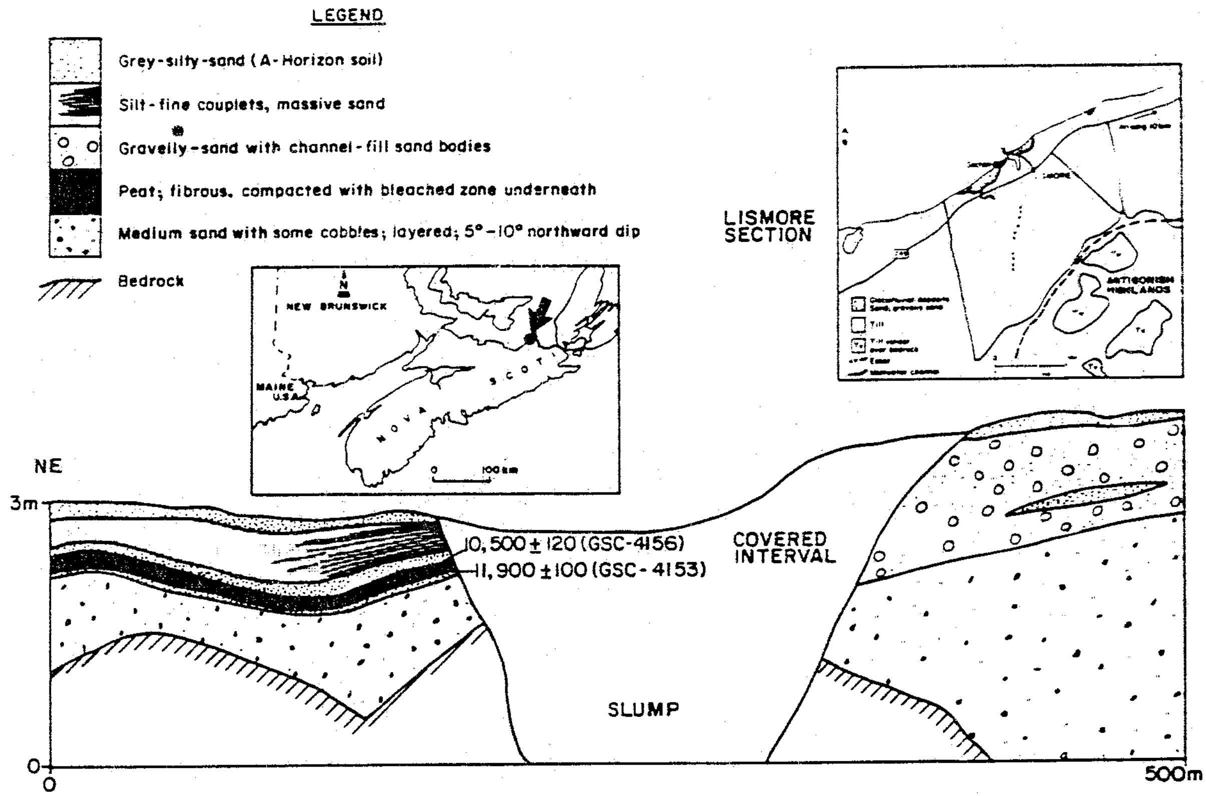

Figure 49. Location, surficial geology and stratigraphy of the Lismore site.

common until only a very few fragments were recovered from the top of the section. The beetle evidence from this site suggests that boreal conditions existed at the site prior to 11,000 yr B.P. However, it does not indicate any change to suggest a cooling climate associated with the younger Dryas event.

INTERPRETATION

The southwest section is interpreted to be an outwash delta deposit with the coarse facies representing topset beds and the fine sand representing foreset beds. Sand underlying the peat section is correlated with foreset sands in the southwest part of the section. It is not known whether the deltas along the Arisaig shore are marine or freshwater. The lack of Late Wisconsinan submergence features in the region and the lack of fossils would suggest that they are freshwater in origin, forming in ice-dammed lakes.

Shallow, wet depressions on the delta surface quickly became sedge meadows and peat began to accumulate about 11,900 yr B.P. Birch shrubs were prominent surrounding the wet areas along with other shrubs, herbs and grasses. By approximately 11,000 yr B.P. birch shrubs had become more abundant on the landscape and vegetation in general was more prolific. Spruce trees probably had not yet reached this area but the moderate *Picea* pollen peak suggests that spruce was migrating toward the area. Pollen analysis at Chance Harbour Lake about 30 km to the southwest (Fig. 47) indicates that spruce trees had migrated there by this time (Jetté and Mott, 1989). An abrupt change then took place beginning with the sandy peat. Shrub birch was replaced by willows, grasses and other herbs. Reduced *Picea* pollen levels suggest deforestation of spruce. This change is dated at 10,600 yr B.P. Environmental change occurred before inundation of the site by inorganic sediments after 10,500 yr B.P.

The laminated, fine grained sediments overlying the peat are interpreted as ponded sediments from a lake or floodplain environment. The restricted area of the deposit and lack of associated fluvial

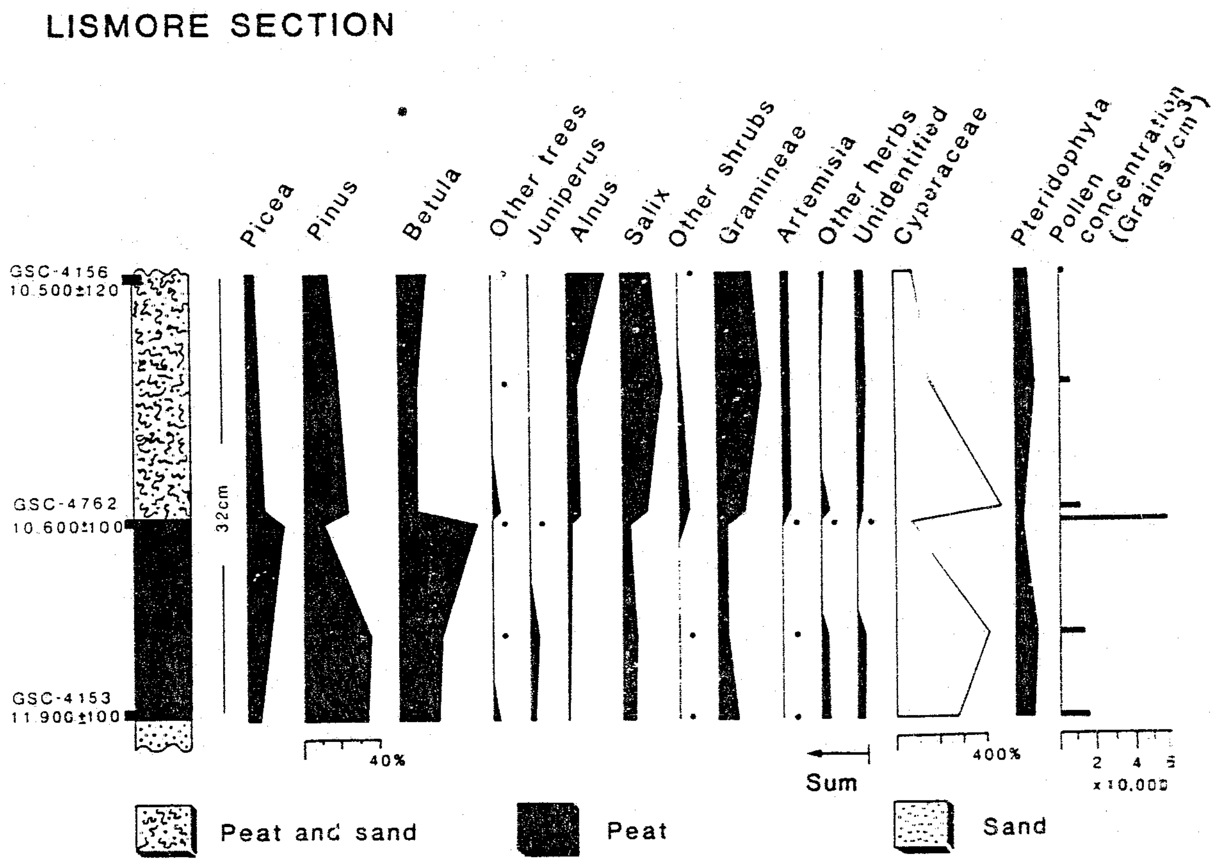

Figure 50. Pollen diagram of the Lismore site.

sedimentary facies favour a lacustrine or glaciolacustrine origin. A relative sea level rise could create ponding conditions either by direct inundation of the land or by re-grading of rivers. This is unlikely, however, because the sea level record in the region shows a relative sea level of -35 m (Quinlan and Beaumont, 1982) at the time of deposition of the laminated sand (ca. 10,500 yr B.P.). Renewed glacial activity, which reformed the lake where the underlying delta had formed, may be an explanation for the genesis of the ponded sediments, but no direct evidence of glaciation is apparent at the site.

STOP 2-5: CAPE GEORGE

Leader: R. R. Stea

Purpose: To view the Wisconsinan sequence overlying an elevated, wave-cut bench believed to represent a high sea level stand during the last interglacial.

Route: Leave Lismore 2:15 PM. Drive 10 km north to junction with Route 337. Follow Route 337 northward 18 km to Cape George Point. Arrive 2:40 PM.

EN ROUTE TO STOP

Wending along this scenic coastal route, we will cross the Hollow Fault at Malignant Cove. This fault separates Lower Paleozoic strata from Hadrynian metavolcanic and metasedimentary rocks (Murphy, 1984). From this point to Cape George we will climb the Antigonish Highlands (Fig. 51). The drift cover thins rapidly as one traverses the highlands, largely underlain by the Precambrian Georgeville Group rocks.

Twenty-four coastal sections were measured from Arisaig on the Northumberland Strait to MacIsaacs Point on the coast of Georges Bay. Eleven representative sections are shown on Figure 51. Common to most of these sections is a sequence of stratified, well sorted sands and gravels (Unit 2) resting on an emerged abrasion platform believed to be a remnant of a Sangamon high sea level stand (Grant, 1980). The height of this rock bench varies from 4-6 m above MSL. At Cape George Point, this bench is developed on mafic rocks. The surface of the bench, exhumed from under sand and till, is scalloped and smoothed. Goldthwait (1924) first described the bench. He noted the similarities between the modern littoral platform and the emerged platform but doubted that it was marine in origin. He stated that discontinuities in the feature and lack of horizontality suggested an alternative mode of origin, possibly faulting. Goldthwait thought that the feature was late- or post-glacial in age rather than pre-glacial.

Lying above the stratified sands in most of the sections is a distinctive, bouldery to gravelly diamicton (Unit 3). The clasts are invariably angular and locally derived. This sediment varies in thickness from 4-20 m and may be a product of mass wasting from adjacent rock slopes during periglacial conditions. It is also possible that it represents a supraglacial flowtill, deposited by advancing glaciers that deposited the overlying till.

Stratigraphically above this diamicton are two till-like diamicton units distinguished primarily by their stone content. The lower unit (Unit 4) is greyish-red with a silty matrix. It is relatively stone free. Robust fragments of Mercenaria sp. are commonly found near the base of this unit. Thick exposures of this unit rest on bedrock surfaces with striations trending 90-130°. Large elongate boulders at the base of some sections with stoss and lee form are oriented parallel to the bedrock striations and have parallel surface striations. This till sheet was probably emplaced by an ice sheet moving to the east and southeast.

The upper unit (Unit 5) is distinguished from the lower shelly till by its abundance of stones and boulders. At some sections it is slightly sandier than the lower till. Till-embedded boulders with striations parallel to their long axes generally trend northeastward. Striations on rock surfaces underneath this till unit along the coast trend from 335 to 045°

STRATIGRAPHY AND CHRONOLOGY

Exposed at this site are six units overlying the wave-cut platform developed on basalt. A brown, massive sand (Unit 2; Fig. 51) rests directly on the bench, which has relict potholes. Overlying the sand is a cobble-gravel diamict (Unit 3; Fig. 50) of local provenance. It attains a thickness of 13 m. A till (Unit 4), bearing shells of Mercenaria sp., overlies the diamict. These tills were emplaced by ice moving southeastward.

A date of >37,000 yr B.P. was obtained on one of the shell fragments (GSC-4048, D. R. Grant, pers. comm., 1984). Amino acid measurements on shell fragments, calibrated with the kinetic model curve of Wehmiller *et al.* (1988), indicate an oxygen isotope Stage 5 age (Fig. 52).

Twelve shell fragments (Mercenaria sp.) from Unit 4 at Cape George were dated using the electron spin resonance method. Radtke *et al.*, (1985) have shown both good and bad results using electron spin resonance dating on different species of molluscs and checked by other dating methods. The bad

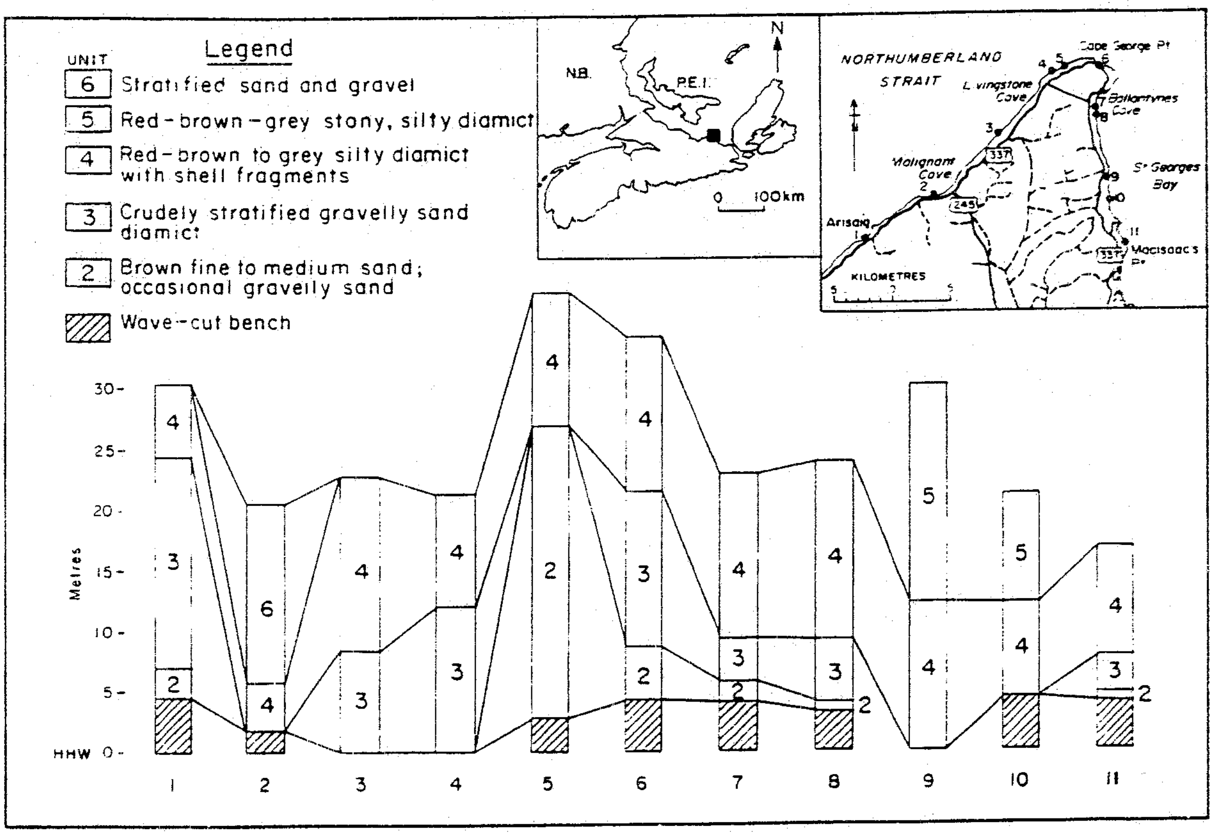

Figure 51. Location and stratigraphy of Quaternary sections along the Northumberland Strait and St. Georges Bay.

results, however, were invariably skewed toward younger ages. The dates range from 99,000-188,000 yr B.P. with a mean value of 136,000 yr B.P. These dates were calculated assuming direct or early uranium accumulation. The older dates are generally on samples with thinner shells and may have had a complicated U-uptake history.

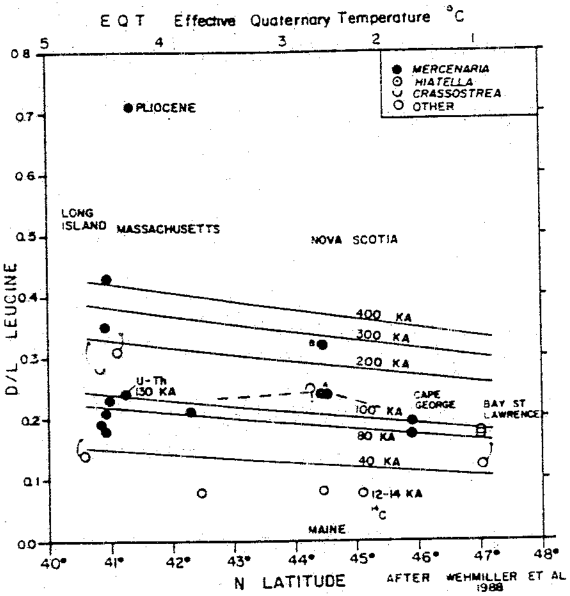

Figure 52. Plot of D/L leucine ratios from fossil invertebrates versus latitude, for New England and the Maritime Provinces. Kinetic age model isochrons after Wehmiller et al., 1988.

STOP 2-6: BALLANTYNES COVE

Leader: R. R. Stea

Purpose: To examine a section of tills formed by ice flows from the northwest and south.

Route: Depart Cape George Point 3:20 PM. Follow Highway 337 5 km. Turn into Ballantynes Cove. Arrive 3:30 PM.

INTRODUCTION

Before examining the section up close, we will stop at the fishing wharf for photographs. From this vantage point the Sangamon platform is clearly visible, cut into differing rock types. The underlying rocks of the section belong to the McAras Brook Formation (basalt, conglomerate) and Wilkie Brook Formation (limestone) (Fig. 53). If time and tide permit we may also visit a site where northward ice flow has carved out large crag and tail features on the near vertical face of a coarse conglomerate.

STRATIGRAPHY AND CHRONOLOGY

The lowland regions that flank the Cobequid Highlands and the Antigonish Highlands, where northward ice flow during phase 3 is pervasive, generally display two to three till units in stratigraphic section. The type section is at Ballantynes Cove south of Cape George where three till units are exposed (Fig. 54). The lower till (Unit 4) is red, matrix-rich and contains shell fragments derived from flow across the Northumberland Strait. It can be correlated with other shell-bearing basal units along the Northumberland Strait (Stea *et al.*, 1987).

One of the distinguishing characteristics of the three till units at Ballantynes Cove is the stone content. The lowest till unit contains less than 10% clasts and exhibits little sample to sample variation. The middle till unit (Unit 5) contains >10% clasts and the till unit above the boulder horizon has >15% clast-sized particles.

The abundances of red-brown coarse sandstone and conglomerate (Lithology 1; Fig. 54) decrease toward the top of the lowest till unit (Unit 4) but grey feldspathic sandstone (Lithology 2) remains fairly constant throughout. Black metasiltstone and shale clasts show a marked increase in abundance at the top of Unit 4 and remain abundant to the top of the section. Green, hard conglomerate and basalt (Lithologies 7 and 8) increase markedly in abundance toward the top of Unit 5 and in Unit 6 near the boulder horizon. Diorite clasts are restricted to Unit 5.

The relative abundance of reddish sandstone and conglomerate clasts, and the presence of well rounded plutonic clasts (Lithology 6; Fig. 54) in Unit 4 are the result of southeastward ice flow across Devonian to Carboniferous conglomerates north of the section (Boucot *et al.*, 1974; Fig. 53). Till fabric at the base of the section shows a strong east-west orientation with a minor southeast mode. Shell fragments in the till and an overall low percentage of clasts confirm a correlation with till on the Northumberland Shore formed by southeastward ice flow.

The marked increase in the proportion of black metasiltstone clasts in Unit 5 and the presence of dioritic clasts indicate erosion of Precambrian Georgeville Group rocks and a change in ice flow from southeastward to northeastward. The northeastward-striking till fabric in Unit 5 is aligned with the change in ice flow direction.

The upper till (Unit 6) is differentiated from Unit 5 by a higher percentage of stones and an increase in greenish, hard conglomerate and basalt clasts. Basalt is found in the McAras Brook Formation, which underlies the section and outcrops in an east to west zone between the Georgeville Stock and Devonian conglomerates (Fig. 53). Their presence indicates a swing in ice flow either to the east or west. The till fabrics measured in Unit 6, however, strike northeast to southwest. If the till fabric is parallel to ice flow, the only way to reconcile this contradictory evidence would be a 180° swing to southwestward ice flow or an unknown source of basalt. Striations along the Northumberland shore near Treen Bluff (Stea and Finck, 1984) attest to late southwestward ice flow although erosional evidence of this ice flow in the Ballantynes Cove area is lacking.

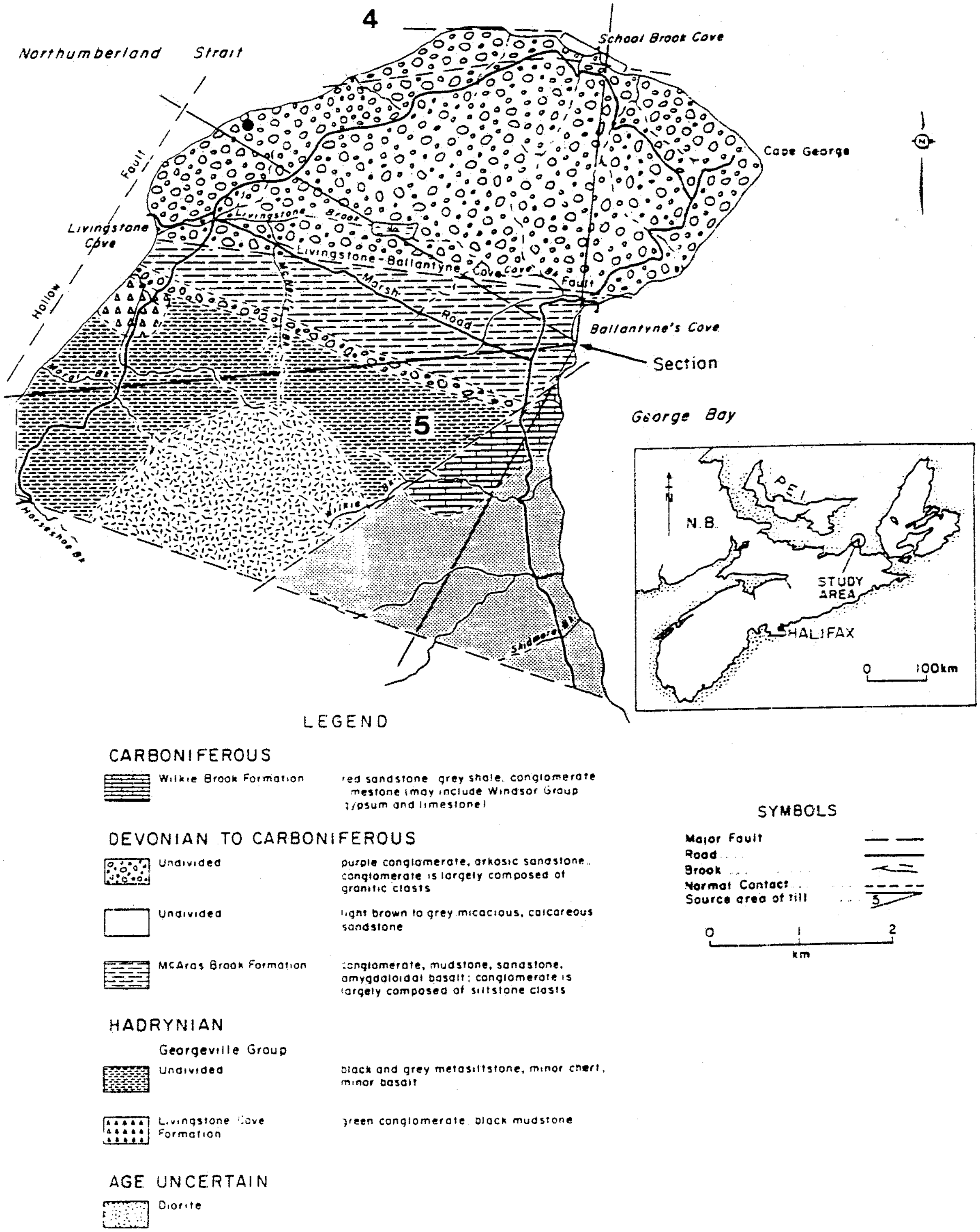

Figure 53. Geology of the Cape George area with source areas of the clasts in till units at Ballantynes Cove.

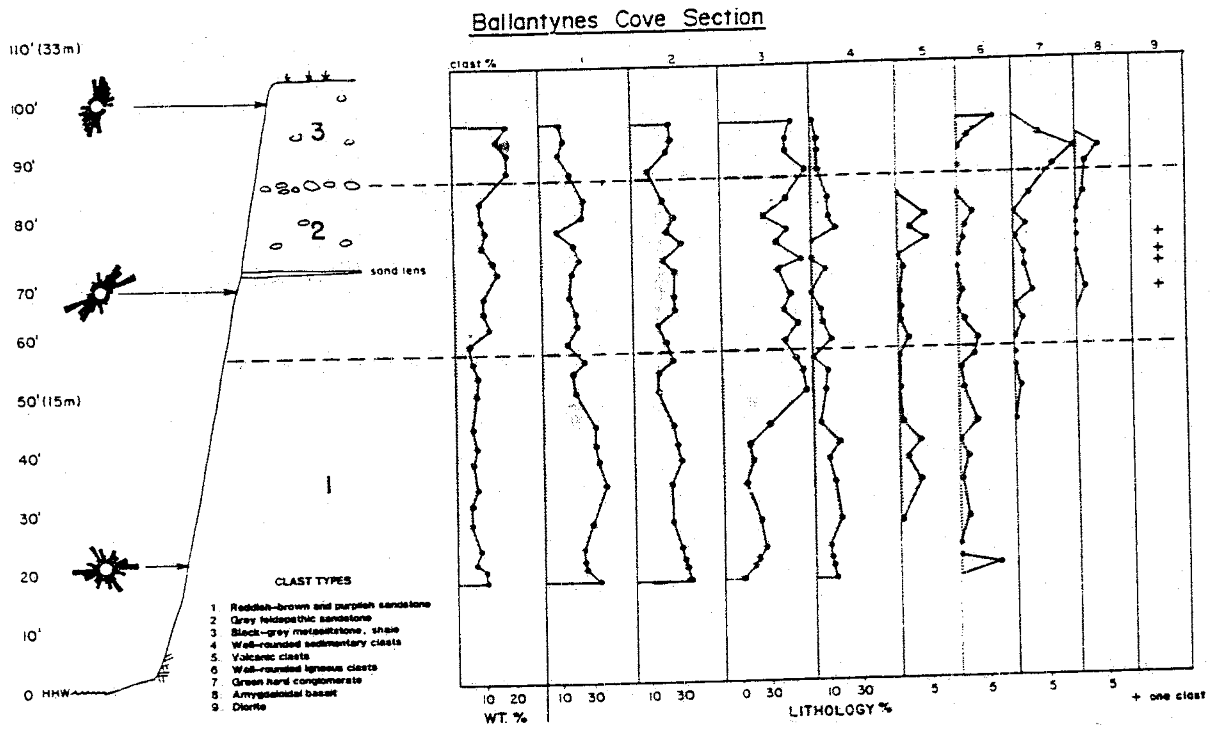

Figure 54. Diagram showing the stratigraphy, fabric and vertical variation in pebble lithology of the Ballantynes Cove section.

STOP 2-7: ADDINGTON FORKS

Leader: R. J. Mott

Purpose: To view a peat bed that was formed during the climatic optimum of the last interglacial.

Route: Leave Ballantynes Cove 4:00 PM. Drive south along Highway 337 32 km through the town of Antigonish to Highway 104. Travel west along the Trans-Canada Highway 2 km to the Brierly Brook road turnoff. Travel 6.6 km along Highway 4. Turn right on road across from Department of Highways depot before bridge. Be careful crossing the main highway. Follow road 2 km. The section is in a road cut on the left past the bridge. Arrive 4:30 PM.

EN ROUTE TO STOP

Driving south along the shore of Georges Bay we pass over the Antigonish Highlands and descend into the flanking Carboniferous basin underlain by limestone, gypsum and sandstone. The white gypsum forms spectacular cliffs along the coast. Crystal Cliffs, a former resort, is located near Antigonish. Among other highlights, Clark Gable once stayed there. Crystal Cliffs was the site of the Massachusetts Institute of Technology's field school in the 1950s. Numerous B.Sc., M.Sc. and Ph.D theses resulted which helped to elucidate the geology of northern Nova Scotia.

The town of Antigonish was largely founded by Highland Scots in the late 1700s. It is the home of the annual Highland Games, a traditional Scottish athletic fest. It is also the home of St. Francis Xavier University, where we will be staying.

After Antigonish we will drive across the large outwash plain flanking the West River (Fig. 55). The outwash consists of a thick succession of channelled gravelly-sand and sand beds. The trend of the channels suggests a northeastward paleoflow. The glacier that formed the outwash receded southward. The outwash plain merges with kame terraces in the Ohio River valley to the south (Fig. 55). A mastodon molar is purported to have come from one of these gravel pits (W. Shaw, pers. comm., 1986).

INTRODUCTION

The site was originally discovered by R. MacNeill when the roadside ditch was excavated for drainage purposes prior to 1969. V. K. Prest re-examined the site and collected samples for pollen analysis and dating. When preliminary pollen analysis revealed spectra suggesting interglacial conditions, Mott and Prest re-excavated the site in 1981 for more detailed study. The steep treed slope above the deposit and the adjacent road precluded extensive excavation into the overlying and underlying glacial deposits.

STRATIGRAPHY AND CHRONOLOGY

About 80 cm of silt with organic seams, organic silt and highly compacted peat make up a nonglacial package of sediments sandwiched between a basal red-brown till and an overlying red till (Fig. 56).

Wood from the site collected by MacNeill (1969) produced a finite date of 33,700 + 2300/-1800 yr B.P. (I-3236). However, Juniper wood collected by Prest gave an age of >42,000 yr B.P. (GSC-1598; Lowdon and Blake, 1973). A third age determination on spruce/tamarack wood gave a date of 36,000 ± 520 yr B.P. Roots from trees on the densely treed slope covering the deposit may be a source of contamination that produced the finite dates.

PALYNOLOGY

The abbreviated pollen diagram (Fig. 56) has been divided into 5 pollen assemblage zones (Mott and Grant, 1985). The basal Zone 1 is dominated by Polypodiaceae and *Osmunda* (fern) spores and tree pollen of *Quercus* (oak) and *Betula* (birch). *Quercus* dominates in Zone 2 along with significant amounts of *Carpinus/Ostrya* (blue beech/ironwood) pollen. *Quercus* decreases and *Pinus* (pine), *Carya* (hickory) and other thermophilous hardwood taxa such as *Fagus* (beech), *Tilia* (basswood) and *Ulmus* (elm) increase in Zone 3. *Cephalanthus* (buttonbush) type and Gramineae along with other herbaceous taxa probably reflect the local flora. *Osmunda* spores attain a second maximum in this zone as well. Decline of *Pinus* pollen and thermophilous hardwood taxa in Zone 4 accompanies in *Abies* (fir), *Picea* (spruce) and *Alnus* (alder). In Zone 5 extremely large *Alnus* values supplant most other taxa.

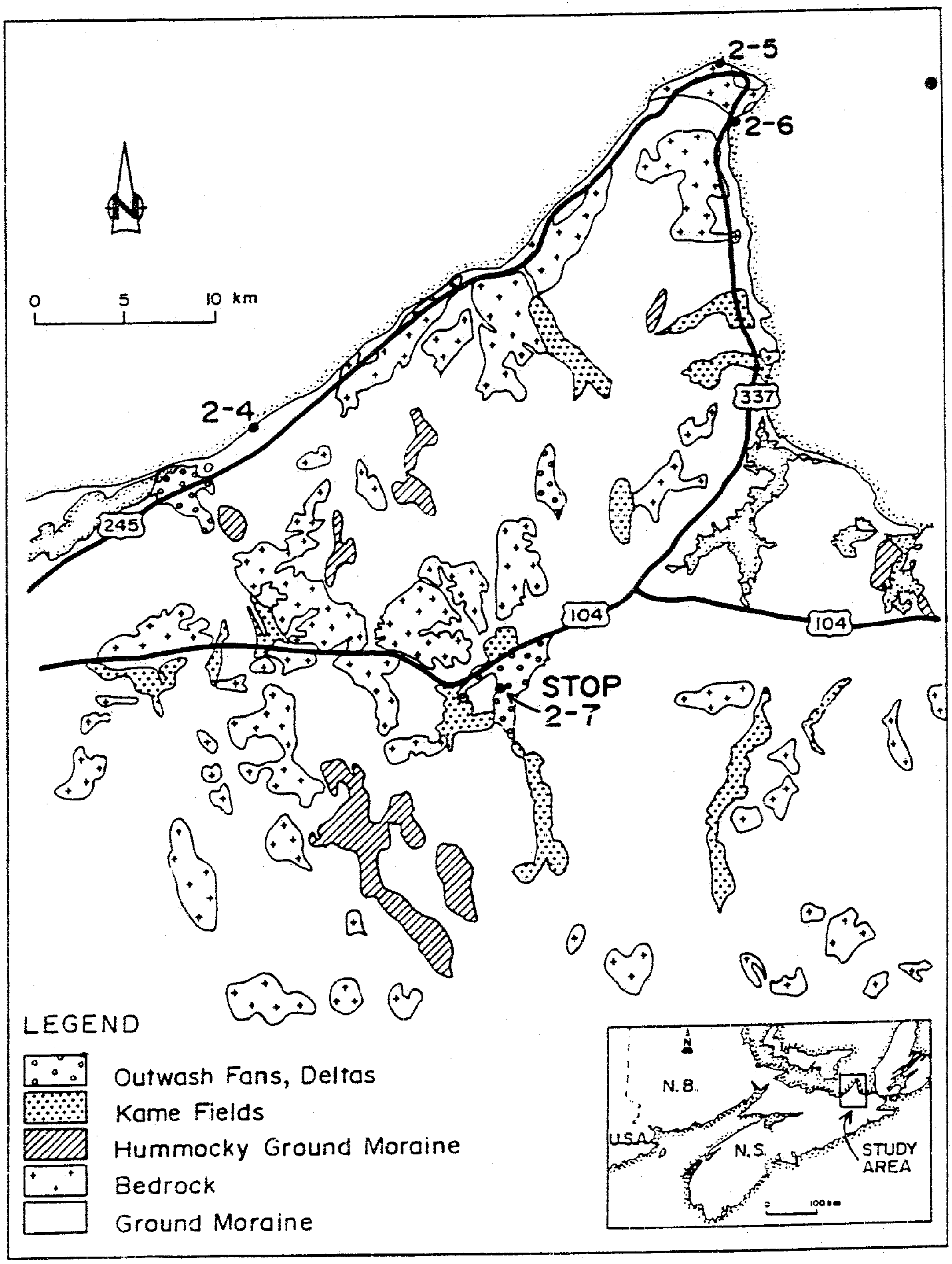

Figure 55. Simplified surficial geology of the Antigonish area and location of Stop 2-7.

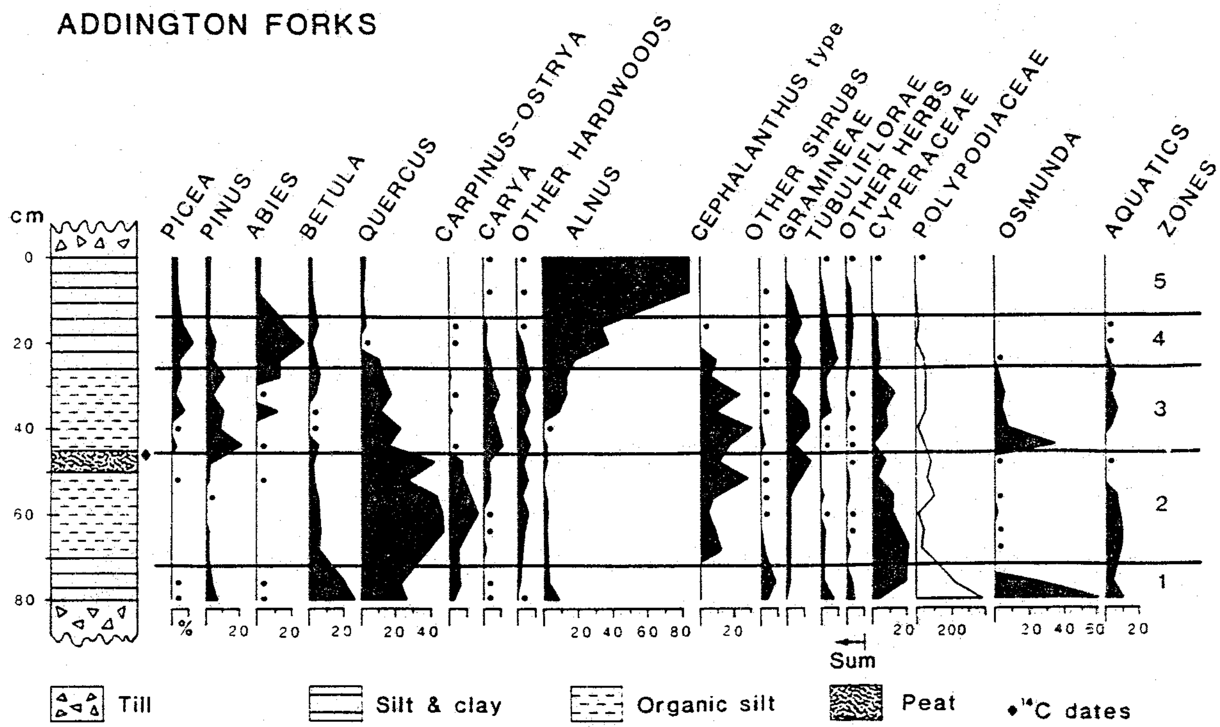

Figure 56. Pollen diagram and stratigraphy of the Addington Forks site.

INTERPRETATION

Despite the two finite dates obtained, the pollen spectra indicate that the organic sediments were deposited under interglacial conditions. Therefore, the dates are considered spurious. The changing pollen spectra indicate that the waning phase of a warm climatic interval is represented. Forests with birch, oak, hickory and other hardwoods were replaced by coniferous forests of balsam fir and spruce. Alder domination mayindicate further cooling to sub-arctic conditions.

This deposit likely relates to the cooling phase of substage 5e with conditions that were warmer than present declining to cool boreal to sub-arctic conditions. If this interpretation is correct, the till at the base of the section would be Illinoisan. The overlying till probably relates to ice flow phases 1 or 2, but no work has yet been done on the sediments.

DAY 3

STOP 3-1: MOOSE POINT

Leaders: R. R. Stea and R. J. Mott

Purpose: To see an interglacial/interstadial peat bed that rests on the emerged wave-cut bench.

Route: Leave Antigonish 8:30 AM. Travel eastward along the Trans Canada Highway 28 km to the junction with Highway 16 at Monastery. Drive south 20 km. Turn left onto Highway 344 and drive 2 km. Turn onto dirt road and follow to Chedabucto Bay shore, 6 km. Arrive 9:20 AM.

EN ROUTE TO STOP

The village of Monastery is the site of a monastery founded by trappist monks in 1825. It is now run by the Augustinian order.

We will pass over rock barrens on the way to Guysborough. These rock surfaces are inscribed with a pervasive set of striations and grooves indicating ice flow to the northwest. Many of these striated surfaces are faulted and offset. Neotectonic activity is suspected, but the age and severity of the deformation are not known.

INTRODUCTION

Sections along the north shore of Chedabucto Bay reveal seven lithostratigraphic units lying above the emerged rock bench (Fig. 57). Along this shore, however, there is a peat bed with wood fragments that lies directly on the bench and predates the deposition of tills. The exposures along this coast are generally in drumlins or drumlinoids whose stoss and lee sides are not well developed.

STRATIGRAPHY AND CHRONOLOGY

The oldest surficial unit is a silt containing more or less abundant organics and compressed peat layers with large wood fragments, which lies directly on regolith from an abrasion platform 4-6 m above tide level. Spruce wood from the compressed peat produced a radiocarbon date of >49,000 yr B.P. (GSC-4419 HP).

Unit II, a silty diamicton, is the thickest unit in the area. It forms the core of many of the drumlins which are the sites of cliff erosion. The unit is interpreted as a till because of its massive and indurated nature, erratic content, and the abundance of stoss-lee boulders parallel to underlying bedrock striations (Fig. 57). The flow associated with this unit is eastward and southeastward parallel to ice flow phase 1 (Fig. 4). The presence of distinctive appinite (mafic pegmatite) boulders in the unit implies an ice flow either to the east or southeast from two of the only known stocks (J. B. Murphy, pers. comm., 1990). Ice flow across the stock north of Antigonish is consistent with the modal trend of the long axes of the till-embedded boulders (115-125°). It appears from underlying bedrock striations that the initial flow was almost due east (Phase 1a; Fig. 4). Shell fragments in the till are presumably derived from Georges Bay.

Units III and IV are waterlain units, which are relatively rare in these drumlin sections. Section 153, however, reveals fairly continuous, rhythmically bedded gravelly-sand and silty-sand beds which separate two massive diamicton units.

Unit V is a diamicton that is distinguished from unit II by greater stone content. This unit contains stoss-lee boulders that indicate a northwestward ice flow, parallel the ice flow inferred by regional striation mapping (ice flow phase 3; Fig. 4).

At Stop 3-2 we will examine deposits from a younger event than the sediments in these sections.

PALYNOLOGY

Preliminary pollen analysis of the organic unit shows pollen spectra near the base with abundant *Picea* (spruce) and *Pinus* (mainly *P. banksiana/resinosa* type) (jack/red pine) pollen and small values for other taxa (Fig. 58). As *Picea* declines and *Pinus* increases slightly above the base of the profile, Polypodiaceae (fern) spores become extremely abundant. *Pinus* reaches a maximum in mid profile and then declines as *Picea* continues to

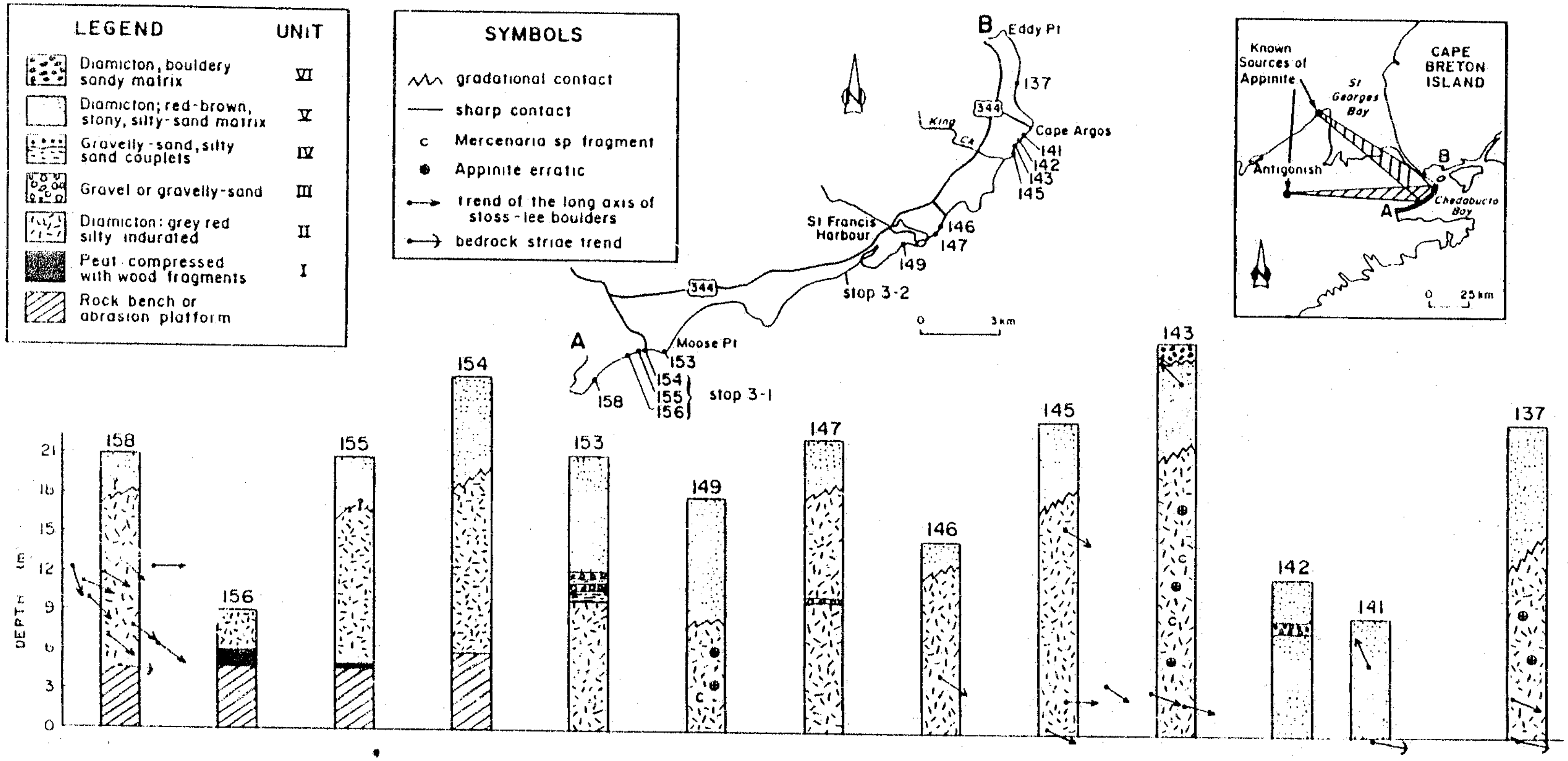

Figure 57. Location and stratigraphy of Quaternary sections along the Chedabucto Bay coast.

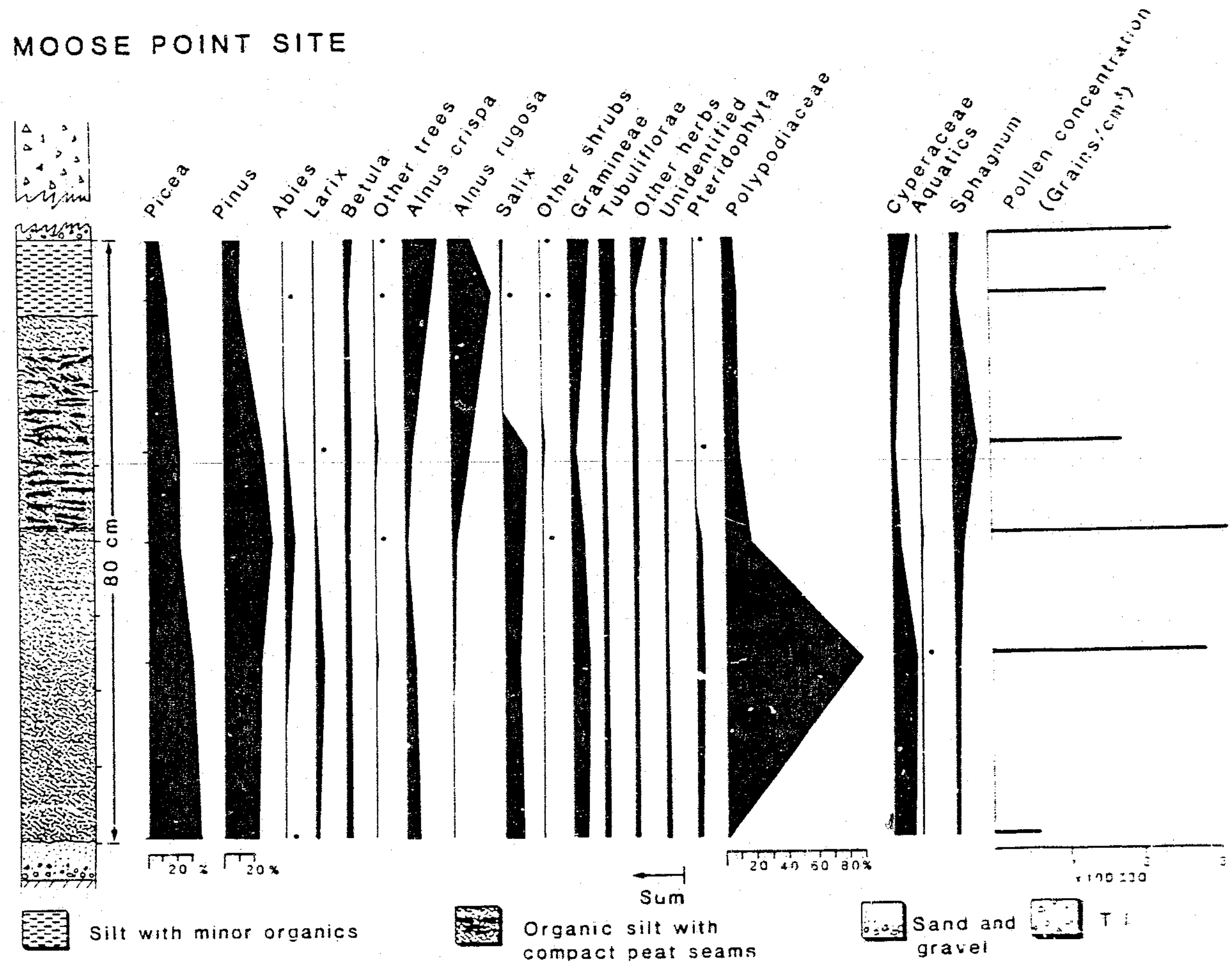

Figure 58. Pollen diagram of the Moose Point site (Stop 3-1).

decline and Polypodiaceae returns to low values. At the top of the profile, *Alnus* (alder) species dominate, *Picea* and *Pinus* values are low, and herbaceous taxa values increase somewhat.

INTERPRETATION

Northern boreal type coniferous forest of spruce and some jack pine occupied the area at time of deposition. Ferns were abundant at least locally for part of the time. Willow, grasses and herbs inhabited open areas. Increasing alder and herbs and declining spruce and pine probably indicate a cooling trend as postulated for the upper part of the A..ngton Forks profile.

The waning phase of a warm interval such as oxygen isotope substage 5e is possible, but more likely this deposit relates to a younger, cooler interval such as substage 5a. The organic intervals position on the raised marine bench requires that its time of deposition postdates the climatic optimum of the last interglacial interval. The overlying till, which is related to phase . ice flow, probably favours the substage 5a possibility.

STOP 3-2: COLLINS POND

Leaders: R. R. Stea and R. J. Mott

Purpose: To look at the best exposed section revealing a glacial diamicton overlying late-glacial peat.

Route: Leave Stop 3-1 10:30 AM. Drive back along dirt road to Highway 344. Proceed eastward 10 km. Pull into small grocery on the right. Walk down path to beach. Arrive 10:40 AM.

INTRODUCTION

The Collins Pond site is located 25 km northeast of Guysborough, along the north shore of Chedabucto Bay (Fig. 59). Till is the predominant surficial deposit in the area. Drumlins trending southeastward occur along the actively eroding shore and sections up to 25 m thick are exposed. Buried peat beds were found in two sections (A and B; Fig. 60) at this site, both on the eastern flanks of the drumlins.

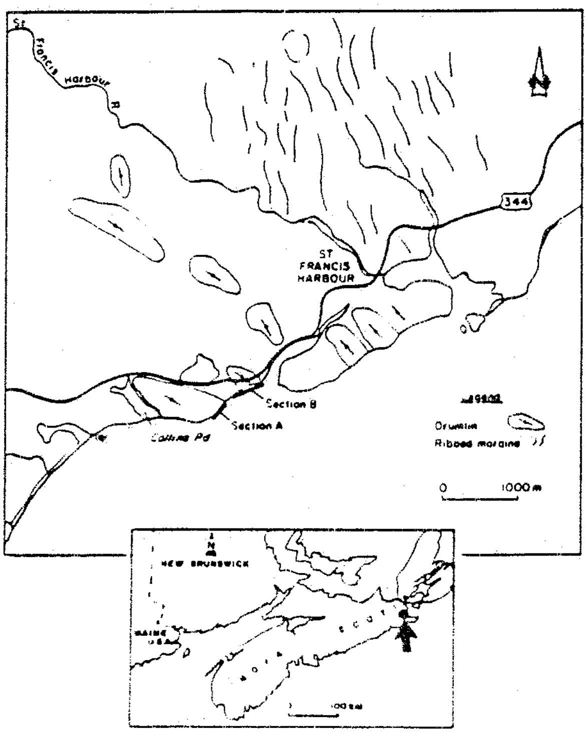

Figure 59. Location and surficial geology of the Collins Pond site (Stop 3-2)

STRATIGRAPHY AND CHRONOLOGY

The drumlins primarily consist of a greyish-red, matrix-supported, silty diamicton (Unit I; Fig. 60) interpreted as till. Clasts in the unit are derived from the Antigonish Highlands terrane to the northwest.

Overlying the bouldery, washed surface of Unit I in a gully between drumlins in the eastern part of section A is a sequence of 10 cm of organic silt and clay, 40 cm of fibrous peat with small twigs, and 10 cm of grey clay with thin organic layers (Fig. 60). The organic bed thins abruptly against the steep flank of the drumlin to the west. This intermittent, thin horizon of matted peat can be traced up the Unit I surface until it pinches out near the drumlin crest. A brown silty sand occurs at the same stratigraphic level and is sometimes intercalated with the organic seam. Thin seams of this brown silty sand as well as seams of organics and organic clay occur within Unit II, the stony diamicton that overlies the organic horizon and Unit I on the northeastern flank of the drumlin. Fabric analyses at three spots within Unit II (Fig. 60) suggest that this diamicton is glacial in origin and was formed by westward-flowing ice.

On the eastern flank of the adjacent drumlin two additional units are exposed overlying Unit I (section B, Fig. 60). Unit III is a diamicton containing thin beds of sand and gravel and relatively few stones. Deformed bodies of organic silt have also been incorporated into this diamicton. The fabric of Unit III is oriented toward the southeast, similar to that of Unit I and parallel to the slope of the drumlin flank at this point (Fig. 60). A thin, gravelly diamicton (Unit IV) is found farther up slope above Unit I (section B, Fig. 60).

The basal 2 cm of the thick peat bed of section A produced a date of 12,700 ± 130 yr B.P. (GSC-4474) and the top 2 cm interval was dated at 11,800 ± 100 yr B.P. (GSC-4367). The thin organic seam within Unit II was dated at 10,900 ± 100 yr B.P. (GSC-4475).

PALYNOLOGY

The pollen profile of the thick peat bed of section A is shown in Figure 61. Cyperaceae increases to a maximum in the middle of the peat along with

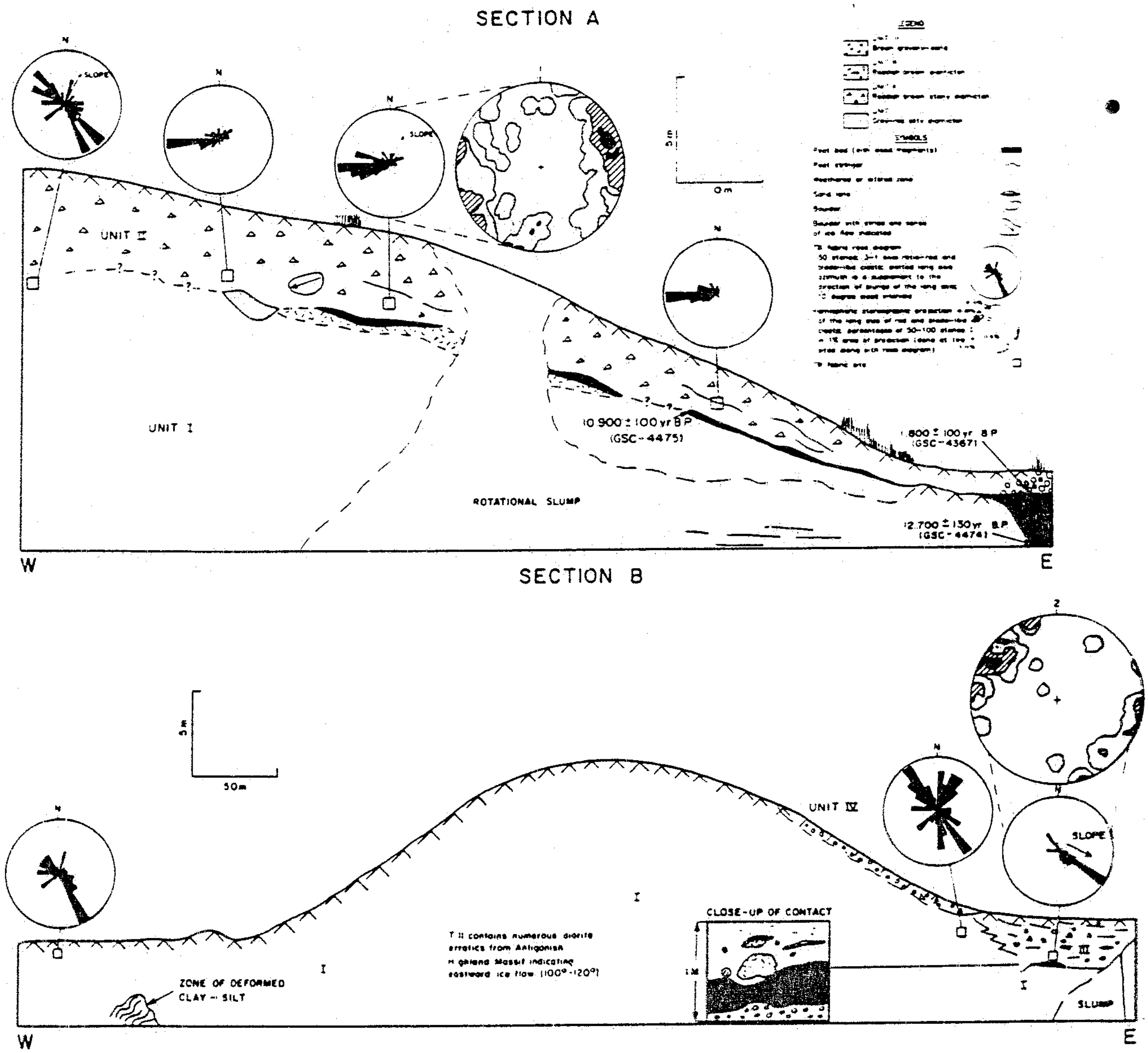

Figure 60. Stratigraphy of the Collins Pond site.

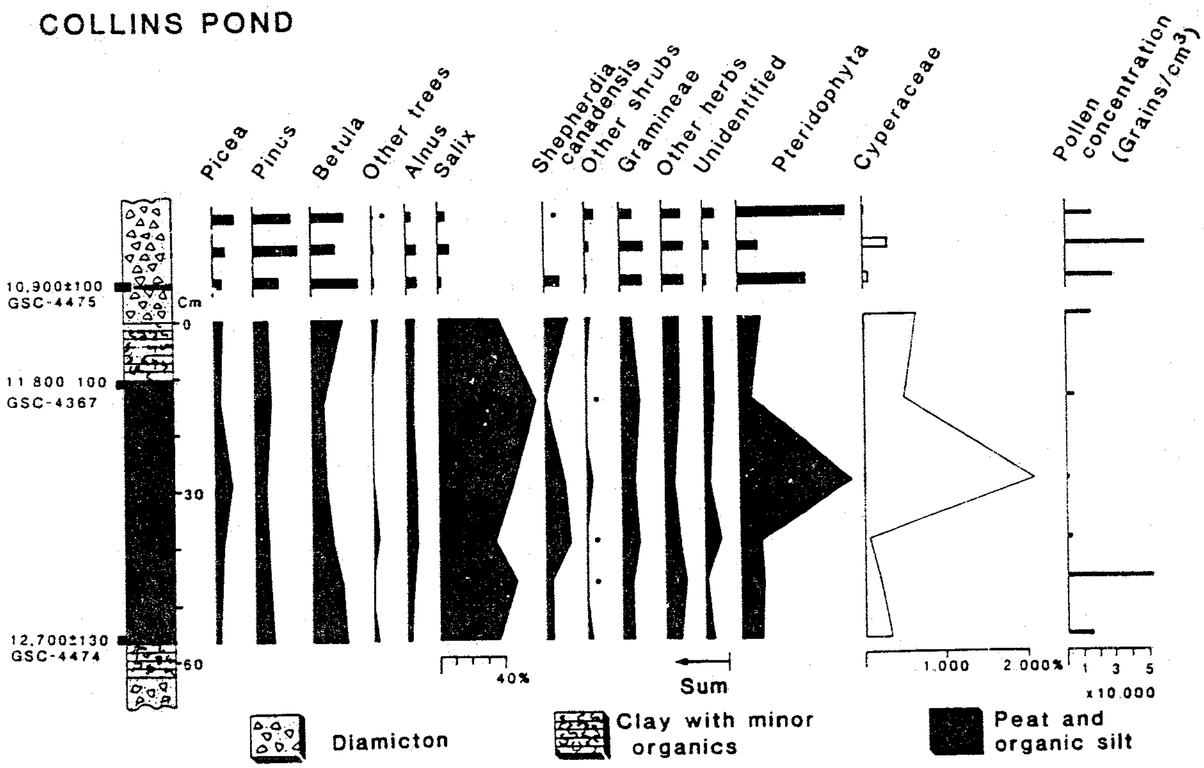

Figure 61. Palynology of the Collins Pond site.

Pteridophyta (fern and fern allies) spores. At this level the peak is almost completely *Equisetum* (horsetail). Near the top of the peat, *Salix* (willow) values exceed 50%, and Cyperaceae values are much lower. *Betula* percentages increase and *Salix* declines in the overlying organic clay.

The pollen assemblage in the narrow organic bed exposed beneath Unit II (Fig. 60) differs from the thick peat profile because of lower *Salix* and Cyperaceae values and higher *Betula*, Pteridophyta, and Polypodiaceae (fern) values. *Betula* values exceed those of *Salix* in both the displaced organic sediments within the Unit III diamicton of section B and the thin peat horizon of section A, but these units differ because of higher percentages of Cyperaceae in Unit III.

INTERPRETATION

Two major, early ice flow events are recorded by striations, erratic dispersal, till fabric and drumlin orientation in the Guysborough area. The first major ice flow was southeastward and southward, followed by northward flow from a divide on the Southern Uplands (Fig. 4). Ice then retreated prior to 12,700 yr B.P. allowing vegetation to migrate into the area. Shallow ponds and sedge meadows occupied low wet areas. Tundra-like shrub communities of willow and possibly some birch surrounded the wetlands and occupied the lower slopes. Sparse herbaceous vegetation occupied open areas on the slopes and dominated the uplands. No trees were present. These conditions remained unchanged, except for increasing shrub birch representation at the expense of willow, until 10,900 yr B.P. An abrupt change then occurred and diamictons were deposited. Diamictons found overlying the peat beds on drumlin slopes could either form as earth or debris flows or by glacier activity.

Earth and debris flows are poorly sorted, locally derived, and have parallel clast orientation, similar to till (Van Steijn, 1987). Long axis orientations in laboratory-simulated flow deposits tend to be highly variable (Van Steijn, 1988). Strong transverse fabrics are also possible, especially at the toe of a flow (Boulton, 1971). The lack of morphological evidence for a flow, such as a toe bulge, or parallel strips at the base of section A, mitigates against a flow hypothesis. Quinty and Filion (1989) describe marked asymmetry in drumlin profiles caused by sheet gelifluction in subarctic climates. This asymmetry is not in evidence in these drumlins. The younger radiocarbon date of 10,900 yr B.P. for the intercalated peat seam upslope from the main peat bed provides a compelling argument against a gelifluction or solifluction origin for the diamicton. A solifluction lobe or earth flow would have tended to remove this peat seam and deposit thick sediment over the peat bed in the gully. In contrast, the gully peat sequence seems to have been truncated and possibly smeared upslope beneath a till layer that increases in thickness upslope.

On the other hand, Unit II (section A, Fig. 60) fulfils most of the criteria for till outlined by Dreimanis and Schluchter (1985). These include high bulk density, parallel clast orientation, erosional basal contacts, multimodal particle distribution and striated clasts. The grain size distribution of the diamicton is similar to those for selected Nova Scotia tills (Fig. 42). A striated bullet-boulder believed to be characteristic of lodgment till (Boulton, 1978; Kruger, 1984) was found embedded in the unit. The fabric and boulder orientations imply a westward ice flow. Late ice in Chedabucto Bay (Fig. 16) or on the outer banks would explain an onshore, westward ice flow. The linear ridges near the section (Fig. 59) are oriented northwest-southeast and suggest an ice margin of this orientation (Goldthwait, 1987). An esker east of Guysborough (Fig. 39) and several striation sites in the region suggest a late, onshore, southwestward ice flow. This may be a small remnant of the ice divide that formed in Chedabucto Bay during the Wisconsin Stage (Stea *et al.*, 1987; Grant, 1988).

Unit III has a till-like fabric, but the main long axis trend is perpendicular to those in section A.

Unit III may be a melt-out till (Shaw, 1982). Melt-out tills in modern glacial environments generally display fabric orientation parallel to ice flow (Lawson, 1979), but transverse fabrics have also been described (Mills, 1977). This unit has either a direct glacial origin or has been modified by solifluction. Evidence for a late, southeastward ice flow was found in the Roman Valley (Fig. 39) and in the Canso Strait area (Grant, 1988). Several small remnant ice caps could have been operating independently during the last stages of deglaciation.

A sediment core from Manassette Lake, less than 1 km inland to the north of the coastal site, was collected to ascertain if the lake site provided a similar scenario to that deduced for the coastal site. The core revealed 300 cm of algal gyttja and clayey gyttja overlying laminated clay to a depth of at least 2 m below the base of the organic sediment. Within the clay about 85 cm below the contact is a 10 cm thick seam of slightly organic clay with minor organic detrital fragments. This organic horizon contains too little organic matter for conventional radiocarbon dating, but an accelerator radiocarbon date is possible. A date on the basal organic sediment (297-300 cm) produced an age of 9910 ± 100 yr B.P. (GSC-4964). A sample at 255-258 cm depth gave a date of 8630 ± 90 yr B.P. (GSC-4965). Preliminary pollen analysis of the base of the organic sediments gives pollen spectra dominated by *Betula* (tree birch), *Picea* (spruce), minor hardwood and shrub taxa and minor amounts of open ground herbaceous taxa. The spectrum from the organic horizon within the clay is dominated by Cyperaceae (sedge) pollen, smaller amounts of *Picea*, *Pinus* (pine) and *Betula* (mainly shrub birch), some shrub taxa and variety of herbaceous taxa. More work will be required before the significance of this sequence is clear, but the results seem to indicate that the coastal buried organic site predates the lake site, and that the climatic oscillation is not recorded in the latter. This is what would be expected if late ice covered the area during late-glacial time.

REFERENCES

Bailey, L. W. 1898: Report on the geology of southwestern Nova Scotia; Geological Survey of Canada, Annual Report, v. 9, pt. M, p. 30-60.

Borns, H. W. 1965: Late glacial ice-wedge casts in northern Nova Scotia, Canada; Science, v. 148, no. 3674, p. 1223-1226.

Boucot, A. J., Dewey, J. F., Dineley, D. L., Fletcher, R., Fyson, W. K., Griffin, J. G., Hickox, C. F., McKerrow, W. S. and Ziegler, A. M., 1974: Geology of the Arisaig area, Antigonish County, Nova Scotia; Geological Society of America, Special Paper 139, 191 p.

Boulton, G. F. 1971: Till genesis and fabric in Svalbard, Spitsbergen; *in* Till, a Symposium, ed. R. P. Goldthwait; Ohio State University Press, p. 41-71.

Boulton, G. F. 1978: Boulder shapes and grain-size distributions of debris as indicators of transport paths through a glacier and till genesis; Sedimentology, v. 25, p. 773-799.

Boulton, G. F. and Jones, A. S. 1979: Stability of temperate ice caps and ice sheets resting on beds of deformable sediments; Journal of Glaciology, v. 24, no. 90, p. 29-43.

Causse, C. and Hillaire-Marcel, C. 1986: Géochimie des familles U et Th dans la matière organique fossile des dépôtsinterglaciaires et interstadiaires de l'est et du nord du Canada; potentiel radiochronologique; *dans* Recherches en Cours, Partie B; Commission géologique du Canada, Etude 86-1B, p. 11-18.

Chalmers, R. 1895: Report on the surface geology of eastern New Brunswick, northwestern Nova Scotia and a portion of Prince Edward Island; Geological Survey of Canada; Annual Report 1894, v. 1, no. vii, pt. m, 144 p.

Clark, P. U., and Hansel, A. K. 1989: Clast ploughing, lodgement and glacial sliding over a soft glacier bed; Boreas, v. 18, p. 201-207.

Davis, W. M. 1922: Peneplains and the geographical cycle; Geological Society of America Bulletin, v. 23, p. 587-598.

Denton, G. H. and Hughes, T. J. 1981: The Last Great Ice Sheets; John Wiley and Sons, Inc., Toronto, Ontario, 484 p.

Donohoe, H. V., Jr. and Wallace, P. I. 1982: Geological maps of the Cobequid Highlands-Colchester, Cumberland and Pictou Counties, Nova Scotia; Nova Scotia Department of Mines and Energy, Maps 82-6 to 82-9.

Dreimanis, A. and Schluchter, C. 1985: Field criteria for the recognition of till or tillite; Palaeogeography, Palaeoclimatology and Palaeoecology, v. 51, p. 7-14.

Flint, R. F. 1951: Highland centres of former glacial outflow in northeastern North America; Geological Society of America Bulletin, v. 62, p. 21-38.

Flint, R. F. 1971: Glacial and Quaternary Geology; John Wiley and Sons, 892 p.

Fulton, R. J. 1984: Summary: Quaternary stratigraphy of Canada; *in* Quaternary Stratigraphy of Canada-A Canadian contribution to IGCP project 24, R. J. Fulton, ed.; Geological Survey of Canada, Paper 84-10, p. 1-5.

Giles, P. G. 1981: The Windsor Group of the Mahone Bay area, Nova Scotia; Nova Scotia Department of Mines and Energy, Paper 81-3, 51 p.

Gipp, M. R., and Piper, D. J. W. 1989: Chronology of the Late Wisconsinan glaciation, Emerald Basin, Scotian Shelf; Canadian Journal of Earth Sciences, v. 26, p. 333-335.

Goldthwait, J. W. 1924: Physiography of Nova Scotia; Geological Survey of Canada, Memoir 140, 179 p.

Goldthwait, R. P. 1987: Analytical classification of glacigene landforms; *in* Genetic Classification of Glacigene Deposits, eds. Goldthwait, R. P. and Matsch, C. L.; final report of the commission on genesis and lithology of glacial Quaternary deposits of the International Union for Quaternary Research (INQUA), p. 267-278.

Grant, D. R. 1963: Pebble lithology of the tills of southeast Nova Scotia; unpublished M.Sc. thesis, Dalhousie University, Halifax, Nova Scotia, 235 p.

Grant, D. R. 1975: Glacial style and the Quaternary stratigraphic record in the Atlantic

Provinces, Canada; *in* Report of Activities, Part B; Geological Survey of Canada. Paper 75-1B, p. 109-110.

Grant, D. R. 1977: Glacial style and ice limits, the Quaternary stratigraphic record, and changes of land and ocean level in the Atlantic Provinces, Canada; Geographie Physique et Quaternaire, v. 31, no. 3-4, p. 247-260.

Grant, D. R. 1980: Quaternary sea-level change in Atlantic Canada as an indication of crustal delevelling; *in* Earth Rheology, Isostasy and Eustasy, ed. N. A. Morner; John Wiley and Sons, London, p. 201-214.

Grant, D. R. 1988: Surficial geology, Cape Breton Island, Nova Scotia; Geological Survey of Canada, Map 1631A, scale 1:125 000.

Grant, D. R. and King, L. H. 1984: A stratigraphic framework for the Quaternary history of the Atlantic Provinces; *in* Quaternary Stratigraphy of Canada - A Canadian Contribution to IGCP Project 24, ed. R. J. Fulton; Geological Survey of Canada, Paper 84-10, p. 173-191.

Graves, R. M. and Finck, P. W. 1988: The provenance of tills overlying the eastern part of the South Mountain Batholith, Nova Scotia; Maritime Sediments and Atlantic Geology, v. 24, p. 61-70.

Grist, A. M., Reynolds, P. H. and Zentilli, M. 1990: Provenance and thermal history of detrital sandstones of the Scotian Basin, offshore Nova Scotia, using apatite fission track and ^{40}AR/^{39}AR methods; *in* Atlantic Geoscience Society, Program with Abstacts, p. 14.

Hack, J. T. 1960: Interpretation of erosional topography in humid temperate regions; American Journal of Science, v. 258-A, p. 80-97.

Hacquebard, P. A. 1984: Composition, rank and depth of burial of two Nova Scotia lignite deposits; *in* Current Research, Part A; Geological Survey of Canada, Paper 84-1A, p. 11-15.

Hadden, K. A. 1975: A pollen diagram from a postglacial peat bog in Hants County, Nova Scotia; Canadian Journal of Botany, v. 53, p. 39-47.

Hennigar, T. W. 1970: A groundwater report on the Nova Scotia Department of Lands and Forests complex at Shubenacadie; Nova Scotia Department of the Environment, unpublished report, 14 p.

Hicock, S. R. and Dreimanis, A. 1985: Glaciotectonic structures as useful ice-movement indicators in glacial deposits: four Canadian case studies; Canadian Journal of Earth Sciences, v. 22, p. 339-346.

Hickox, C. F., Jr. 1962: Pleistocene geology of the central Annapolis Valley, Nova Scotia; Nova Scotia Department of Mines, Memoir 5, 36 p.

Hillaire-Marcel, C. and de Vernal, A. 1989: Isotopic and palynological records of the late Pleistocene in eastern Canada and adjacent ocean basins; Géographie Physique et Quaternaire, v. 43, p. 263-290.

Honeyman, D. 1876: On the varieties of transported boulders from the Nova Scotia shore; Proceedings of the American Philosophical Society, v. 16, p. 237.

Hughes, O. L. 1957: Surficial geology of the Shubenacadie map area; Nova Scotia; Geological Survey of Canada, Paper 56-3, 10 p.

Jetté, H. and Mott, R. J. 1989: Palynostratigraphie du tardiglaciaire et de l'Holocène de la région du lac Chance Harbour, Nouvelle-Ecosse; Géographie Physique et Quaternaire, v. 43, no. 1, p. 27-39.

King, L. H. 1969: Submarine end moraines and associated deposits on the Scotian Shelf; Geological Society of America Bulletin, v. 80, p. 83-96.

King, L. H. 1972: Relation of plate tectonics to the geomorphic evolution of the Canadian Atlantic Provinces; Geological Society of America Bulletin, v. 83 p. 3083-3090.

King, L. H. and Fader, G. B. 1986: Wisconsinan glaciation of the Continental Shelf of southeast Atlantic Canada; Geological Survey of Canada, Bulletin 363, 72 p.

Kruger, J. 1984: Clasts with stoss-lee form in lodgement tills: a discussion; Journal of Glaciology, v. 30, p. 241-243.

Lawson, D. E. 1979: A comparison of the pebble orientations in ice and deposits of the Matanuska glacier, Alaska; Journal of Geology, v. 87, p. 629-645.

Lin, C. L. 1970: Hydrogeology of the Musquodoboit River valley, Nova Scotia; Nova Scotia Department of Mines, Report 70-3, 60 p.

Livingstone, D. A. 1968: Some interstadial and postglacial pollen diagrams from Eastern Canada; Ecological Monographs, v. 38, p. 87-125.

MacDonald, G. F. 1968: Debert, a Palaeo-Indian Site in Central Nova Scotia; Anthropology Papers, National Museum of Canada, no. 16, 207 p.

MacNeil, R. H. 1951: Pleistocene geology of the Wolfville area, Nova Scotia; M.Sc. thesis, Acadia University, Wolfville, Nova Scotia; Nova Scotia Department of Mines and Energy, Thesis 241, 59 p.

MacNeill, R. H. 1969: Some dates relating to the dating of the last major ice sheet in Nova Scotia; Maritime Sediments, v. 5, no. 1, p. 3.

MacNeil, R. H. and Purdy, C. A. 1951: A local glacier in the Annapolis-Cornwallis Valley; Proceedings of the Nova Scotian Institute of Science, v. 23, pt. 1, p. 111.

Martinson, D. G., Pisias, N. G., Hays, J. D., Imbrie, J., Moore, T. C., Jr. and Shackleton, N. J. 1987: Age dating and the orbital theory of the ice ages: Development of a high resolution 0 to 300,000 year chronostratigraphy; Quaternary Research, v. 27, p. 1-29.

Miller, A. A. L., Mudie, P. and Scott, D. B. 1982: Holocene history of Bedford Basin, Nova Scotia; Foraminifera, Dinoflagellate and pollen records; Canadian Journal of Earth Sciences, v. 19, no. 12, p. 2342-2367.

Mills, H. H. 1977: Differentation of glacier environments by sediment characteristics: Athabasca glacier, Alberta, Canada; Journal of Sedimentary Petrology, v. 47, no. 2, p. 728-737.

Mott, R. J., Anderson, T. W. and Matthews, J. V., Jr. 1982: Pollen and macrofossil study of an interglacial deposit in Nova Scotia; *in* Special Edition Devoted to Papers from the Eleventh INQUA Congress, ed. P. Richard; Geographie Physique et Quaternaire, v. 36, no. 1-2, p. 197-208.

Mott, R. J. and Grant, D. R. G. 1985. Pre-Late Wisconsinan paleoenvironments in Atlantic Canada; Geographie Physique at Quaternaire, v. 39, p. 239-254.

Mott, R. J., Grant, D. R. G., Stea, R. R. and Occhietti, S. 1986b: Late-glacial climatic oscillation in Atlantic Canada equivalent to the Allerod-Younger Dryas event; Nature, v. 323, no. 6085, p. 247-250.

Mott, R. J., Matthews, J. V., Jr., Grant, D. R. and Beke, G. J. 1986a: A late glacial buried organic profile near Brookside, Nova Scotia; *in* Current Research, Part B; Geological Survey of Canada, Paper 86-1B p. 289-294.

Murphy, J. B. 1984: Geology of the southern Antigonish Highlands, Nova Scotia; *in* Current Research, Part A; Geological Survey of Canada, Paper 84-1A, p. 587-595.

Nicks, L. 1988: The study of the glacial stratigraphy and sedimentation of the Sheldon Point Moraine, Saint John, New Brunswick; M. Sc. thesis, Dalhousie University, Halifax, Nova Scotia, 171 p.

Nielsen, E. 1976: The composition and origin of Wisconsinan tills in mainland Nova Scotia; Ph.D. thesis, Dalhousie University, Halifax, Nova Scotia, 256 p.

Ogden, J. G. 1987: Vegetational and climatic history of Nova Scotia; I. Radiocarbon-dated pollen profiles from Halifax, Nova Scotia; Canadian Journal of Botany, v. 65, no. 7, p. 1482-1490.

Oldale, R. N., Grimes, J. R. and Whitmore, F. C., Jr. 1988: Late Wisconsinan elephant teeth from the western Gulf of Maine; *in* Geological Society of America, Northeastern Section, 23rd Annual Meeting Program with Abstracts.

Peltoniemi H. 1985: Till lithology and glacial transport Kuhmo, eastern Finland; Boreas, v. 14, p. 67-74.

Prest, V. K. 1973: Surficial deposits of Prince Edward Island; Geological Survey of Canada, Map 1366A.

Prest, V. K. 1977: General stratigraphic framework of the Quaternary in eastern Canada; Geographie Physique et Quaternaire, v. 31, p. 7-14.

Prest, V. K. and Grant, D. R. 1969: Retreat of the last ice sheet from the Maritime Provinces - Gulf of St. Lawrence region; Geological Survey of Canada, Paper 69-33, 15 p.

Prest, V. K., Grant, D. R., MacNeill, R. H., Brooks, I. A., Borns, H. W., Ogden, J. G., III, Jones, J. F., Lin, C. L., Hennigar, T. W. and Parsons, M. L. 1972: Quaternary geology, geomorphology and hydrogeology of the Atlantic Provinces; 24th International Geological Congress, Excursion Guidebook, A61-C61, 79 p.

Prest, V. K. and Nielsen, E. 1987: The Laurentide ice sheet and long distance transport; *in* INQUA Till Symposium, Finland 1985, eds. R. Kujansuu and M. Saarnisto, Geological Survey of Finland, Special Paper 3, p. 91-102.

Prest, W. H. 1896: Glacial succession in central Lunenburg, Nova Scotia; Proceedings and Transactions of the Nova Scotian Institute of Science, Halifax, v. 9, session 1895-96, p. 158-170.

Purdy, C. A. 1951: Pleistocene geology of the Kentville area, Nova Scotia; M.Sc. thesis, Acadia University, Wolfville, Nova Scotia; Nova Scotia Department of Mines and Energy, Thesis 299, 51 p.

Quinlan, G. and Beaumont, C. 1982: The deglaciation of Atlantic Canada as reconstructed from the postglacial relative sea-level record; Canadian Journal of Earth Sciences, v. 19, no. 12, p. 2232-2246.

Quinty, F. et Filion, L. 1989: Decrochements pelliculaires et gelifluxion en nappe sur le versant nord de drumlins, region de lac Minto, Quebec subarctique; Geographie Physique et Quaternaire; v. 43, no. 1. p. 51-64.

Radtke, U., Mangini, A. and Grun, R. 1985: ESR dating of marine fossil shells; Nucular Tracks, v. 10, no. 4-6, p. 879-884.

Railton, J. B. 1972: Vegetation and climatic history of southwestern Nova Scotia in relation to a South Mountain ice cap; Ph. D. Thesis, Dalhousie University, Halifax, Nova Scotia, 146 p.

Rampton, V. N., Gauthier, R. C., Thibault, J. and Seaman, A. A. 1984: Quaternary geology of New Brunswick; Geological Survey of Canada, Memoir 416, 77 p.

Schnikter, D. 1987: Late Glacial paleoceanography of the Gulf of Maine; International Union of Quaternary Research, Twelfth International Congress, Programme and Abstracts, p. 260.

Scott, D. B., Medioli, F.S. and Miller, A. A. L. 1987: Holocene sea levels, paleooceanography and late glacial ice configuration near the Northumberland Strait, Maritime Provinces; Canadian Journal of Earth Sciences, v. 24, p. 668-675.

Shaw, J. 1982: Melt-out till in the Edmonton area, Alberta, Canada; Canadian Journal of Earth Sciences, v. 19, p. 1048-1569.

Shepard, F. P. 1930: Fundian faults or Fundian glaciers; Geological Society of America Bulletin, v. 41, p. 659-674.

Stea, R. R. 1982: The properties, correlation and interpretation of Pleistocene sediments in central Nova Scotia; M.Sc. thesis, Dalhousie University, Halifax, Nova Scotia; Nova Scotia Department of Mines and Energy, Thesis 503, 215 p.

Stea, R. R. 1983: Surficial geology of the western part of Cumberland County, Nova Scotia; *in* Current Research, Part A; Geological Survey of Canada, Paper 83-1A, p. 197-202; also *in* Mines and Minerals Branch, Report of Activities, 1982; Nova Scotia Department of Mines and Energy, Report 83-1, p. 357-358.

Stea, R. R. 1984: The sequence of glacier movements in northern mainland Nova Scotia determined through mapping and till provenance studies; *in* Correlation of Quaternary Chronologies, ed. W. C. Mahaney, p. 279-297.

Stea, R. R. 1988: Holocene sea levels, paleooceanography and late glacial ice configuration near the Northumberland Strait, Maritime Provinces: Discussion; Canadian Journal of Earth Sciences, v. 25, no. 2, p. 348-350.

Stea, R. R. and Finck, P. W. 1984: Patterns of glacier movement in Cumberland, Colchester, Hants and Pictou Counties, northern Nova Scotia; *in* Current Research, Part A; Geological Survey of Canada, Paper 84-1A, p. 477-484.

Stea, R. R., Finck, P. W. and Wightman, D. M. 1985: Quaternary geology and till geochemistry of the western part of Cumberland County, Nova Scotia (sheet 9); Geological Survey of Canada, Paper 85-17, 57 p.

Stea, R. R. and Mott, R. G. 1989: Deglaciation environments and evidence for glaciers of younger Dryas age in Nova Scotia, Canada; Boreas, v. 18, p. 169-187.

Stea, R. R., Scott, D. B., Kelley, J. T., Kelley, A. R., Wightman, D. M., Finck, P. W., Seaman, A., Nicks, L., Bleakney, S., Boyd, R. and Douma, M. 1987: Quaternary Glaciations, Geomorphology and Sea Level Changes: Bay of Fundy Region; NATO Advanced Course on: Sea Level Correlations and Applications, Halifax, Nova Scotia, Field Trip Guidebook, 79 p.

Stea, R. R. and Wightman, D. M. 1987: Age of the Five Islands Formation and the deglaciation of the Bay of Fundy; Quaternary Research, v. 27, p. 211-219.

Stevenson, I. M. 1959: Shubenacadie and Kennetcook Map Areas, Colchester, Hants and Halifax Counties, Nova Scotia; Geological Survey of Canada, Memoir 302, 88 p.

Stuckenrath, R., Jr., Coe, W. R. and Ralph, E. K. 1966: University of Pennsylvania radiocarbon dates IX; Radiocarbon, v. 8, p. 348-385.

Swayne, L. E. 1952: The Pleistocene geology of the Digby area, Nova Scotia; M.Sc. thesis, Acadia University, Wolfville, Nova Scotia; Nova Scotia Department of Mines and Energy, Thesis 362, 56 p.

Van Steijn, H. 1987: The role of debris flows in the development of stratified slope deposits as inferred from modern analogues; International Union for Quaternary Research, INQUA 87, July 31 - August 9, 1987, Ottawa, Canada; Programme with Abstracts, p. 280.

Van Steijn, H. 1988: Debris flows involved in the development of Pleistocene stratified slope deposits; Zeitschrift fuer Geomorphologie, v. 71 p. 45-58.

Vernal, (de) A., Causse, C., Hillaire-Marcel, C., Mott, R. J. and Occhietti, S. 1986: Palynostratigraphy and Th/U ages of upper Pleistocene interglacial and interstadial deposits on Cape Breton Island, eastern Canada; Geology, v. 14, p. 554-557.

Watson, P. D. 1984: The sedimentology of glaciofluvial deposits near Upper Nine Mile River, Nova Scotia; B. Sc. thesis, Dalhousie University, Halifax, Nova Scotia, 57 p.

Wehmiller, J. F., Belknap, D. F., Boutin, B. S., Mirecki, J. E., Rahaim, S. D. and York, L. L. 1988: A review of the aminostratigraphy of Quaternary mollusks from United States Atlantic Coastal Plain sites; *in* Dating Quaternary Sediments, ed. D. J. Easterbrook, Geological Society of America, Special Paper 227, p. 69-110.

Welsted, J. E. 1971: Morphology and evolution of the Bay of Fundy with emphasis on changes of sea-level during the Quaternary; unpublished Ph.D. dissertation, Bristol University, Bristol, England; Nova Scotia Department of Mines and Energy, Thesis 504, 253 p.

Wightman, D. M. 1980: Late Pleistocene glaciofluvial and glaciomarine sediments on the north side of the Minas Basin, Nova Scotia; Ph.D. thesis, Dalhousie University, Halifax, Nova Scotia; Nova Scotia Department of Mines and Energy, Thesis 405, 426 p.