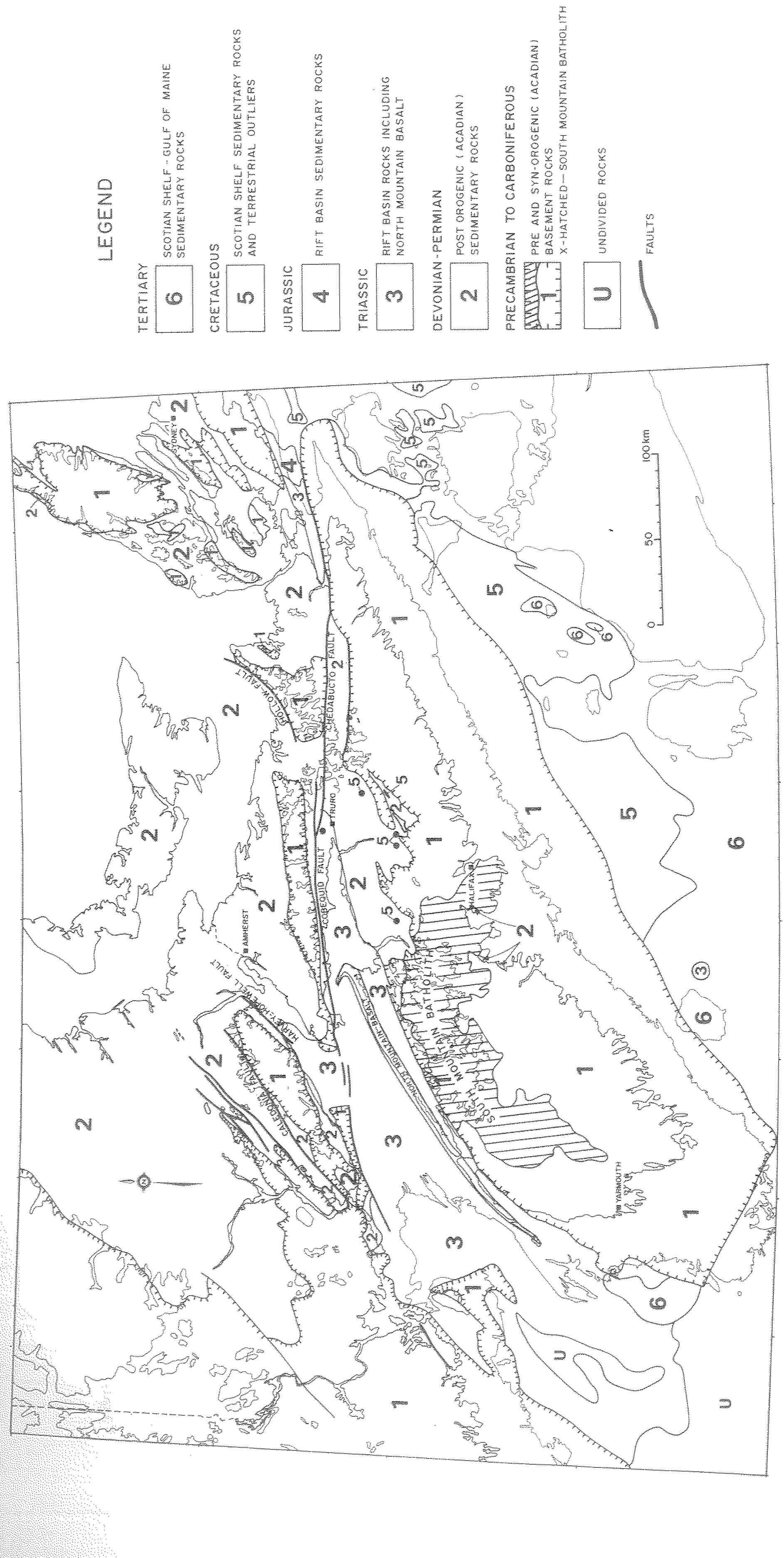


Figure 3. Generalized geology of the Bay of Fundy region

two basic theories of landform development - (1) the concepts of the geomorphic cycle (Davis, 1922) and (2) the steady-state landscape hypotheses (Hack, 1960).

Goldthwait (1924) interpreted the geomorphic evolution of Nova Scotia in the light of Davisian concepts. He speculated that the entire region was infilled after the Triassic and then eroded to a peneplane. The age of this cycle would predate the deposition of kaolinitic clays and silica sands found in isolated sites in lowland regions of Nova Scotia. Goldthwait believed that these clays were correlative with Cretaceous clays in New Jersey and therefore a Cretaceous age was postulated for peneplanation. The remnants of that peneplane are the erosional surface represented by the flat accordant upland surfaces of the Caledonian Highlands in New Brunswick, and the North Mountain, the Cobequid Highlands and the Atlantic Uplands of Nova Scotia. Southward flowing rivers became superposed 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, align with modern rivers. Examples of this alignment are the Parrsboro and Folly Gaps with the Shubenacadie and Avon Rivers, respectively (Fig. 2).

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

Welsted (1971) argued that the trough of the Bay of Fundy existed since the Triassic. He contended that no evidence existed for extensive post-Triassic cover of rocks and little evidence for superposition of streams. Welsted interpreted the flat uplands as exhumed surfaces of great antiquity that did not form during the same cycle of erosion. 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 cuesta was formed by longitudinal streams cutting through the basalt into the underlying softer sediments while tributaries denuded the basalt slopes.

Recent work has brought to light evidence favoring Goldthwait's interpretation of the formation of the Bay of Fundy. King (1972a) suggested that the lowlands of the Bay of Fundy may have been covered by extensive Cretaceous and Tertiary sediments and subsequently exhumed by uplift. However, he did question the existence of a peneplane. The presence of widespread unconformities in the offshore sequence prompted King (1972a) to recognize that rejuvenation of the landscape had occurred several times from the Cretaceous to the Recent.

Hacquebard (1984) calculated that the depth of burial of lignites in the Early Cretaceous outliers was 700 - 1000 m (2300 - 3280 ft). At this depth of burial Early Cretaceous or younger sediments would easily have overstepped the present day Atlantic Uplands as well as the Cape Breton Highlands which attain heights of 600 m (1968 ft; Fig. 2). The fact that Cretaceous outliers are found on Triassic rocks near Truro, as well as on Carboniferous rocks in Nova Scotia and New Brunswick (Fowler and Stea, 1979), suggests that the Bay of Fundy basin was also covered (Fig. 3). Tertiary sediments are known 40 km (25 m) southwest of Yarmouth on the Meguma Group rocks of the Atlantic Uplands as well as in numerous outliers in the Gulf of Maine (King, 1972b; G. B. Fader, personal communication, 1986; Fig. 3).

Implicit in Welsted's (1971) theory of the evolution of the Bay of Fundy is that the present topographical relationships were maintained since the formation of the bay. If this theory is correct and a Cretaceous or younger basin fill of up to 1000 m (3280 ft) is assumed, then a minimum of 1200 m (3937 ft) of erosion of the Atlantic Uplands is required to maintain the present height differences between the Atlantic Uplands and the Bay of Fundy. This does not take into account the possible subsidence in the fault-bound basins. Extensive erosion of the Atlantic Uplands in post-Viséan time is unlikely given the preservation of thin skins of Windsor Group reef carbonates on the upland surface (Giles, 1981).

The sum of this evidence implies that the Bay of Fundy formed after the Triassic-Jurassic rifting episode and was infilled and exhumed at least once and perhaps twice in the Cretaceous and Tertiary periods. Studies of the provenance of heavy minerals in the Cretaceous sediments (Stea and Fowler, 1981; Dickie, 1986) support this concept. Goldthwait (1924) postulated that the surface of the Cretaceous peneplane was coursed by southward flowing consequent streams. The source of heavy minerals in some of these Cretaceous outliers appears to be areas to the north which are not connected by present drainage patterns.

TIDES OF THE BAY OF FUNDY

The tides of the Bay of Fundy are the highest in the world reaching 16 m (53 ft) at the head of bay (Fig. 4). The unusual height of the tides is due to a funnelling effect and to the resonance period of the Fundy Basin. The 12.4 hr 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 3 hrs 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 accumulate on one side of the bay due to the Coriolis effect. The range of the tides are up to 1.5 m (5 ft) 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 (70.9 ft) at the head of the bay.

Tides Through Time

The tidal range in the Bay of Fundy has not always been as great as today. Wightman (1976) studied the sedimentology of Late Wisconsinan raised beach deposits in the Advocate Harbour area (see Stop 2; Day 6). He based paleotidal estimates on the assumption that the foreshore facies extends from the subtidal to the supratidal zone and that the thickness of this facies approximates the paleotidal range. He estimated that a 3.4 m (11.2 ft) tidal range existed during the deposition of the beach deposits 13,000-12,000 yr B.P. The lower range of the tides during the deglaciation period may have been due to the presence of residual ice streams in the Minas and Cumberland Basins. Ice in these basins would effectively change the oscillation period of the Bay of Fundy.

Greenberg (1978) produced a model which can simulate the tides in the Bay of Fundy if the configuration in the basin is known (i.e. the water depth). Scott and Greenberg (1983) tackled the problem of Holocene paleotides of the Bay of Fundy. Essentially, tidal ranges showed little change in the last 4000 years but had increased 30-50% between 7000 and 4000 yr B.P. Perhaps more importantly, they showed that water depth over Georges Bank (not within the bay) controlled the tidal amplification. At 7000 yr B.P., when sea level was 30-40 m lower than present on Georges Bank, most of the bank was exposed and the volume of water into the Bay of Fundy decreased. As the bank was submerged, the flow into the bay increased and the funnelling and resonance effects combined to produce a high tidal range.

Sea level researchers who use standlines or other high water markers as a measurement baseline should be aware of tidal ranges through time. The high water mark could vary by several metres, with only a small tidal expansion.

A famous feature of the Bay of Fundy is the tidal bore created when the tidal surge enters a shallow estuary. Under ideal conditions the bore can be a wave 0.9 m (3.0 ft) high with speeds up to 13 km/hour (8 mi/hour). We may see the bore in our travels around the bay with particular opportunity at Truro (Day 6).

QUATERNARY EVENTS

In the late 1800s when the glacial theory started to gain favour, a controversy emerged about the nature of glaciation in the Bay of Fundy region. The question exists whether the ice was local, originating in upland areas and confined to the land masses, or part of a great continental mass which crossed the Bay of Fundy. Reverend D. Honeyman, who was curator of the provincial museum in the late 1800s, discovered amygdaloidal basalt boulders along the Atlantic coast near Halifax (Honeyman, 1876). These boulders were derived from the North Mountain, a basaltic cuesta which forms the border of the Nova Scotia side of the Bay of Fundy, and transported a distance of 130 km (80 mi). Honeyman used this observation to support the concept of a continental-based ice movement that crossed the Bay of Fundy.

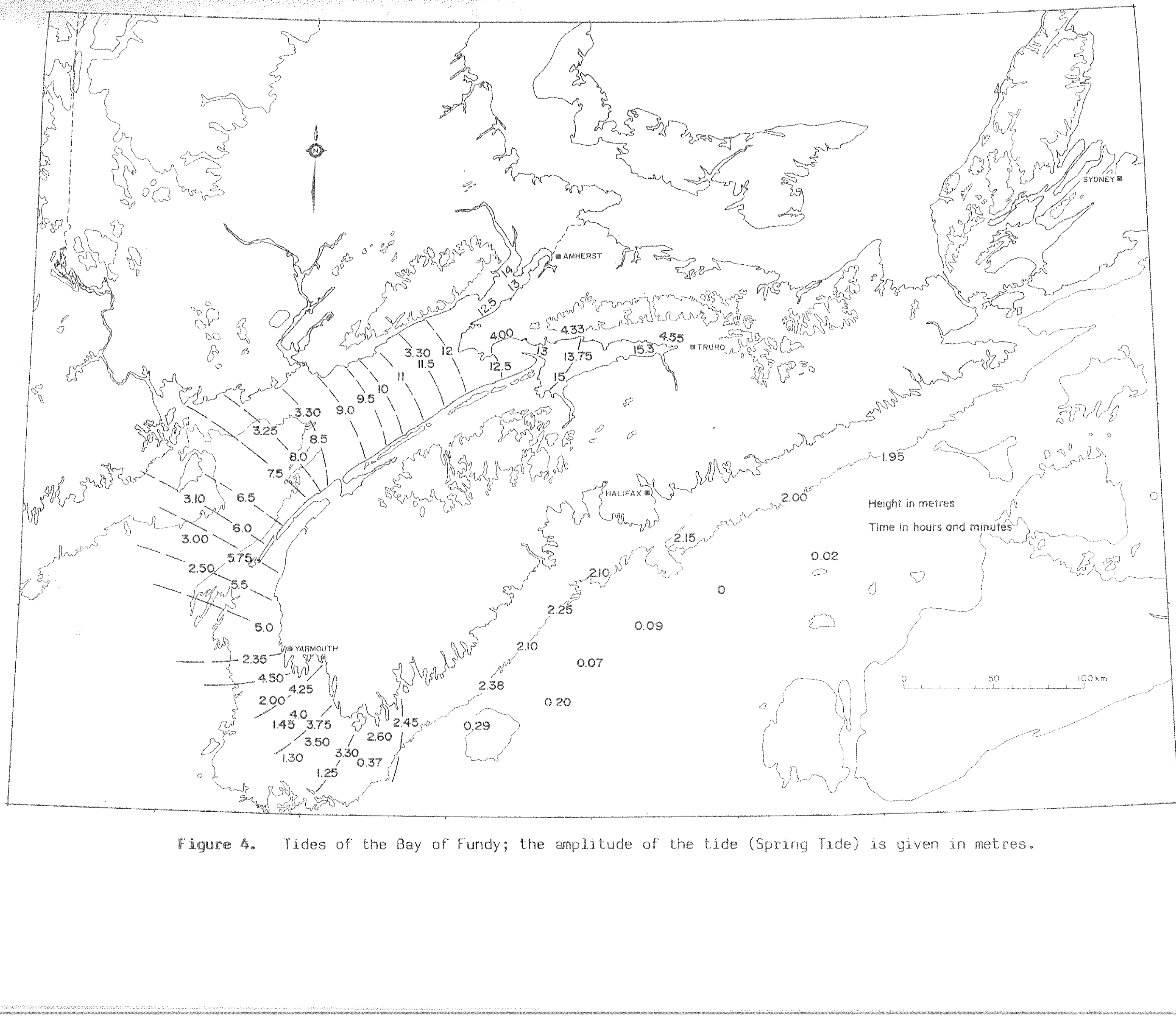


Figure 4. Tides of the Bay of Fundy; the amplitude of the tide (Spring Tide) is given in metres.

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 was glaciated largely by local glaciers with floating ice a secondary agent in low-lying areas. Contrary to Honeyman, Chalmers did not believe that a glacier had extended across the Bay of Fundy. Chalmers (1895, p. 95m) stated:

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

L. W. Bailey (1898) and W. H. Prest (1896) worked in mainland Nova Scotia and observed erratics that supported the view of ice movement across the Bay of Fundy and the existence of local ice caps. Bailey stated the compromise position (1898, p. 26m):

> 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 "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 stemming from ice in the Gulf of St. Lawrence called the Acadian Bay Lobe. Goldthwait's (1924) ideas gained ascendancy in the following years, and much of the earlier work was discredited. Wickenden (1941, p. 143) for example dismissed Chalmers (1895) work as "without value".

Pleistocene mapping in the Annapolis Valley which was initiated at Acadia University, Wolfville (MacNeill, 1951; Purdy, 1951; Swayne, 1952) revived the concepts of continental, then local, glaciations (MacNeill and Purdy, 1951). At the same time, Flint (1951) proposed on the basis of work by Chalmers (1895) that two local ice centres existed in Nova Scotia. 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 time, radiocarbon dating was established and processes of glacier mechanics were clarified. The timing of glaciations in Nova Scotia emerged 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. These authors believed 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. The Late Wisconsinan substage (25,000-10,000 yr B.P.) was established as the time of the last glacial maximum. King (1969) dated the last glacial maximum in Nova Scotia at 18,800 yr B.P. The Scotian Shelf morainal system was concluded to represent the terminus of a major continental ice advance. Grant (1975) proposed a southward flood of ice prior to 39,000 yr B.P. followed by a retreat about 38,000 yr B.P. 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 extended across the Bay of Fundy and that mainland Nova Scotia, as well as Cape Breton Island, harboured their own ice caps. This was accepted as the minimum model. Grant (1977b, 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 the earlier workers the Late Wisconsinan Laurentide ice sheet filled the Bay of Fundy, crossed Nova Scotia and terminated offshore (Flint, 1971). Figure 2 shows the ice distribution in the maximum and minimum models. 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.

There is evidence for a more complex picture of ice movements than that envisioned by Grant (1977) in the Late Wisconsinan and a compromise between the minimum and maximum models. Quinlan and Beaumont (1981) through modelling of relative sea level, inferred a Late Wisconsinan ice distribution somewhere between the minimum and maximum models. Recent mapping in Nova Scotia has outlined a complex pattern of Early to Late Wisconsinan ice flows. The ice movement history of Nova Scotia is characterized by diachronous advances and retreats of ice caps centred north of the Province on the Atlantic Uplands, Antigonish Highlands and Cape Breton Lowlands (Stea, 1984). The Bay of Fundy region was first overrun by ice flowing eastward then southeastward during Ice Flow Phase 1 (Fig. 5) in the Early Wisconsinan. Following this major ice flow which extended to the shelf edge, ice flowed southward and was funneled through the Bay of Fundy from a centre near Prince Edward Island (Ice Flow Phase 2; Fig. 5). Ice divides were then created across the Nova Scotia peninsula and ice flowed northward (Ice Flow Phase 3; Fig. 5). Marine incursion in the lower reaches of the Bay of Fundy halted the spread of this ice mass. The last event was flow out of the Minas Basin from a centre in the Antigonish Highlands (Ice Flow Phase 4). The Isle Haute Moraine (Fig. 2) may have formed as an interlobate moraine early during Ice Flow Phase 4 as ice was confluent from the Minas Channel and Chignecto Bay. The last three ice flow phases are believed to span Middle to Late Wisconsinan time.

DEGLACIATION

Concepts on the deglaciation of the Bay of Fundy are linked to the maximum or minimum views of the Late Wisconsinan ice sheets. Denton and Hughes (1981) suggested that the pattern of radiocarbon dates on emerged marine sediments in the region are an indirect indication of the presence of calving ice streams in the Bay of Fundy around 14,000 yr B.P. Dates from the head of the Bay of Fundy, however, are notably lacking. Radiocarbon dates on sediments formed directly by retreating glaciers at the head of the Bay of Fundy are therefore crucial in the resolution of the extent of ice in the Late Wisconsinan.

Recent dates on glaciomarine sediments at Spencers Island (see stop 3; Day 3) have shed some new light on this problem. Three AMS dates have been obtained on shell material from the bottomset beds of a glaciomarine delta at Spencers Island, Nova Scotia near the head of the Bay of Fundy. The sediments in the delta are part of the previously undated Five Islands Formation, and are the first direct indication of the age of deglaciation in this region. The Five Islands Formation encompasses glaciofluvial and glaciomarine sediments that form raised and terraced outwash plains, deltas and raised beaches along the shores of the Minas and Cumberland Basins, Nova Scotia (Swift and Borns, 1967; Wightman, 1980; Stea et al. 1986; Fig. 6).

The Five Islands Formation is divided into two members - the Advocate Harbour Member and the Saints Rest Member. The Saints Rest Member comprises glaciofluvial deposits which include the outwash and topset beds of the deltas. The Advocate Harbour Member includes the bottomset and foreset parts of the deltas and raised littoral deposits (Fig. 6). The deltas mark prominent ice marginal stand positions from the retreat of the Late Wisconsinan glaciers (Wightman, 1980). Figure 7 shows the stages of formation of the Minas deltas.

The dates on the Spencers Island delta range 14,300 - 12,600 yr B.P and record the deposition of the delta and a diamicton under the deltaic deposits (Stea and Wightman, in press). The diamicton formed approximately 14,000 yr B.P. under ice shelf or calving bay conditions, or by a re-advance of grounded ice. The Spencers

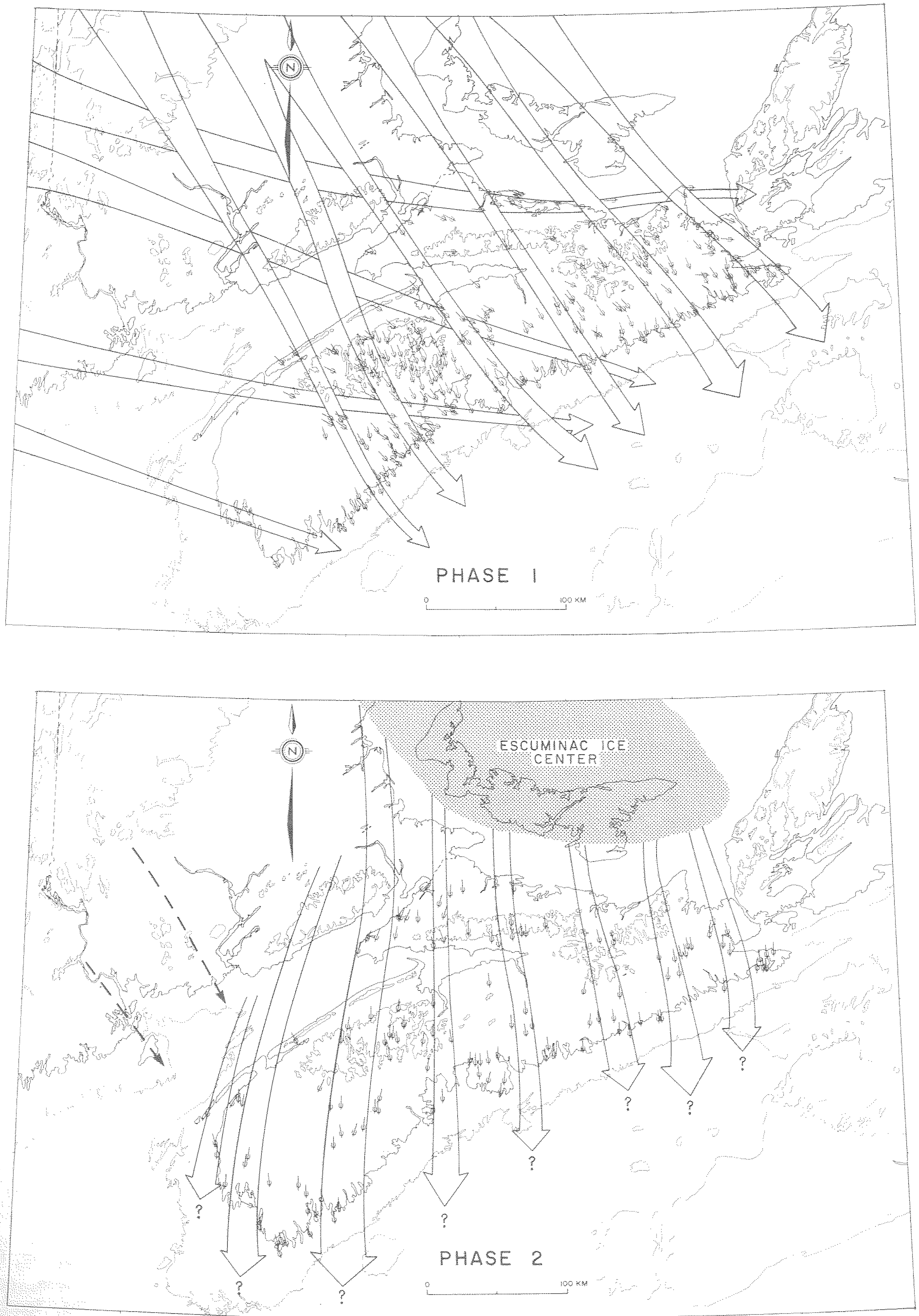


Figure 5a. Ice flows in Nova Scotia during the Wisconsinan Stage; Phase 1 - Illinoian? to Early Wisconsinan, Phase 2 - Middle to Late Wisconsinan.

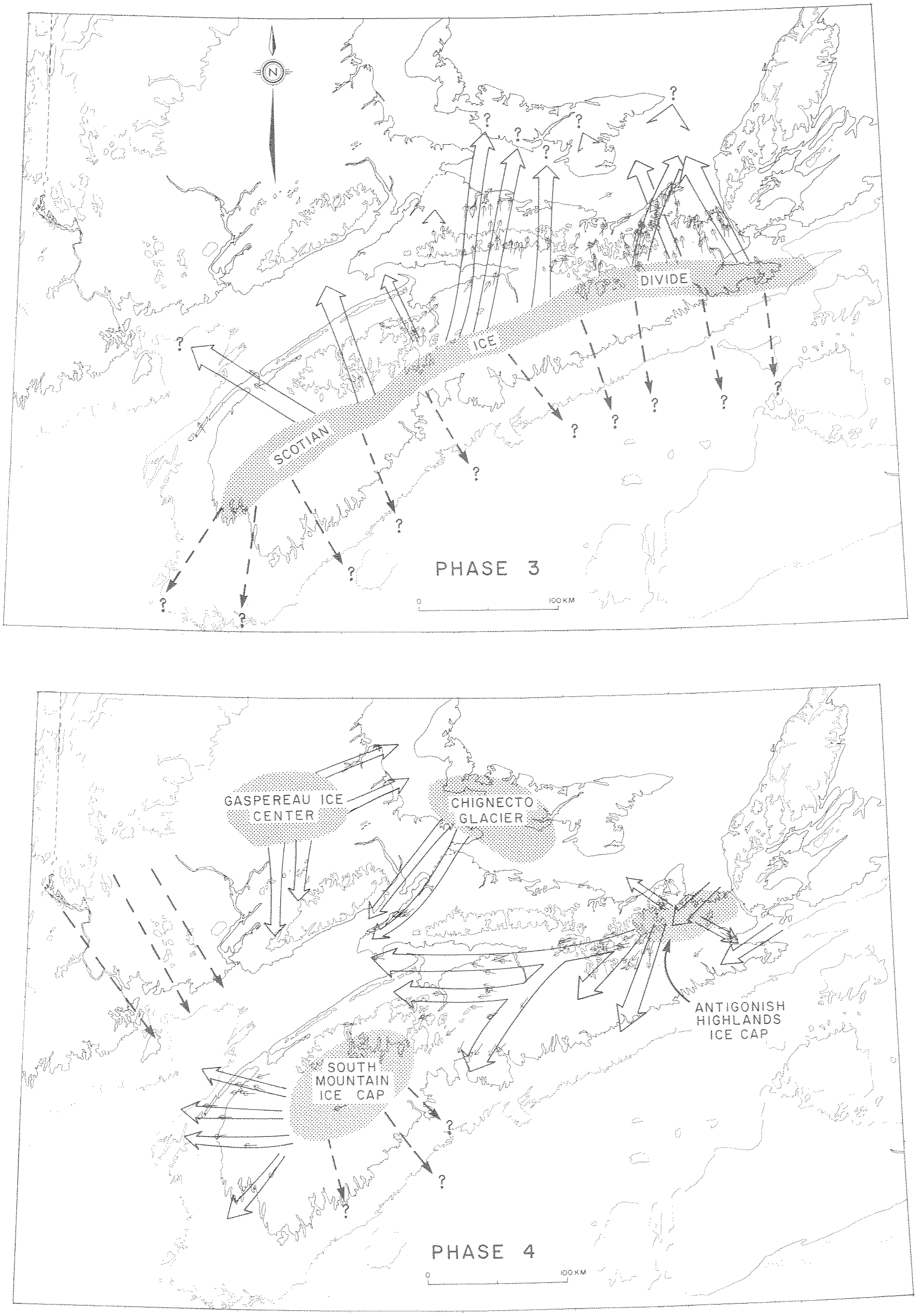
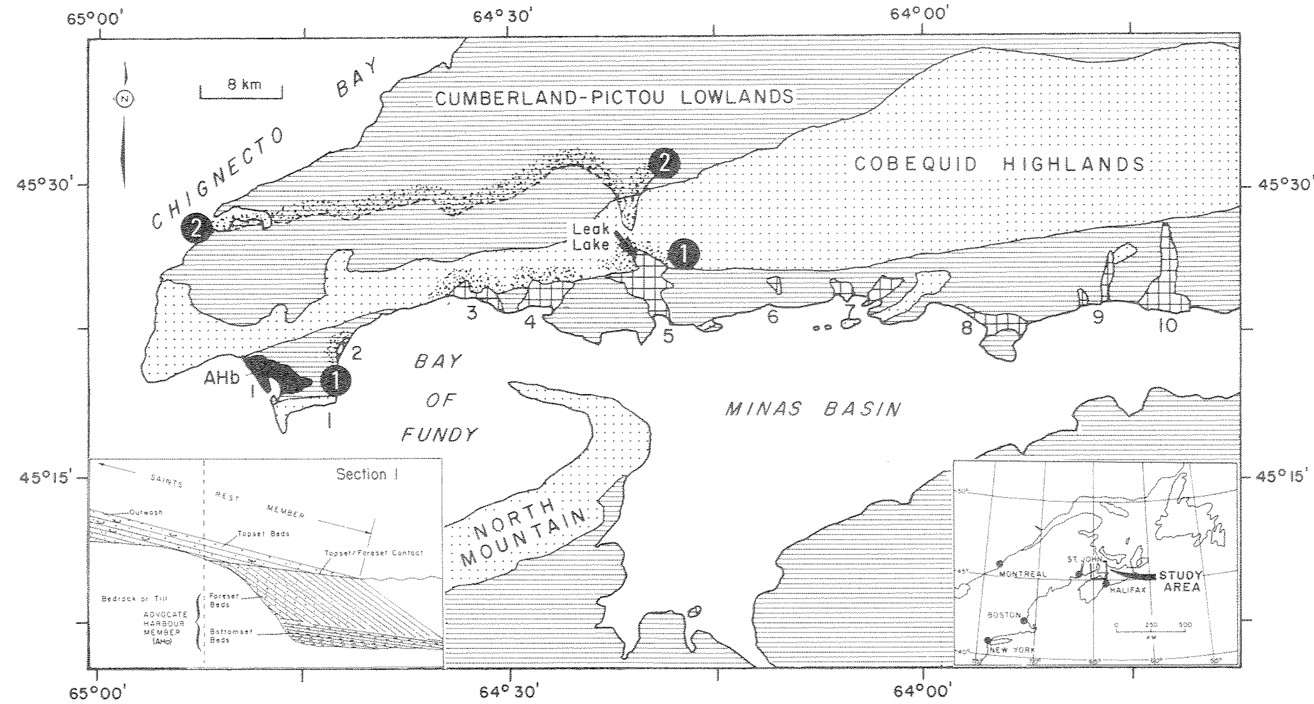


Figure 5b. Ice flow in Nova Scotia during the Wisconsinan Stage; Phase 3 - Late Wisconsinan, Phase 4 - Late Wisconsinan.



LEGEND

QUATERNARY
LATE WISCONSINAN
FIVE ISLANDS FORMATION:
Glaciofluvial, glaciolacustrine, and glaciomarine deposits laid down during retreat of the glaciers and rise and fall of sea level

ADVOCATE HARBOUR MEMBER; MARINE SEDIMENTS: AHa, marine gravel, sand, silt and clay; forming a progradational sequence of littoral and sublittoral deposits: planar bedded gravel and sand beds dipping 10° to 35° and rhythmites of silt, clay, sand forming foreset and bottomset beds of a glaciomarine delta sequence (seen in section only, section 1). AHb, marine sand and gravel, shallow, seaward dipping (2°to 14°) strata, forming forebeach and backbeach deposits

SAINTS REST MEMBER:
Glaciofluvial gravel, sand and minor silt: massive to horizontally stratified, channel sequences common; grain size increases from proximal to distal delta and upwards in sequence, forms topset beds which unconformably overlie marine foreset and bottomset (Advocate Harbour Member) in a classic "Gilbert Delta" sequence (See section 1) : thickness 1-15m

ICE MARGIN # 1

ICE MARGIN # 2

PRE-QUATERNARY
Sedimentary Lowlands

Igneous and Metamorphic Highlands

Figure 6. Map of the deltas and morainal stands of the Chignecto Peninsula, Nova Scotia with a cross-section of an idealized raised marine Gilbert-Type delta and its constituent members

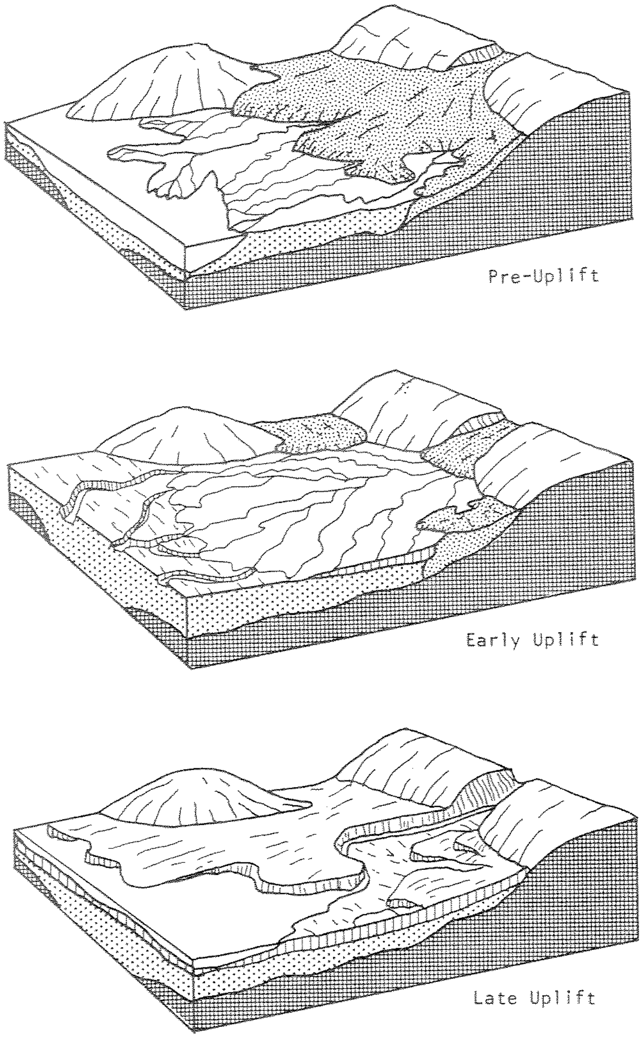


Figure 7. Evolution of the Minas north shore outwash-delta terrace. Top- growth of marine deltas. Middle-subaerial fans overstep the delta. Bottom-modern terrace after uplift, dissection and sea level rise. (after Swift and Borns, 1967)

Island delta is part of a prominent ice marginal stand marked by numerous deltas along the Minas Basin. The time of formation of the deltas and the inferred ice margin is 12,000 - 13,500 yr B.P. based on the dates of the Spencers Island sediments and the base of lake sediment cores from the delta surface (the latter is confirmed by palynology). Ice marginal glaciomarine deposits near Saint John, New Brunswick record a range of radiocarbon dates similar to the Spencers Island dates (Fig. 8). The Presumpscot Formation in Maine (Bloom, 1963) is the Late Wisconsinan equivalent of the Advocate Harbour Member. It is a grey, laminated, fossiliferous marine mud. Dates within this formation range 13,320 - 11,500 yr B.P. The synchroneity of deglacial dates obtained at the head and mouth of the bay would suggest that the bay was virtually ice free at 14,000 yr B.P. Several total carbon dates on the marine/freshwater transition in lakes in New Brunswick may, if valid, extend marine conditions back to 16,000 yr B.P. (Scott and Medioli, 1980).

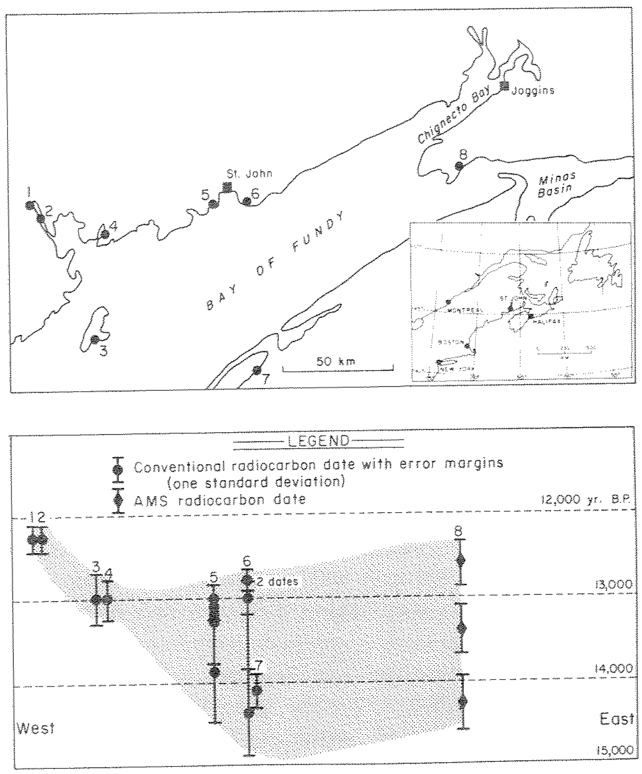


Figure 8. Dates of marine shells from moraines and ice-marginal deltas along the Bay of Fundy

RELATIVE SEA LEVEL CHANGE

INTRODUCTION

The first researchers of the Bay of Fundy region were made aware of changing relative sea level (RSL) by the presence of emerged shorelines. The age of the shorelines or the rates of sea level change could not be determined as no absolute dating technique existed. The first comprehensive work on changing RSL in the Maritimes was by Grant (1970) who dated, by the C^{14} method, many sites of Holocene sea levels. The RSL during the Holocene was shown to be rising in all areas. Scott and Medioli (1982), Scott and Greenberg (1983) and Scott et al. (1981, 1984, 1987, in press) have provided rates of RSL rise for 15 locations around Eastern Canada some of which will be visited as part of this field trip.

The rates of RSL rise tend to be higher along the Atlantic coast of Nova Scotia and inside the Bay of Fundy and show a trend of decreasing rates from east to west (i.e. rate of RSL rise decreases towards the former ice centre). However, the rate of RSL rise is anomalously high compared with areas away from the former ice margin (i.e. Florida, Bermuda) and comparable with those in New England (Redfield, 1967).

Various hypotheses explain the anomalous RSL rise. Grant (1970) suggested water loading on the continental shelf. Although the continental shelf off Florida is analogous to the Scotian Shelf, the former is not experiencing anomalous RSL rise and thus the water loading hypothesis does not appear feasible. A model incorporating the earth's rheology and crustal response to deglaciation was suggested (Peltier and Andrews, 1976). This general model, which suggested a peripheral forebulge collapsing across the formerly glaciated terrain, predicts differing RSL rises as well as anomalously high RSL rise along a former glacial margin (Fig. 9). The model was applied to Eastern Canada specifically (Quinlan and Beaumont, 1981; 1982) and the results are in agreement with the observed level changes. The model predictions and field observations (Fig. 10) indicate that the Late Wisconsinan ice cover was somewhere between the minimum and maximum models with little or no grounded ice in the southern Gulf of St. Lawrence. Thin ice (<1 km (0.6 mi) thick) covered much of the Bay of Fundy region. The effect of the retreating forebulge produced a series of RSL 'zones' based on the changes of RSL since deglaciation (Fig. 10). A zone of continuous submergence (zone D) produced by collapse of the leading edge of the forebulge is located at the periphery of the ice load. Zones C, B and A parallel the migration of the forebulge and show various times of emergence and submergence (Fig. 9).

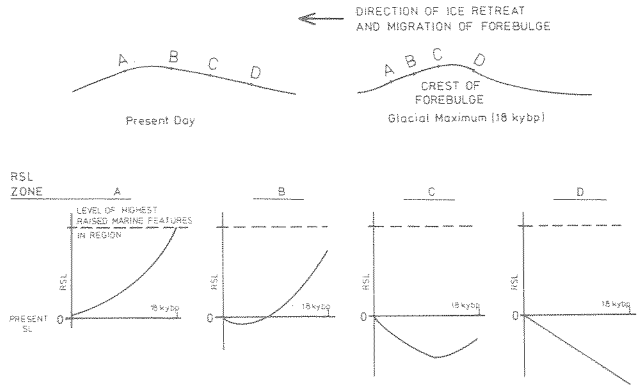


Figure 9. The upper diagram is a schematic representation of a peripheral bulge at 18,000 yr B.P. and present time. The bulge migrates in the direction of the arrow affecting RSL at sites A, B, C and D. The lower panel shows the RSL history at each of the sites A through D resulting from bulge migration (after Quinlan and Beaumont, 1981)

Scott and Medioli (1980a) and Honig and Scott (1987) use a method of coring lakes within marine limit to determine the RSL history of a region. The method is relatively simple involving these steps: (1) core a basin, (2) date the marine/freshwater transition and (3) measure the elevation of the lake sill. An emergence curve for the Bay of Fundy region (Fig. 11) has been produced using this method. Several total carbon dates of 15,000-16,000 yr B.P. were recorded at the base of lake cores, indicating that at 16,000 yr B.P. the Bay of Fundy may have been ice free to the head. However, Rampton et al. (1984) feel these dates are too old as similar dates obtained from the base of lake cores by Mott (1975) do not correlate with palynological assemblages. The older dates suggest deglaciation started much earlier, as

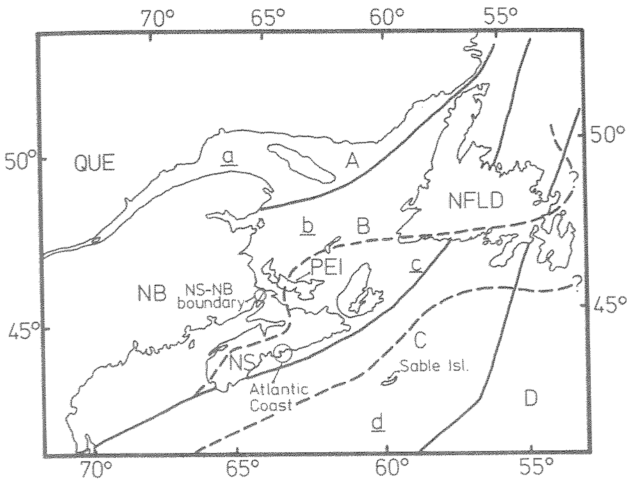


Figure 10. Theoretical sea level zones predicted in accordance with the maximum ice load configuration of Quinlan and Beaumont (solid lines, capital letters) with modifications based on field observations (dashed lines)

designated in the North Atlantic record of Ruddiman and MacIntyre (1981). Shell dates of 12,600 - 14,000 yr B.P. are from glacio-marine sediments associated with delta and moraine formation. These dates mark sea levels associated with a delta surface. If the older dates are correct then the emergence curve developed by Scott and Medioli (Fig. 11) implies that these younger dates may not be associated with maximum marine emergence and that the ice margin receded well inland before the formation of the coastal deltas.

The sea level record based on field observations from emerged features and lake cores will be discussed in three segments: (1) Sangamon sea levels, (2) Late Wisconsinan sea levels and (3) Holocene records.

SANGAMON SEA LEVELS (ca. 120,000-100,000 yr B.P.)

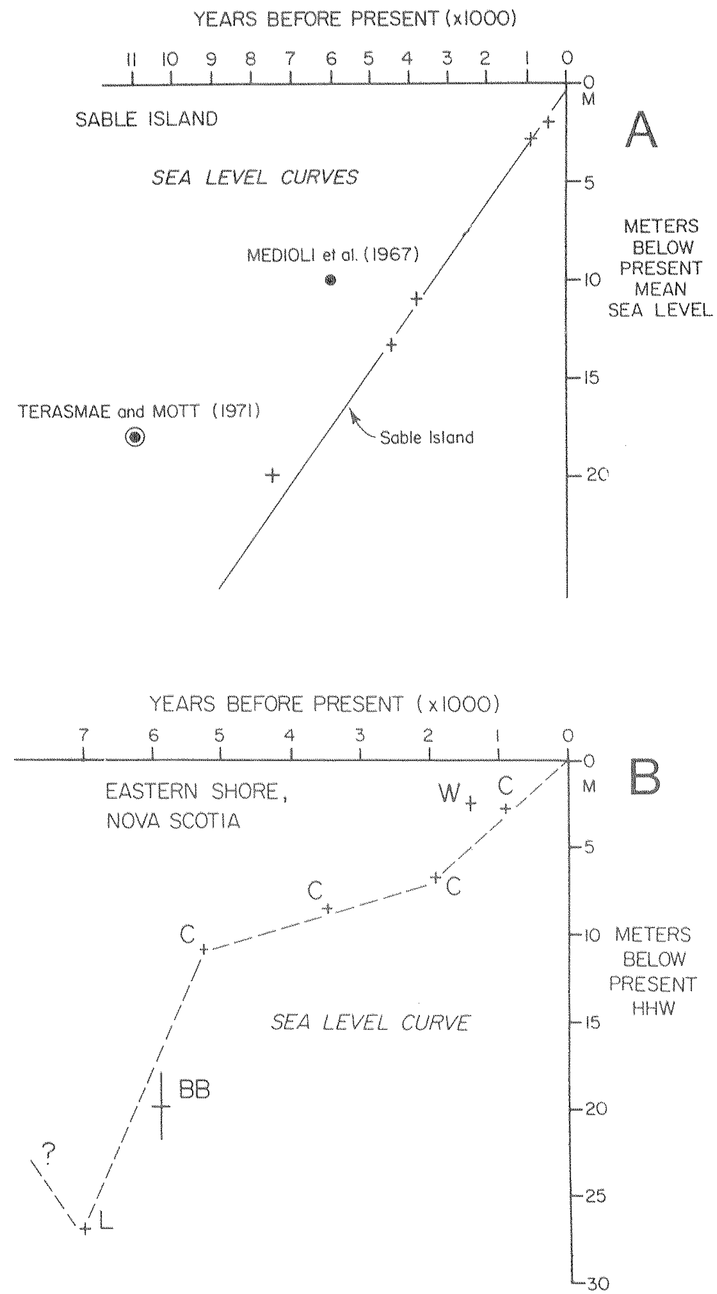
The earliest record of former sea levels in the Bay of Fundy region predates the initial eastward and southeastward ice flows across Nova Scotia (Fig. 5). Tills formed during these regional ice flows lie above a rock bench at an elevation of 4-6 m (13 - 20 ft) above sea level. This bench has been found at similar elevations in Nova Scotia (Grant, 1975) and Newfoundland (Henderson, 1972). Gravel beds, sea stacks (Grant, 1980) and striated potholes (Rampton et al., 1984) have also been found on the surface of this bench. Grant (1980) inteprets this bench as a wave-cut feature relating to higher sea levels during the Sangamon Interglacial. There are no direct dates on this feature but peat beds beyond the range of radiocarbon dating have been found above the bench in Cape Breton Island. Palynology of the beds suggest that they were formed during an Early Wisconsinan interstadial. We will make several stops along the shores of Georges Bay and the Northumberland Strait of Nova Scotia (Fig. 1) to view this feature and the overlying stratigraphy.

Grant (1980) pointed out that the feature records a former equilibrium state between sea level and crustal rebound after a major glaciation. The present interglacial period will reach this state in another 2000 yr assuming the current rate of submergence of 0.3 m/century (1 ft/century).

LATE WISCONSINAN SEA LEVELS (16,000-10,000 yr B.P.)

The Bay of Fundy is part of RSL zone B - a zone of emergence followed by subsidence (Fig. 10). Raised marine features dating 14,000 - 12,500 yr B.P. are found at various levels around the Bay of Fundy. Figure 12 shows the distribution of these features in the Bay of Fundy. Wave-cut terraces, raised beaches and deltas were the major landforms produced during the late-glacial high stand of sea level. Isolines of marine emergence based on the elevation of these features were constructed for the Bay of Fundy coastal region. The general pattern is a north to south decrease in marine emergence from >80 m (262 ft) to <40 m (131 ft) in the western region near Passamaquoddy Bay and a west to east decrease from 40 m (131 ft) to 0 m (0 ft) at the head of the bay (Fig. 12). Major perturbations in the 40 m (131 ft) isoline occur along the south shore of the Bay of Fundy and the Chignecto Peninsula areas where anomalous emergence has been noted.

The eastward decline of the paleoshore from Cape Chignecto to Truro at the head of the Bay of Fundy has been explained by two models - (1) an eastward deglaciation causing delay in marine incursion and (2) differential ice loading. Wightman and Cooke (1978) and Grant (1980)



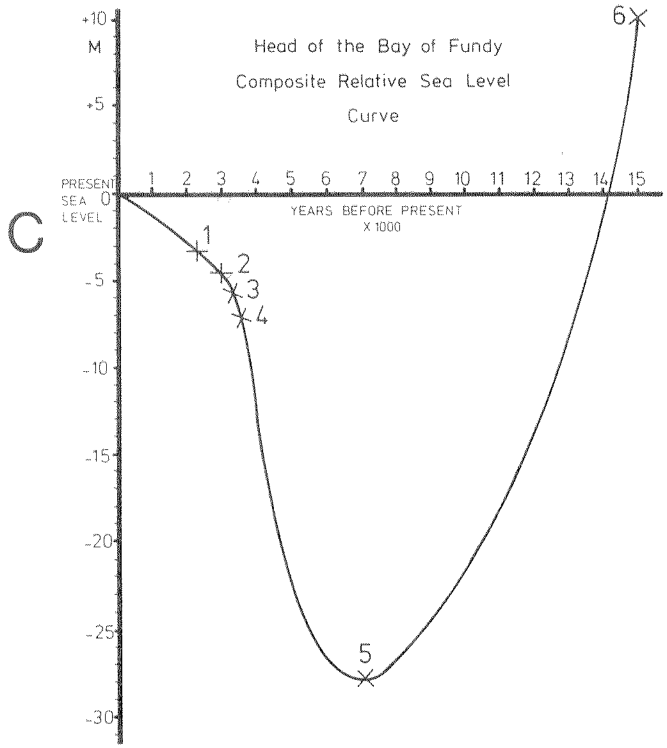


Figure 11. Composite observed RSL curves for Atlantic Canada from several sources (after Scott et al., in press)

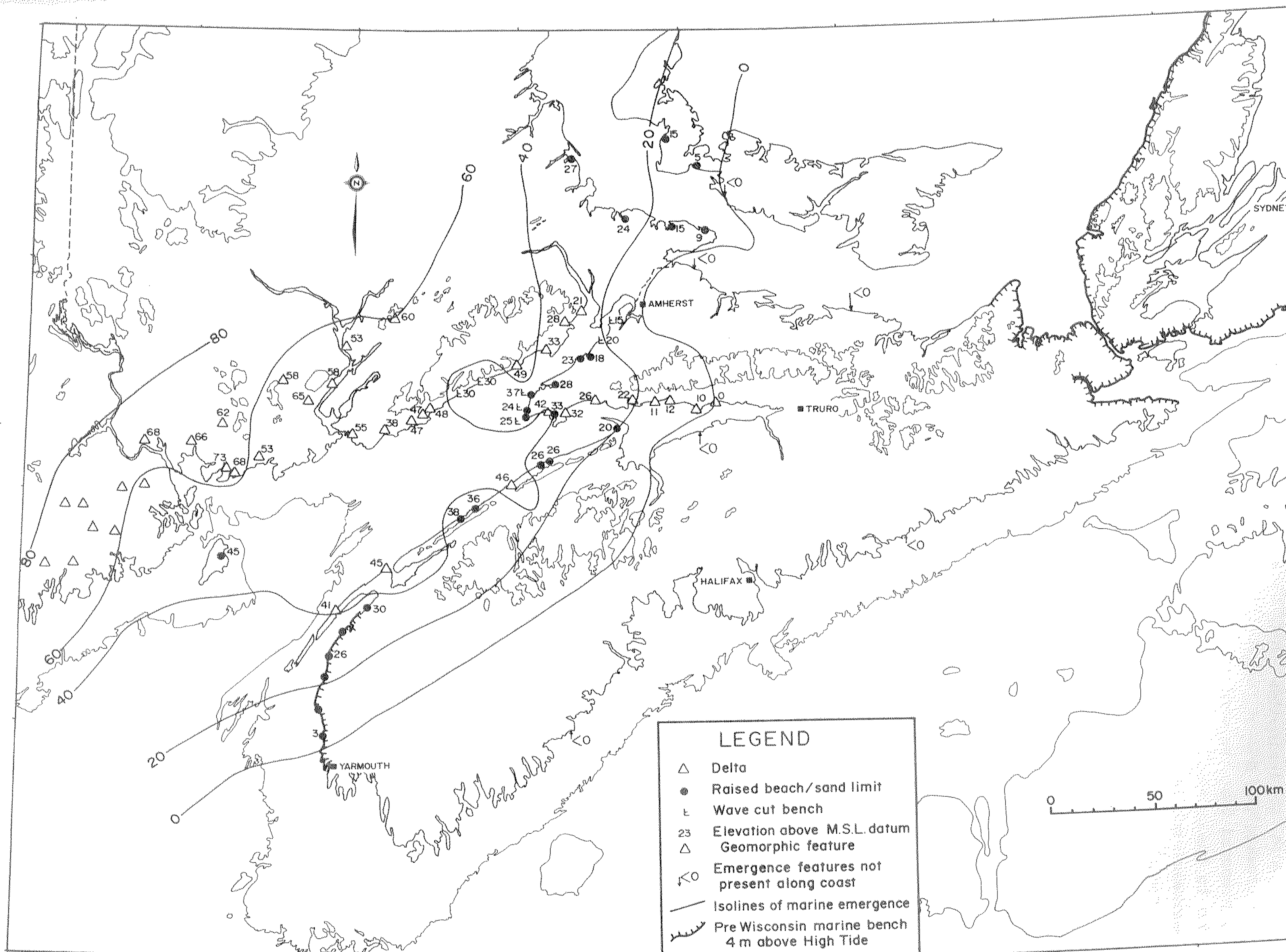


Figure 12. Plot of elevation isolines of raised marine features around the Bay of Fundy (data from Welsted, 1971; Grant, 1980; Kelley et al. 1986; Rampton et al. 1984; Wightman, 1980; Stea, 1983). Some data (Grant, 1980, Welsted, 1971) showing high water markers as datum were adjusted to mean sea level datum by using tide tables.

attributed the decline in emergence to southeastward tapering of the effective regional ice load. This theory is implicitly connected to the maximum model because it insinuates that Late Wisconsinan ice crossed the Bay of Fundy. Prest (1970, p. 710) proposed that the declining marine limits were due to the gradual uncovering of the shore by an eastward retreating ice lobe in the Minas Basin. Scott and Medioli (1980a) showed that the elevation of marine deposits in lakes also decreases between Saint John and Sackville, New Brunswick, with marine maximum at +75 m (246 ft) near Saint John and +10 m (33 ft) near Moncton. These authors proposed a late ice model (Scott et al., 1987; Fig. 13) based on the anomalous RSL curves and emergence in the area. The late ice cap is centred over the tip of the Chignecto Peninsula where local emergence is greatest. However, Stea et al. (1986) have shown that the tip of the peninsula was the first area to become deglaciated. An end moraine system, north of the tip of the peninsula, marks the margin of the last glacier which receded up Chignecto Bay (Fig. 6). As well, ice receded into the Minas Basin. Westward trending striations near Spencers Island imply that the westward flow of ice out of the Minas Passage (Ice Flow Phase 4; Fig. 5) reached well into the deeper parts of the Bay of Fundy.

The maximum ice load model cannot explain the decline of marine limit in Chignecto Bay. The axis of the bay is subparallel to the general trend of isobases inferred by the ice load model

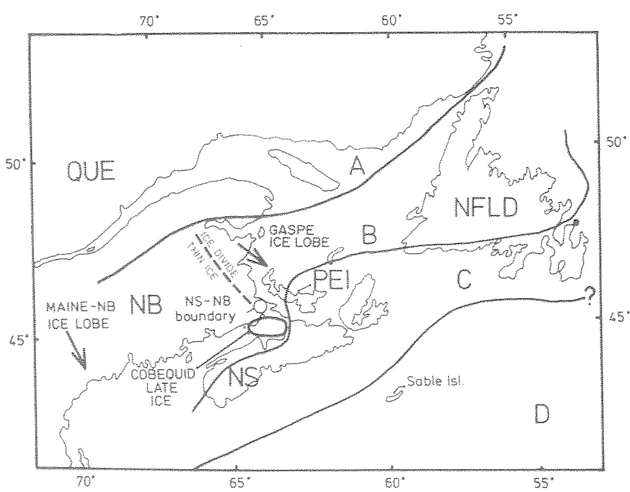


Figure 13. Map of the Maritime region indicating ice lobe positions for the Late Pleistocene and sea level zones

so there should be little differential uplift along its length. The emergence data, however, shows a deflection of isolines to the northwest so that they are perpendicular to the axis of the bay and to the direction of ice flow during the Ice Flow Phase 4 (Fig. 12). This implies that marine incursion was delayed as this ice lobe retreated up the bay and inland. Rampton and Paradis (1981a) also explain discordancies in marine limits in New Brunswick by the persistence of residual ice caps.

Perturbations in the 40 m (131 ft) isoline along the North Mountain shore (Fig. 12) are a result of the 20 m (66 ft) variation in elevation between deltaic deposits and nearby beaches. This variation may be due to a glacier advance. Hickox (1962) proposed a glacier readvance (ca. 11,000-10,500 yrs B.P.) over the Margaretsville delta (Stop 2; Day 2). The readvance may have interrupted emergence before the formation of the raised beach after deglaciation. Grant (1980) observed similar displacements of shorelines due to glacier readvances in Newfoundland.

It is apparent that the marine-limit plane defined by deltas along the Maine shoreline (Fig. 12) has been isostatically uplifted in response to deglaciation. Greater uplift occurred to the northwest, where the late Wisconsinan ice sheet was thicker. The tilt, as measured across the isolines in central Maine, is 0.49 m/km (2.60 ft/mi).

Figure 12 shows an area of anomalously low deltas in the vicinity of Passamaquoddy Bay. It is likely that this area experienced Holocene crustal subsidence, especially in view of geodetic leveling data and other evidence for recent downwarping (Anderson et al., 1984). An alternate possibility is that a late-glacial ice mass lingered in eastern Maine while isostatic uplift was in progress. The result of the later scenario would have been that deltaic sedimentation persisted longer in the Passamaquoddy Bay region as RSL was falling, and thus the eastern Maine deltas would not be at lower altitudes than the deltas in the central to southwestern coastal lowland. There was, in fact, a major standstill of the ice margin (of uncertain duration) along the Pineo Ridge moraine system. However, deltas that formed contemporaneously along this moraine system and were graded presumably to the same RSL are now lower to the east. This fact, as well as the abruptness of the flexure of the isolines, seems to favor a neotectonic explanation for the low altitudes of the eastern deltas.

HOLOCENE SEA LEVELS (10,000 yr B.P. to present)

Sea level data is minimal for the time prior to 7000 yr B.P. Milliman and Emery (1968) report sea levels of -40 m (131 ft) in Georges Bank at 10,000 yr B.P. Sea level reverse points were dated at 7000 yr B.P. by Scott and Greenberg (1983) and Scott and Medioli (1982).

Much of the coastline was emergent before 7000 yr B.P. Since 7000 yr B.P., RSL has been rising in most of southeastern Canada at varying rates which are dependent upon the position relative to the collapsing forebulge (Fig. 10). Where RSL data dates back to 4000 yr B.P., there is a break in the rate of RSL rise at 2500 yr B.P., as reported for New England (Redfield, 1967). Rates of RSL rise prior to 2500 yr B.P. can be as high as 1 m/century (3.28 ft/century). However, rates after 2500 yr B.P. are usually less than 20 cm/century except at the edge of the continental shelf where rates do not appear to have changed since 7000 yr B.P. (Scott et al., 1984).

SYMPOSIUM FIELD TRIP

ITINERARY

Day 1 - July 20, Monday

Leaders: Stea, Scott, Boyd, Douma

Fundy Tide (09:30) HIGH
(WINDSOR) (15:40) LOW

Time	Event
0800	muster at Dalhousie University in Halifax; travel to Wolfville
0845-1015	Stop 1-1
1045-1130	Stop 1-2
1145-1330	Stop 1-3
1415-1430	Stop 1-4
1450-1600	Stop 1-5
1615	arrive in Wolfville; stay at Acadia University

Day 2 - July 21, Tuesday

Leaders: Stea, Scott

Fundy Tide (10:25) HIGH
(WINDSOR) (16:35) LOW

Time	Event
0810-0820	Stop 2-1
0850-1000	Stop 2-2
1130	arrive at Ferry Terminal in Digby, Nova Scotia
0300	board ferry to Saint John, New Brunswick; depart immediately for Michias, Maine
1700	arrive and stay at University of Maine in Machias

Day 3 - July 22, Wednesday

Leaders: Kelley, Kelley

Fundy Tide (10:00) HIGH
(MACHIAS) (16:05) LOW

Time	Event
0800	muster at University of Maine
0820-0930	Stop 3-1
0945-1030	Stop 3-2
1045-1300	Stop 3-3
	Stop 3-4
	Stop 3-5
1400-1500	Stop 3-6
1700	arrive at Shiretown Inn in St. Andrews, New Brunswick

Day 4 - July 23, Thursday

Leaders: Seaman, Nicks

Fundy Tide (11:05) HIGH
(SAINT JOHN) (17:15) LOW

Time	Event
0800	muster at St. Andrews
0830-0915	Stop 4-1
0100-1300	Stop 4-2

... continued

1315-1330	Stop 4-3
1400-1500	Stop 4-4
1700	arrive Mount Allison University, Sackville, New Brunswick

Day 5 - July 24, Friday

Leaders: Seaman, Scott, Stea
Fundy Tide (12:30) HIGH
(MONCTON) (18:30) LOW

Time	Event
0800	muster at Mount Allison University
0830-0915	Stop 5-1
1000-1100	Stop 5-2
1115-1130	Stop 5-3
1200-1400	Stop 5-4
1430-1600	Stop 5-5
1630	return to Mount Allison University

Day 6 - July 25, Saturday

Leaders: Stea, Wightman, Finck, Scott
Fundy Tide (13:25) HIGH
(JOGGINS) (19:30) LOW

Time	Event
0800	muster at Mount Allison University
0845-1030	Stop 6-1
1100-1300	Stop 6-2
1315-1430	Stop 6-3
1500-1530	Stop 6-4
1545-1630	Stop 6-5
1700-1730	Stop 6-6
1800-1830	Stop 6-7
1900	arrive in Truro; stay at Palliser Motel

Day 7 - July 26, Sunday

Leader: Stea

Time	Event
0930-1000	Stop 7-1
1030-1200	Stop 7-2
1300-1430	Stop 7-3
1700	arrive in Halifax

DAY 1 - CENTRAL NOVA SCOTIA (JULY 20)

Stop 1-1. Lantz clay quarry

Leader: R. R. Stea

Purpose: To examine evidence of a late-glacial climatic oscillation equivalent to the Allerod/Younger Dryas in Europe

Route: Leave Dalhousie University, Halifax at 0800 hr; take the A. Murray Mackay bridge to Highway 118 and merge with Highway 102; take Exit 7 to Enfield and proceed east on Highway 2 for approximately 10 km (6.2 mi); arrive at 0845 hr.

Introduction:

The first stop of the trip will be in the clay quarry of the L. E. Shaw brick plant in Lantz (Fig. 14). 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 purpose of this stop is to examine evidence of a late-glacial climatic oscillation equivalent to the Allerod/Younger Dryas event in Europe. A buried peat bed of boreal-tundra climatic affiliation, dated at 10,900 - 11,700 yr B.P., was found at this site. This site is one of 15 documented sites of this age in Nova Scotia (Mott et al., 1986). Many of these sites are found along the coastline but none of the sediments in the exposed sections are known to be marine. There is a lack of RSL data for this time period, but some indirect evidence of sea levels at this time does exist. Indians occupied a site at Debert (Fig. 2) approximately 10,585 yr B.P. near the end of this warm interval. Land migration routes to this site may have been blocked by residual ice. An attractive theory proposed by Borns (1966) suggests that the Minas Basin and large parts of Chignecto Bay were land during the 11,000 - 10,000 yr B.P. period and utilized by the Indians as a migration route from New England. The chalcedony used in the Debert projectile points is found along the Minas Basin shore near Five Islands (Fig. 2; Macdonald, 1968).

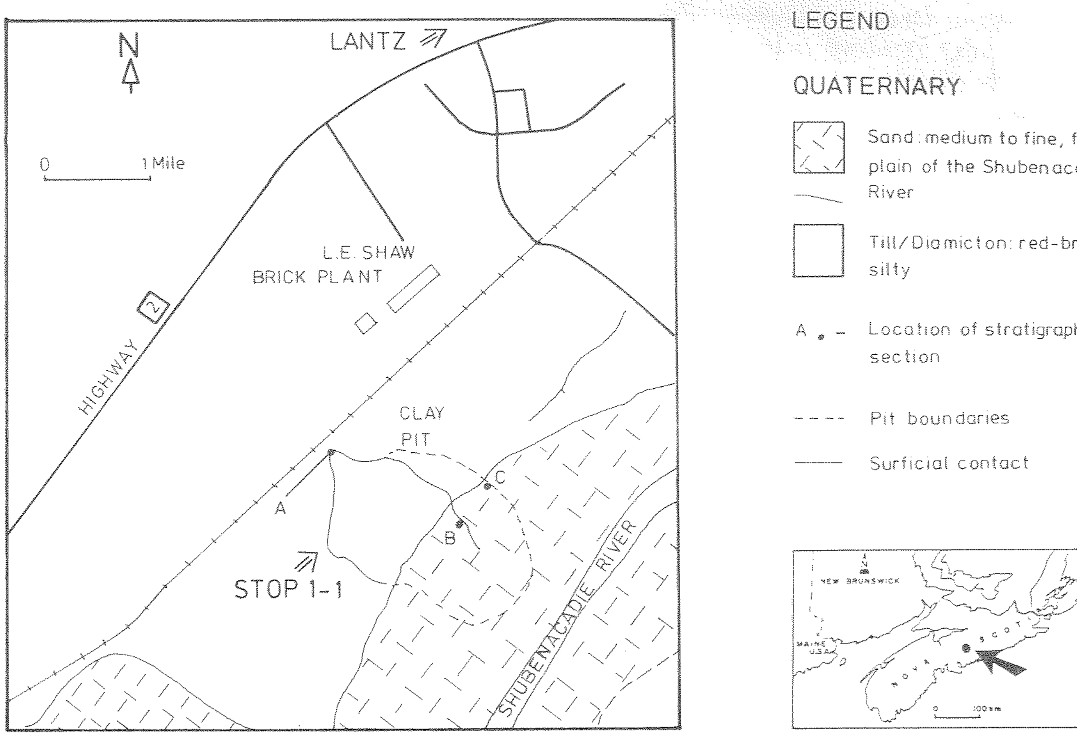


Figure 14. Location and surficial geology of the Lantz site (Day 1, Stop 1). A, B and C are section locations shown in Figure 15

Site Description:

The Lantz site is located in the clay quarry of L. E. Shaw Limited in Lantz, Halifax County (Fig. 14). The site is part of the Hants - Colchester Lowland region which is underlain by Carboniferous and Triassic rocks (Figs. 2 and 3). The lowlands are bounded on the south by the Atlantic or Southern Uplands and on the north by the Cobequid Highlands. The site is underlain by Early Viséan Windsor Group rocks including gypsum and limestone. The clay quarry area is located in a gently undulating plain which lies above the flood plain of the Shubenacadie River. The tidal surge reaches up river as far as the village of Shubenacadie.

The stratigraphy of the site was gleaned from pits that have been exposed during the last 5 years. Three sections in the clay quarry area are shown in Figure 15. The first section (section A) is part of a 20 m (66 ft) wide swale. The units described in the section all pinch out at the edge of the swale and grade into sand which immediately overlies the quarried clay.

Unit 1 is a massive clay to rythmically laminated greyish brown clay-silt. Calcareous concretions are commonly found weathering out of exposed clay blocks. Unit 1 underlies most of the area of the brick plant to a depth of 10 m (33 ft). Auger holes reveal that sand and gravel underlie the clay to depths of 5 m (16 ft). Rootlets were found in the upper metre (3.3 ft) of the clay unit but these did not continue through the underlying sand. This unit is presently being quarried.

Unit 2 is a grey to brownish sand which becomes oxidized toward the contact with Unit 3. A placon is developed in some areas at the top of Unit 2.

Unit 3 is locally gravelly at the base with a discontinuous greenish clay zone, which in turn is overlain by a peat layer of variable thickness (0-0.3 m; 0-1.0 ft). In observed sections, the peat zone pinches out at the edge of the swale. A bulk sample of the peat was dated at 11,100 ± 100 yr B.P. (GSC - 3116). Samples of the bottom and top of the peat zone were dated at

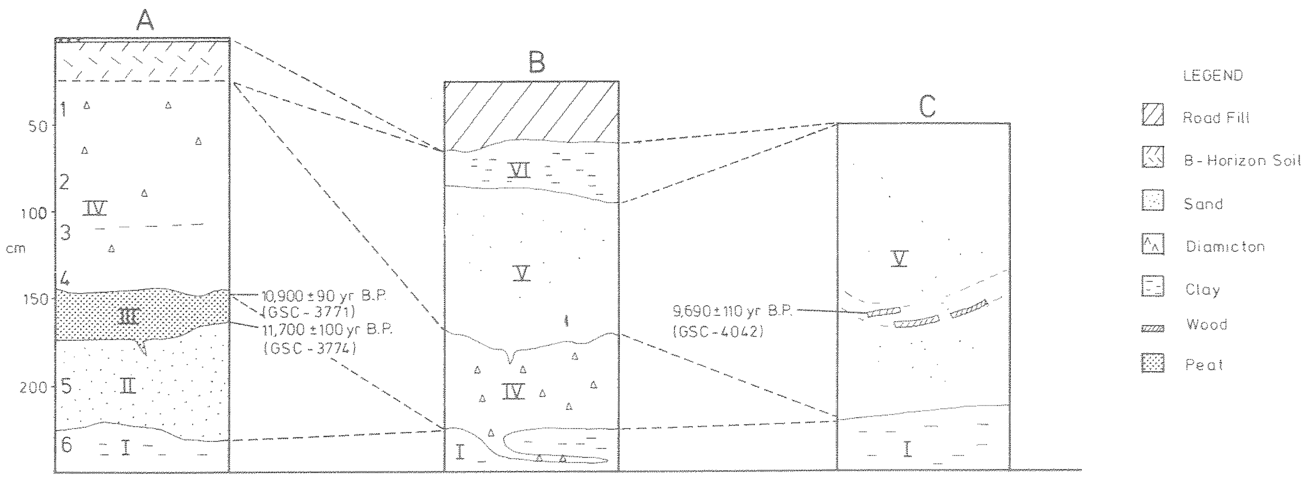


Figure 15. Stratigraphy of 3 sections exposed at the quarry

11,700 ± 100 yr B.P. (GSC - 3774) and 10,900 ± 90 yr B.P. (GSC - 3771) respectively.

Unit 4 is a grey, blocky clay which becomes reddish with higher sand and stone percentage toward the contact with the overlying peat. This deposit constitutes the bulk of the constructional landform which contains the peat. This reddish diamict is correlated with a sediment at section B which has an increased stone and sand content, and overlies and intrudes Unit 1.

Unit 5 occurs at the southern end of the quarry nearest the Shubenacadie River. It is a buff, gravelly sand with reddish clay layers and overlies the reddish diamict. The gravelly sand is capped by a reddish clay layer at section B. Organic clay lenses and wood fragments were found in arcuate channel-fills within the unit. A wood (Populus sp.) fragment from these organic layers was dated at 9690 ± 110 yr B.P. (GSC - 4042).

The environment of deposition and the paleoclimatic interpretation of these units is summarized as follows:

Unit 1 - the clay at the site is believed to have a glacio-lacustrine origin based on the presence of concretions, and banded or rythmite bedding, and the lack of fossils. The clay is located throughout the Shubenacadie Valley and adjacent lowland regions (Fig. 2). At some locations, the clays are rythmically bedded and contain large dropstones. During regional deglaciation (ca. 14,000 yr B.P.), a large lake formed due to the blockage of northward drainage by residual ice. Lacustrine deposits, comprising mainly clays, formed in the ice-dammed lake.

Units 2 and 3 - the ice dam was breached during the ensuing warm period (ca. 11,500 yr B.P. - 10,500 yr B.P.) which lasted 500 - 1000 yr. The pollen profile of the Lantz site indicates that it was a bog environment. Shrubs, notably sedge (Cyperaceae) and birch (Betula) migrated to the site followed by spruce trees (Picea).

Unit 4 - the interpretation of this sediment is contentious and has regional climatic implications. The unit is not well exposed in the quarry area. It is clay-rich in some parts of the quarry but stonier in others. Pollen within the unit suggests that it may be a lake or flood plain sediment (Mott et al., 1986). Evidence of intrusion and deformation of the sediments beneath it suggests that it may be glacigenic (Fig. 16). Recently, till-like diamictons were found overlying peats of the same age in other parts of the Province. If the deposit is a till, then a glacial interval truncated the warm period when the peat was deposited.

Unit 5 - this unit represents a flood plain deposit.

Stop 1-2. Selmah Bar

Leaders: R. Boyd, M. Douma, D. B. Scott

Purpose: To examine sedimentary features formed in the macrotidal environment of the Bay of Fundy

Route: Leave Lantz at 1015 hr and proceed east along Highway 2 for 7 km (4.3 mi); turn left onto Highway 14; rejoin Highway 102 at Exit 9; take Exit 10 to Highway 214 and proceed north for 25 km (15.5 mi); arrive at 1045 hr.

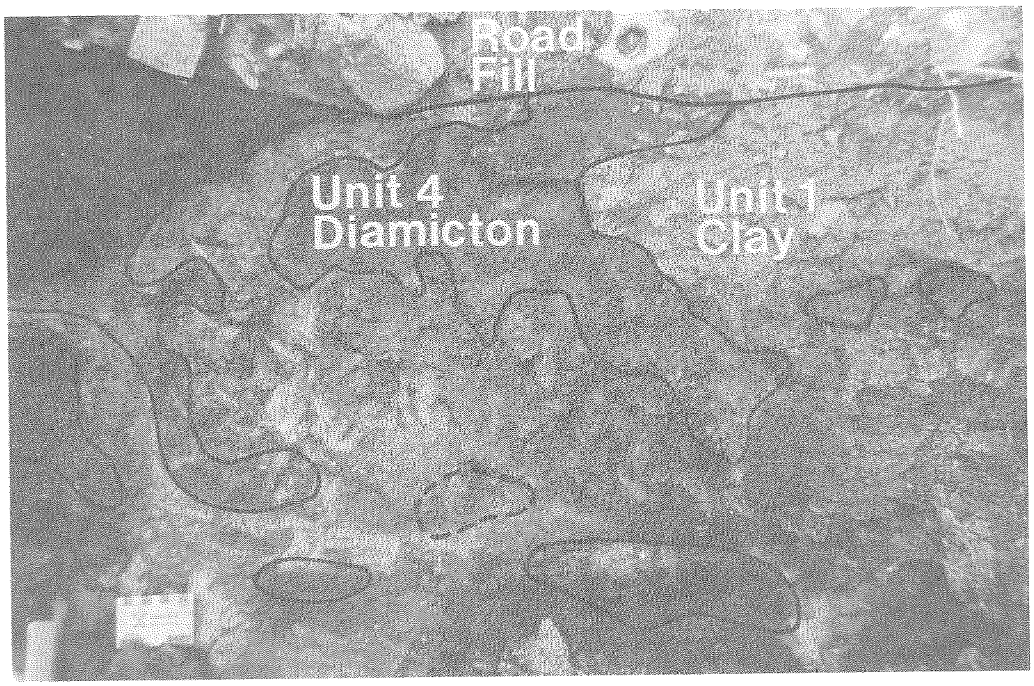


Figure 16. Photograph of clastic wedge of Unit 4 intruding Unit 1

Introduction:

We will be driving along the Shubenacadie River as we approach Stop 1-2 (Figs. 2, 17). The name "Shubenacadie" is derived from the Micmac Indian word "place where wild potatoes grow". The river winds its way through the Hants-Colchester Lowlands underlain by Carboniferous clastic and evaporitic rocks. The tidal bore can be seen from the cantilevered bridge at South Maitland (Fig. 2). Drumlin exposures along the lowlands of the Shubenacadie River form spectacular cliffs that are constantly undercut by the powerful tidal surge travelling up the river. Three till units are exposed in the section. The orientation of the drumlins and striations indicate a major period of southward ice flow (Ice Flow Phase 2; Fig. 5). The terrain fabric and surface outcrops of most of this region, however, relate evidence of two subsequent ice flows, to the northeast (Ice Flow Phase 3) and to the west (Ice Flow Phase 4; Fig. 5).

This stop is a very popular site. Many of our classes as well as previous conferences have included it as part of the program. The Selmah Bar is one of the classic 'mega' features formed as a result of the expanded Fundy tidal range. The feature has been studied intensively since 1971 due to its easy accessibility and excellent exposure.

WARNING: Depending on the state of the tide, it is possible to walk to the outer edges of this feature. However, we must caution you not to put any large channels between yourself and the shore since the tide floods rapidly west beginning 1.5 - 2.0 hr after low water level. If you become cut off, it is a difficult swim in very cold water and fast currents as the bar is entirely submerged at high tide.

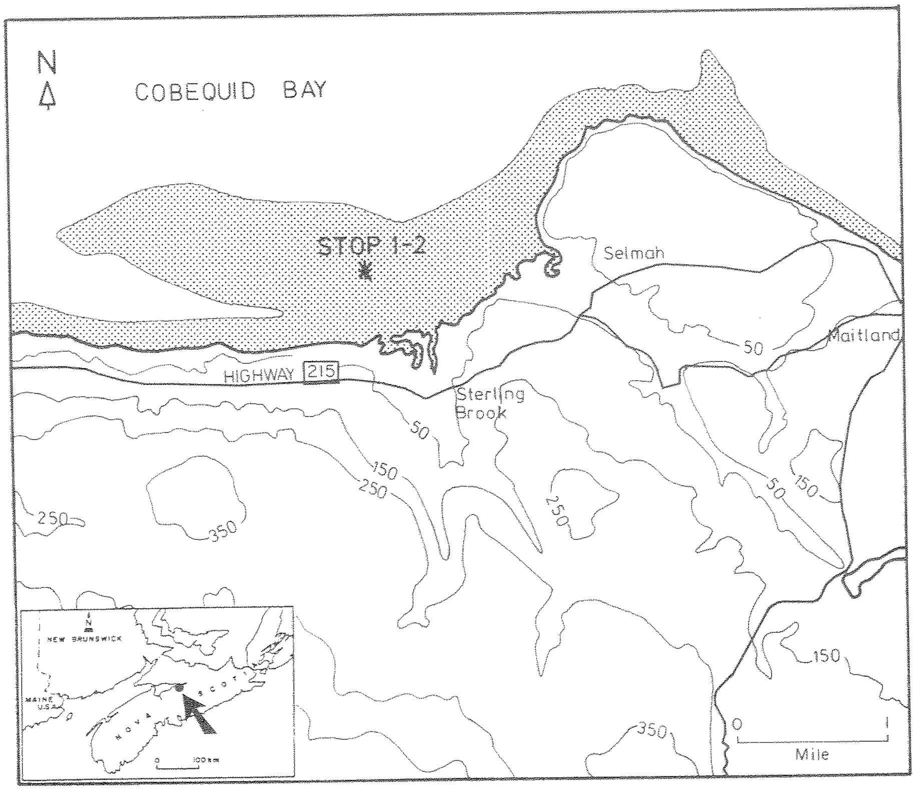


Figure 17. Location or Selmah Bar (Day 1, Stop 2)

Site Description:

We will briefly summarize some of the highlights. The shore cliffs near Selmah Bar range 2-10 m (7-33 ft) in height and are mostly till with some red Triassic sandstones and shales at Salter Head. There are some fringing salt marshes but we will see much better examples of these high tidal range marshes at later stops.

Gravel beaches, mud flats and a wave-cut platform face out into the bar between the coastal cliffs and the bar.

Selmah Bar:

The bar is a large, depositional sand body measuring 4 km (2.5 mi) long and up to 1.5 km (0.9 mi) wide with a relief sometimes exceeding 6 m (20 ft). The feature we will see occupies the southeastern most portion of the large complex of similar sand bars that fill Cobequid Bay west of Salter Head (Fig. 17). The general size and shape have remained relatively constant since the earliest records (Knight, 1980; Amos and Long, 1980).

The topography of the bar is dominated by a large swatchway (Robinson, 1960) cut diagonally across the feature. The cross-section of the entire area is generally asymmetrical. The symmetry is controlled by the flood tide component.

Sediments:

The sediment that makes up Selmah Bar is predominantly well sorted medium sand with coarser sand in areas adjacent to the foreshore and finer sands on the eastern edge of the bar. There are isolated patches of gravel and boulders as a result of ice rafting.

Features to Observe:

What distinguishes this bar from other tide-dominated estuaries is the tidal range. This allows us to closely examine the bedforms at low tide (Fig. 18). Dalrymple et al. (1978; 1982) recognize four morphologically and hydraulically distinct types:

1. Ripples: height <0.05 m (0.16 ft), wave length <0.3 m (1.0 ft), straight to linguoid in plan (Fig. 18),

2. Type 1 megaripples: height 0.05-0.5 m (0.16 - 1.6 ft), wave length 0.2-5.0 m (0.65-16 ft), relatvely flat profiles,

3. Type 2 megaripples: height 0.05-0.7 m (0.16 - 2.3 ft), wave length 0.05-14.0 m (0.16-50 ft), steep profiles, and

4. Sandwaves: height 0.15-3.4 m (0.5-11 ft), wave length 5-215 m (16-705 ft), may have first three types incorporated within them.

At low tide, ripples exist almost everywhere superimposed on larger features. Types 1 and 2 megaripples occur on most parts of the bar, either alone where grain size is finer or superimposed on sandwaves where sediment is coarse. Sandwaves occur along the south side of the bar wherever grain sizes are larger than 0.31 mm (0.01 in).

Stop 1-3. Tennycape Quarry

Leaders: R. R. Stea, P. W. Finck

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

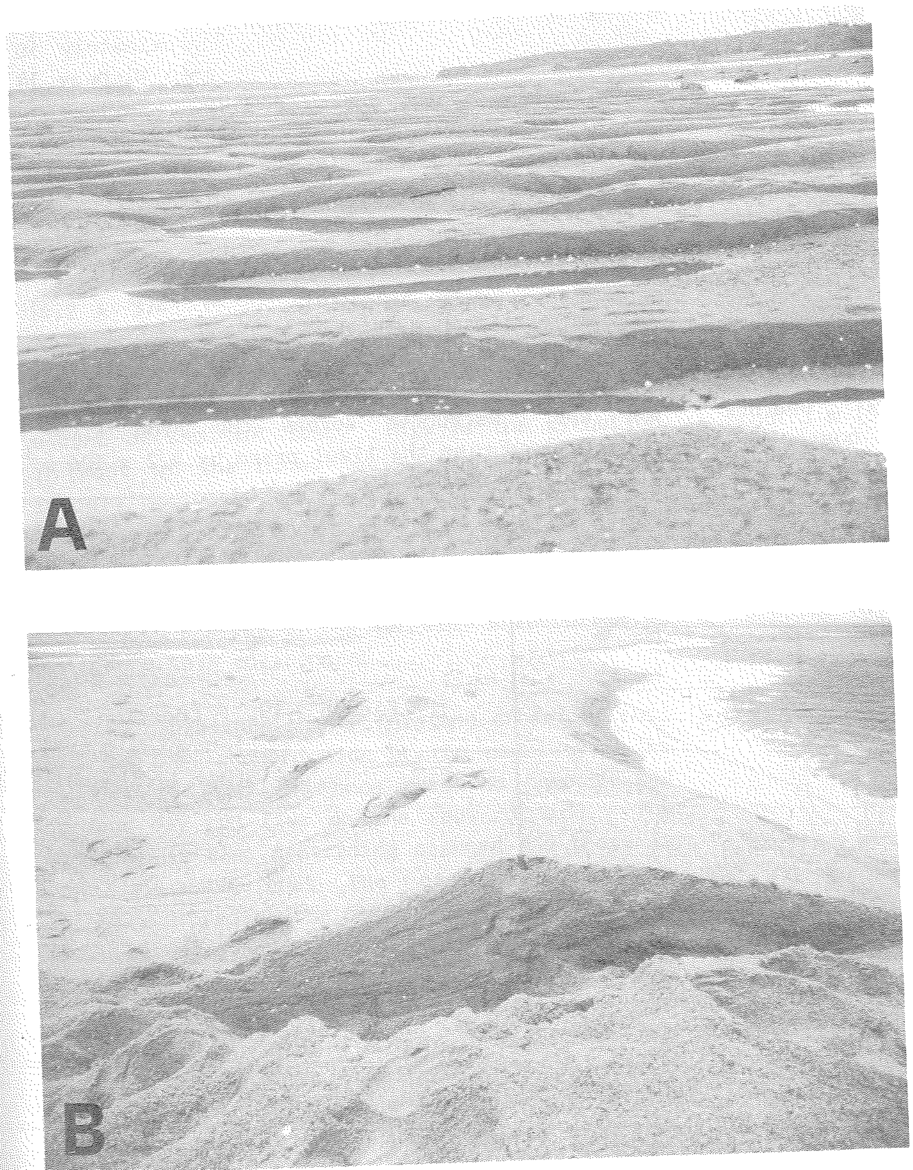


Figure 18. Photographs of the tidal channel and bar facies at Selmah Bar. A - Type 2 megaripples on the surface of Selmah bar, B - cross-section of Type 1 megaripple showing flood dominated cross-stratification at the base of the trench and ebb tide reworking at the crest.

Route: Leave Selmah Bar at 1130 hr and proceed westward along Highway 215 for 20 km (12 mi); at Tennycape bridge turn left and travel along dirt road to quarry; arrive at 1145 hr.

Introduction:

The site at Tennycape Quarry (Fig. 19) reveals evidence of most of the ice flows that have affected the area during the Wisconsinan Stage. The sequence of the last two ice flows is also revealed at a location east of this area where a flat siltstone surface bearing striations (trending 247°) crosscut wide grooves (trending 020°) stained with iron oxides. The preferential staining of the grooves may indicate an ice-free period of subaerial weathering. Large east-west oriented esker systems in the Shubenacadie region have predominantly west trending paleocurrent indicators (P. Watson, personal communication, 1983).

Site description:

All regional flow trends can be seen at the Tennycape Quarry site (Fig. 5). A westward sloping bedrock surface of Carboniferous siltstone is exposed at the quarry. The bedrock is buried under 8 m (26 ft) of till(s). The lower surface reveals striations trending 138° cut by striations trending 180° (Fig. 20). Bedrock surfaces in the upper part of the quarry are inscribed with two sets of striations, one trending 028° truncated by a second trending 281°. Pebble 'shadows' or 'mini' crag and tail features on conglomerates at nearby localities also indicate evidence of the same two ice flows. A boulder of porphyritic granite embedded in the upper part of the till section at the quarry is believed to have been emplaced by a flow which crossed a section of the batholith to the southwest (Fig. 3).

Stop 1-4. Windsor Mud Flats

Leader: D. B. Scott

Purpose: To show the effect of a dam across a major tidal channel

Route: Leave Tennycape at 1330 hr and proceed along Highway 215 for approximately 50 km (31 mi); turn right onto Highway 14 and travel west for 7 km (4 mi); turn right onto Highway 101 and travel northwest for 4 km (2.5 mi); arrive at 1415 hr.

Introduction:

This will be a relatively short stop to look at a recently formed mud flat (Fig. 21). In 1970, a 900 m (2953 ft) long, impermeable causeway was constructed across the Avon River, directly reducing the tidal prism by 4.2 x 106 m^3

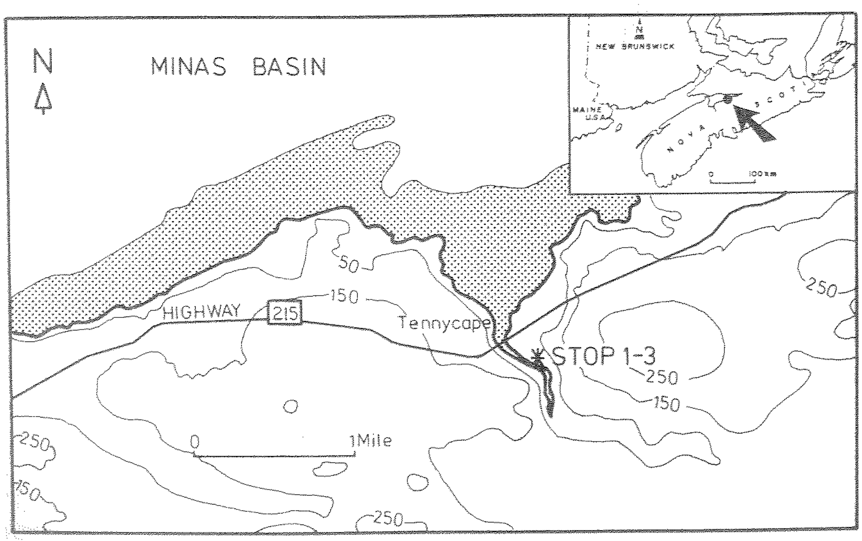


Figure 19. Location of Tennycape Quarry (Day 1, Stop 3)

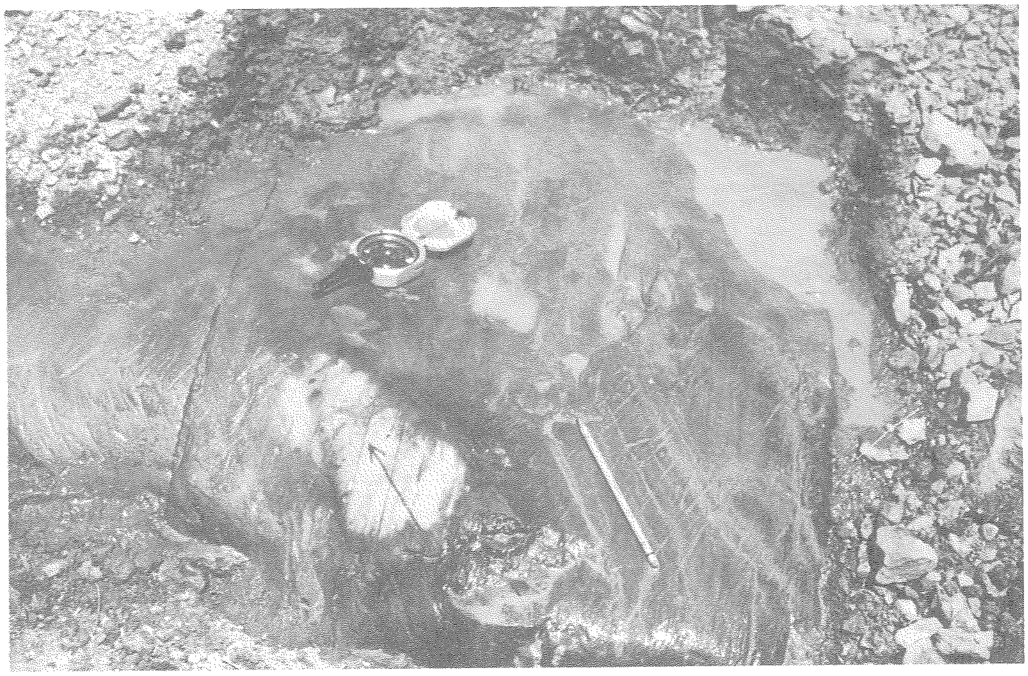


Figure 20. Photograph of the striated bedrock surface at Tennycape Quarry; ice flow to the northeast (028°) marked by pencil and ice flow to the west (281°) marked by compass.

(4.2 x 3743 ft^3). This, of course, caused a substantial decrease in tidal currents and turbulence. If you look closely at the water you will notice it is virtually opaque from the large amount of sediment load. This is the situation in most of the Minas Basin. The sediment is largely derived from the soft bedrock cliffs in the area. Any decrease in the hydraulic regime will cause a dramatic increase in sedimentation rate. The Windsor mud flat is the most pronounced example of how man-made structures can influence this environment.

To appreciate the 'before and after' effect, look on the landward side of the causeway. The channel is deep and once extended through the present day mud flat on the seaward side. The mud flat is also prograding seaward and there is some evidence that the causeway may have slightly decreased the tidal range. Studies here indicate rapid accretion in the summer (mean rate 5 cm/month; 2.0 in/month; maximum 14.6 cm/month, 5.7 in/month) which is partially balanced by ice plucking in the winter (Amos, 1978). The mud flat appears to have stabilized at about 4 m (13 ft) of mud with its surface 3.5 m (11.5 ft) below HHW (Amos, 1978). Salt marsh plants (*Spartina alterniflora*) have already begun to colonize its surface and once these take hold, they will promote another rapid growth phase until the marsh surface reaches the high marsh phase (i.e. about 1 m (3.3 ft) below HHW).

Stop 1-5. Wolfville

Leaders: S. Bleakney, D. B. Scott

Purpose: To examine buried oysters in a Holocene section near Wolfville

Route: Leave Windsor 1430 hr and travel northwestward along Highway 101 for 18 km (11 mi); take Exit 10 and proceed along side road for 2 km (1.2 mi); turn right towards Evangeline Beach and travel north for 4 km (2.5 mi); arrive 1450 hr.

Introduction:

Evangeline was the fictional heroine of the poet Longfellow in the poem of the same name. The poem depicts the plight of the Acadians French settlers who colonized the Minas basin region in the 1600s and were deported from their

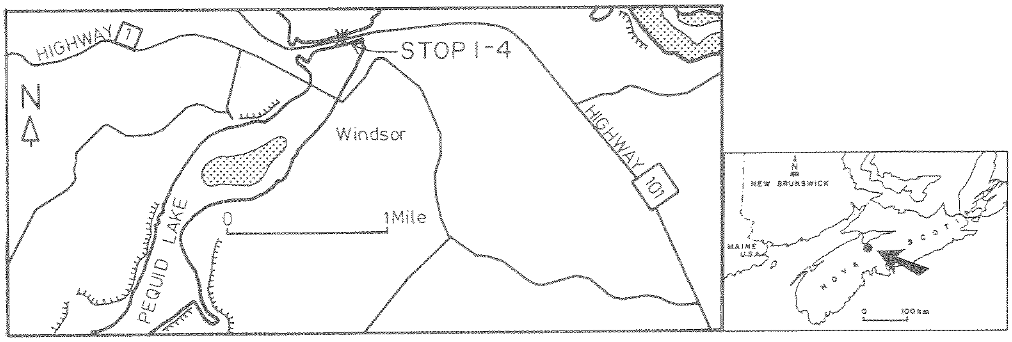


Figure 21. Location of Stop 4 of Day 1

homelands by the British in 1755. The majority of the extensive dykelands in the Minas Basin area was built by the Acadians.

This site (Fig. 22) is famous for both its interest as a ancient level site and for the extraordinary size of the oysters found there (Fig. 23). Although at 4000 yr B.P. the climate in Nova Scotia was considerably warmer, these oysters do not live this far north at present.

To be able to see everything, we require a spring low tide. Unfortunately, this situation will not exist today. However, we will have some examples of the oysters and you will be able to see the general area.

The significance of the oysters is that the upper limit of their habitat is LLW level. In combination with dates of tree stumps, which indicate the HHW position, measurement of the paleotidal range is possible (Bleakney et al., 1981). It is suggested that at 3800 yr B.P. the tidal range was 4-5 m (13-16 ft) compared with the present 16 m (53 ft) range. The former is about 30% of the latter. These figures contradict the Scott and Greenberg (1983) calculations of paleotides for 4000 yr B.P. which were about 82-88% of the present range (i.e. about 13.14 m; 43.11 ft).

We will not try to resolve this conflict here but let the participants in the field trip come up with their own ideas.

DAY 2 - NORTH MOUNTAIN AND ANNAPOLIS VALLEY (JULY 21)

Stop 2-1. Turner Brook

Leaders: R. R. Stea and D. B. Scott

Purpose: To view an emerged strandline and the remnants of a wave-cut cliff along the North Mountain cuesta

Route: Leave Acadia University, Wolfville at 0730 hr and travel east along Main Street for 3 km (1.9 mi); turn left onto Highway 101 and proceed westward for 30 km (18.6 mi); turn right onto Highway 360 and travel north for 14 km (8.7 mi); arrive at 0810 hr.

Introduction:

A veneer of beach deposits mantle the shore of the North Mountain below elevations of 20 m (66 ft; Figs. 2, 24). MacNeill (1956) and Hickox (1962) mapped these deposits and assigned varying elevations for emerged strandlines. Hickox stated that RSL was never higher than 20 m (66 ft) in his map area. However, MacNeill (in Prest et al., 1972) noted that the elevation of the Margaretsville delta (Day 2, Stop 2) suggested a much higher stand of sea level (45 m; 148 ft). Recent mapping by the authors revealed an increase in strandline elevations from northeast to southwest along the North Mountain. The strandline increases in elevation from 20 m (66 ft) to 38 m (125 ft) above mean sea level from Scots Bay to Hampton along the North Mountain (Fig. 12). There is a discontinuity of 20 m (66 ft) between the RSL indicated by the delta at Margaretsville and the nearby strandlines. Hickox postulated that ice advanced over the delta. The RSL discontinuity is probably a result of rebound that occurred when ice covered the delta before the formation of the strandline. Grant (1980) postulated that glacier readvance in

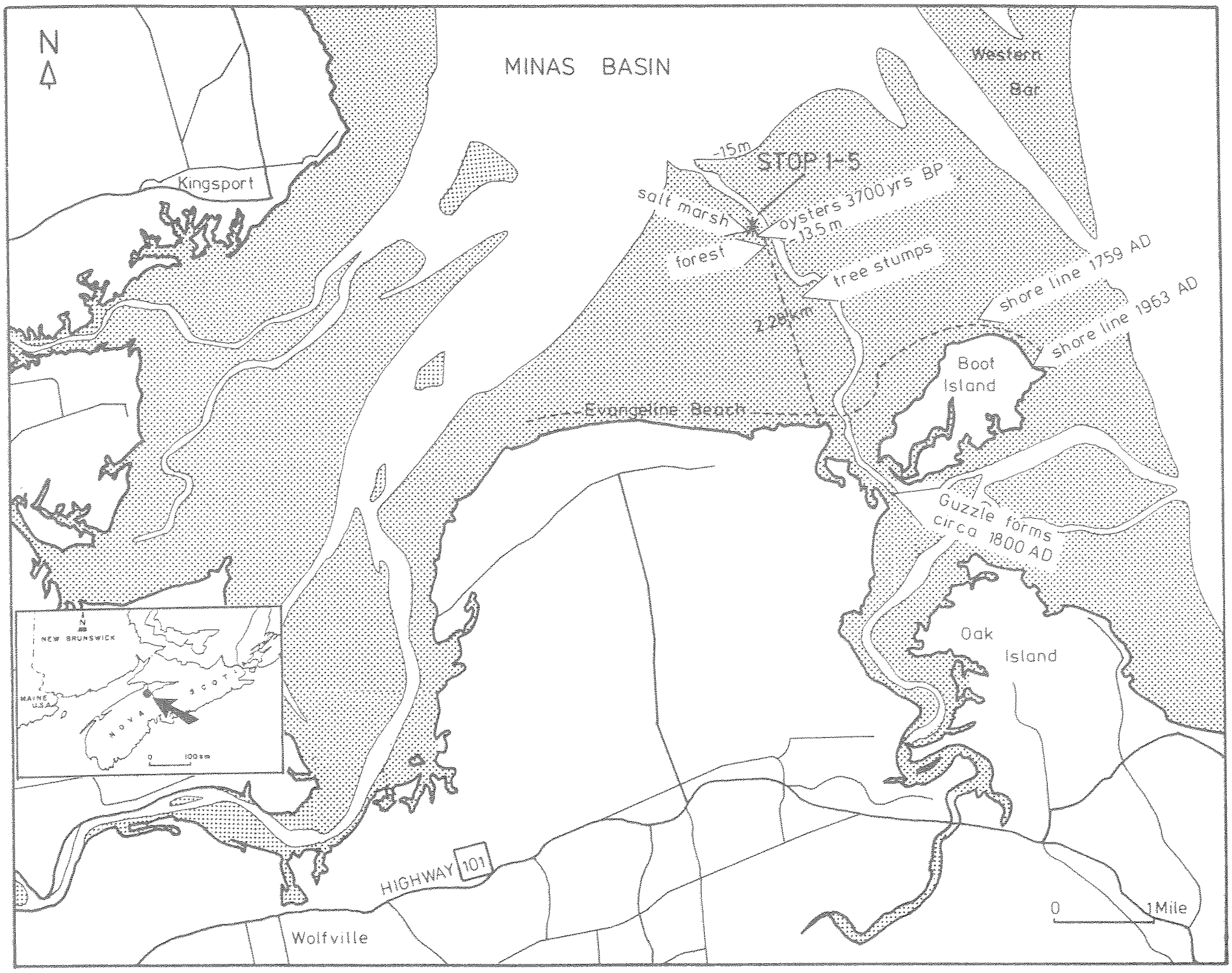


Figure 22. Location of Stop 5 of Day 1

Newfoundland created similar discontinuities in emergence.

Site Description:

The Turner Brook site reveals a wave-cut bench and cliff eroded in till, with a thickening wedge of beach deposits (Fig. 25). The emerged sea cliff is 14 m (45 ft) high (Fig. 24). The beach terrace slopes seaward at 3-7° and terminates in the modern sea cliff.

Stop 2-1. Margaretsville

Leaders: R. R. Stea and D. B. Scott

Purpose: To see an emerged delta with a kame field superimposed on the delta surface

Route: Leave Turner Brook at 0820 hr and proceed south for 14 km (8.7 mi); turn left onto Highway 221 and travel 20 km (12.4 mi); turn right onto side road and travel north for 8 km (5.0 mi); arrive at Margaretsville at 0850 hr.

Introduction:

The Margaretsville delta covers 3 km (1.9 mi) of shoreline between Margaretsville and Port George (Fig. 26). It was described by Hickox (1962) whose work we will largely draw on for this stop description. A surficial geology map of the area shows a large area of ice contact stratified drift over the delta front (Fig. 26). The delta was fed by two meltwater channels that presently harbour northward flowing misfit streams. These channels end abruptly in a

V-shaped notch that cuts southward through the North Mountain to join the Annapolis River. Hickox proposed that advancing ice dammed the Annapolis valley forming a proglacial lake. This explains the lacustrine deposits in the valley. When the lake reached an elevation of 82 m (270 ft), the height of the drainage divide, meltwater spilled across the North Mountain and deposited the delta. Delta deposition occurred when RSL was 45 m (150 ft) higher than today. The ice then thickened to override the mountain, depositing kames upon retreat. MacNeill (in Prest et al., 1972) proposed a late westward flow in the Annapolis Valley. This ice may have surged over the North Mountain. Figure 27 shows the stages of formation of the delta according to Hickox.

Site Description:

The first exposure we will see is a pit cut into a kame. This is part of a kame field that rests directly on the delta surface. The kame exhibits chaotic cut and fill structures, a wide range of grain sizes between beds, and marginal slumping of beds. The delta front is exposed in cliffs 33 m (110 ft) high below the kames. The front consists primarily of sand foresets dipping 5-20° seaward.

DAY 3 - MAINE (JULY 22)

INTRODUCTION

Pleistocene glaciation was the last major geological occurrence to leave an impact on Maine. Although many glaciations probably occurred in New England during the Pleistocene, there is clear evidence for only the Late Wisconsinan glaciation in the coastal zone.

Within the area of the field trip (Fig. 1) the oldest radiocarbon dated material is seaweed from the Pond Ridge moraine dated at 13,320 yr B.P. by Stuiver and Borns (1975). This moraine (Fig. 2), which contains interbedded till and marine sediment, marks a significant grounding

Figure 23. Photo of oysters (Crassostrea virginica) found near Wolfville dated at 3800 yr B.P.

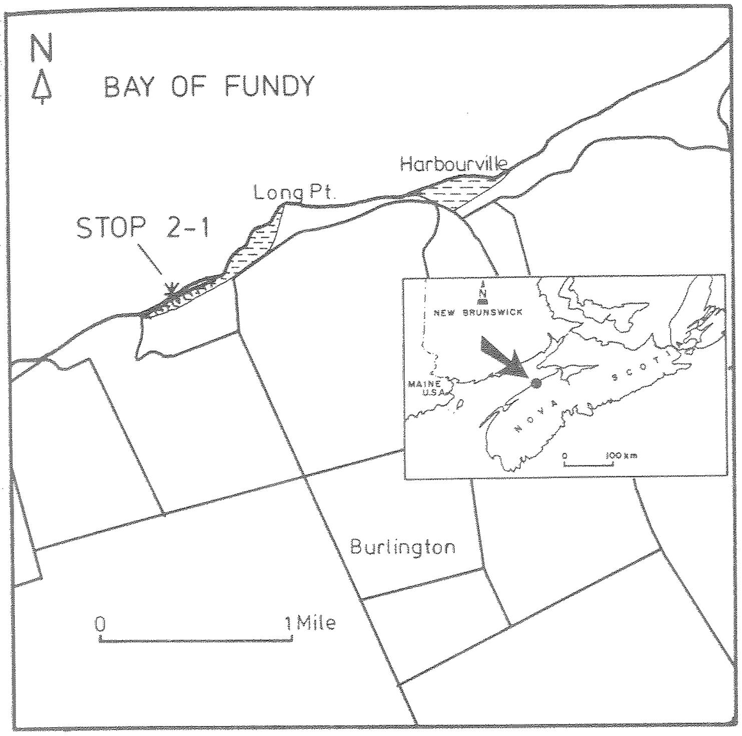


Figure 24. Location and surficial geology of Turner Brook (Day 2, Stop 1; from Hickox, 1962)

line position for a retreating, marine-based ice sheet in the Gulf of Maine. Similar stratified moraines occur elsewhere in coastal Maine and are prominent features on the State Surficial Geologic Map (Thompson and Borns, 1985). Since these features cross most of the embayments of the present coast, they also exercise an important role in the evolution of the State's estuaries by blocking embayments and providing sediment through erosion.

Following and contemporaneous with retreat of the ice, a marine submergence extended across the isostatically depressed landscape (Fig. 28). Glaciomarine deltas mark the landward extent of submergence - the marine limit. Glaciomarine mud, the Presumpscot Formation (Bloom, 1963), is common within this area (Fig. 29). The Presumpscot Formation consists of interbedded sand and mud layers with soft-sediment deformation structures at its base (ice-proximal) and more massive mud with dropstones (ice-distal) throughout a typical exposure. Fossils are locally common within the Presumpscot Formation and dates range 13,000 - 11,500 yr B.P.

An erosional unconformity marks the top of the Presumpscot Formation. It was subjected to gullying and subaerial weathering during the regression which followed its deposition (Fig. 29). In the upper Kennebec River valley, sand deposits overlying this unconformity are termed the Embden Formation (Borns and Hager, 1965). Although sand is common above the Presumpscot Formation, bog deposits may be observed as well. As yet, the Embden Formation has not been mapped outside of its type locality.

While regressive shorelines are present on the front of deltas, remarkably few such features have been mapped. It is possible that sea level dropped rapidly and few shorelines were preserved. Although the timing is uncertain, it appears that the sea level lowstand was reached at a present depth of 65 m (213 ft) around 9500 yr B. P. (Schnitker, 1974; Belknap et al., 1986a). Most evidence is from seismic reflection profiles run offshore across a conspicuous shoreline at the aforementioned depth (Kelley et al., 1986; Belknap et al., 1986b).

The landscape that was transgressed by the most recent rise of sea level in Maine was very unlike that which existed along the barrier island coastline of the central and south portions of the east coast of the United States. The drainage was deranged and all deep, preglacial bedrock valleys which entered coastal embayments were filled with till and glaciomarine sediment. The fluvially carved embayments were

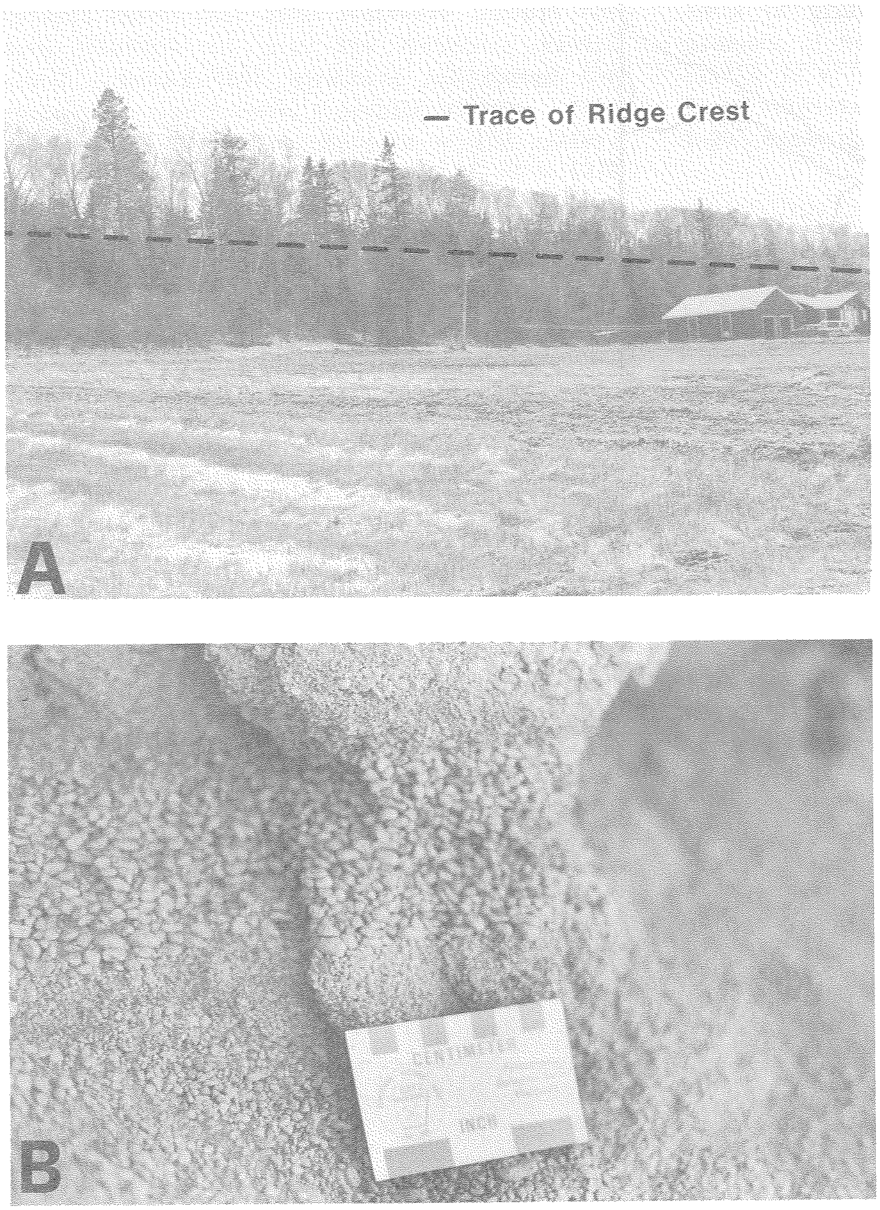


Figure 25. Photograph of the Turner Brook site. A - relict wave cut cliff in background 10 m high covered with trees. B - openwork rounded gravel in beach cut.

often blocked by moraines that similarly impeded drainage, and bogs occupied many depressions on the immature terrane.

Stop 3-1. Bar Island, Jonesport

Leaders: J. T. Kelley and A. R. Kelley

Purpose: To examine the Quaternary stratigraphic section in an 'ideal' exposure on the eroding coast

Route: Leave Machias at 0800 hr and proceed southwestward along U.S. Route 1 for 12 km (7.5 mi); turn south on U.S. Route 187 for 11 km (6.8 mi); arrive at Popplestone Beach at 0820 hr.

Introduction:

Changing environmental conditions in the Quaternary are reflected in a variety of lithologies present in the stratigraphic column in Jonesport (Fig. 29). Local bedrock, the Pleasant Bay Gabbro (Devonian), is overlain by till connecting Bar Island to the mainland. A boulder lag with some sand is all that remains of the moraine which was a peninsula within historic times.

Directly overlying the till is a well laminated series of sand, mud and organic-rich layers (Fig. 29). This is the basal, ice-proximal facies of the Presumpscot Formation. The sand layers presumably represent seasonal high energy conditions associated with ice melting. The mud and organic layers may have formed by sedimentation beneath an ice-covered water surface.

Above the well laminated unit is a more massive mud deposit containing numerous drop-stones (Fig. 29). This probably was deposited during a time when the ice had retreated considerably (ice-distal). Blocks of till that were deposited from icebergs are occasionally seen in this unit.

A discolored unit of sand and cobbles is present (arrows) at the top of the Presumpscot Formation. This is interpreted as an erosional unconformity resulting from the early Holocene sea level regression. When buried by Holocene mud offshore, the unconformity surface is a strong acoustic reflector on seismic reflection records (Knebel and Scanlon, 1985).

The 4 m (13 ft) high bluff in the Jonesport area is capped by a terrestrial peat deposit. This formed in a depression associated with the moraine which has long ago eroded. The sand at the very top of the bluff is fill associated with poorly-sited house construction.

Stop 3-2. Tracy Corner

Leader: J. T. Kelley, A. R. Kelley

Purpose: To examine a cross-section of a stratified moraine

Route: Leave Stop 1 at 0930 hr; follow Route 187 for 15 km (9.3 mi) to Tracy Corner; travel 2 km (1.2 mi) north of Tracy Corner in gravel pit; arrive at 0945 hr.

Introduction:

The Tracy Corner moraine marks one of the many ice-recessional positions inland from the present coast (Fig. 30). In plan view (Fig. 31 a,b) the moraine is asymmetric with a steep, rocky ice-proximal slope (left in Fig. 5a) and a gentle, boulder-free ice-distal side. The moraine has about 15 m (49 ft) of relief, and an active gravel pit affords a cross-sectional view of the moraine (Fig. 31b).

Several sedimentary units are well exposed in the gravel pit. A muddy gravel interpreted as till is well exposed in most walls of the pit. Near the distal side of the pit this unit over-rides (arrow, Fig. 31b) folded, stratified sand

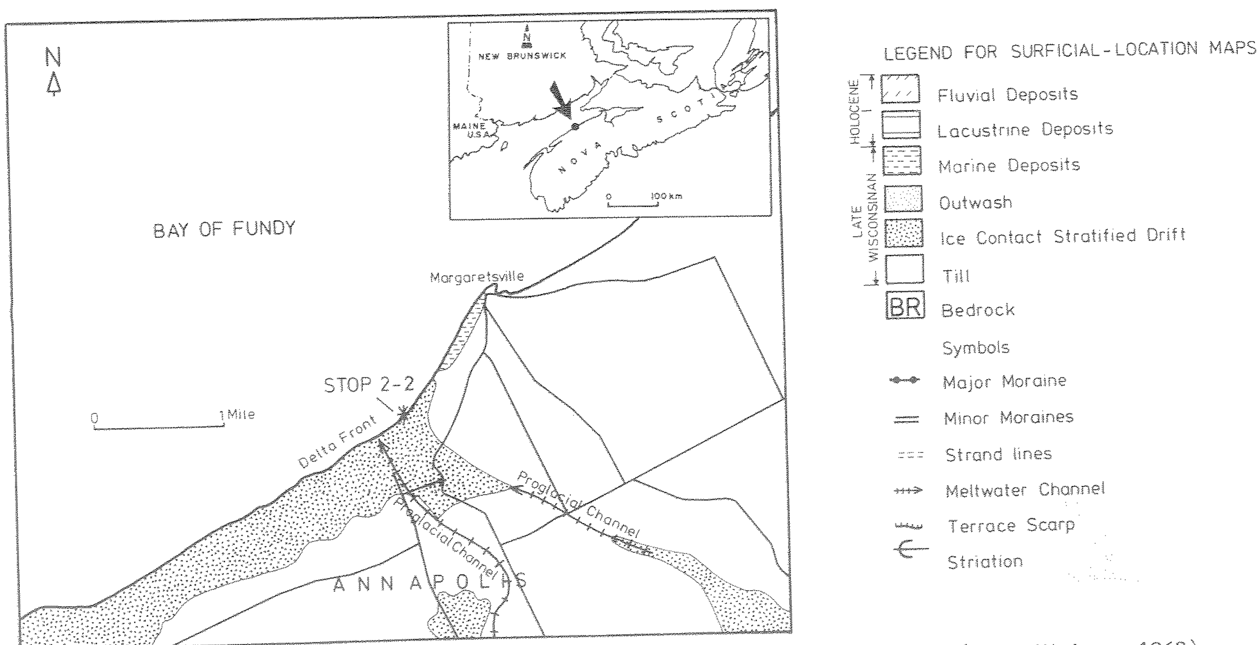


Figure 26. Location and surficial geology of Stop 2 of Day 2 (from Hickox, 1962)

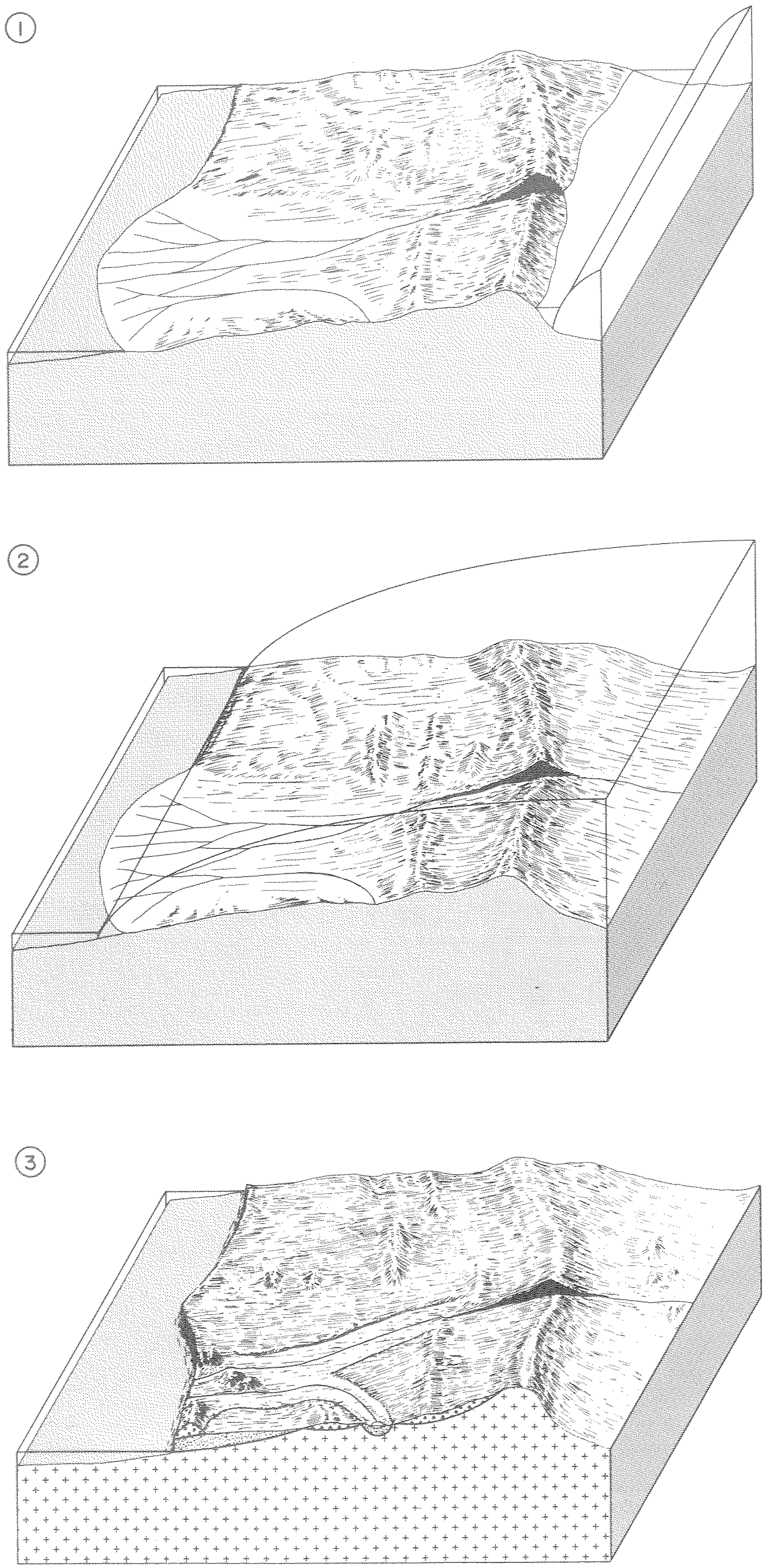


Figure 27. Stages in the evolution of the Margaretsville delta. 1 - ice in the Annapolis Valley dams a lake which drains northward through a notch and the Victoria Vale; a delta is formed where the meltwater enters the ocean at a relative sea level of 40 m, 2 - ice advances over delta and deposits kames, 3 - present day.

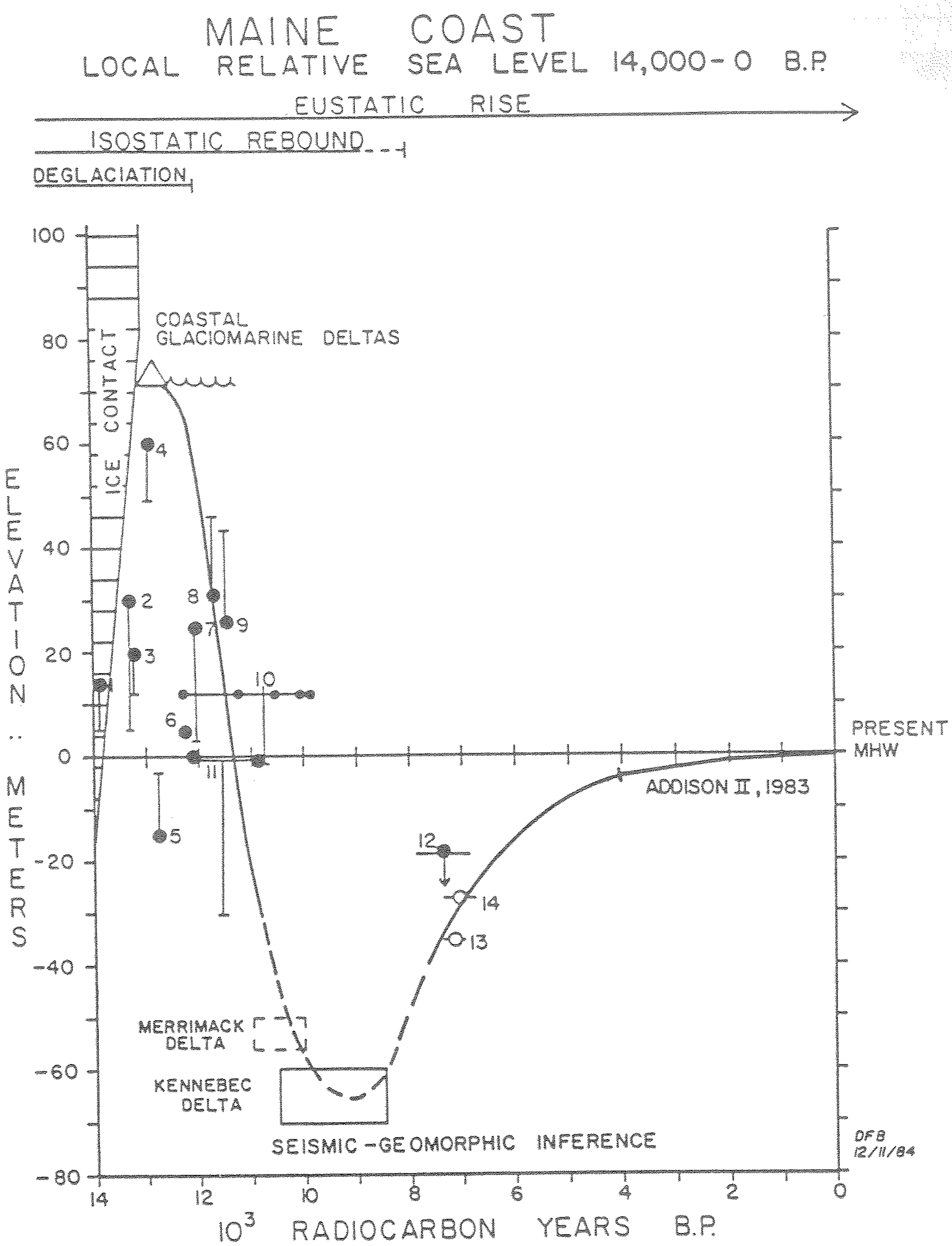


Figure 28. Relative sea level curve for Maine (from Belknap et al., 1986)

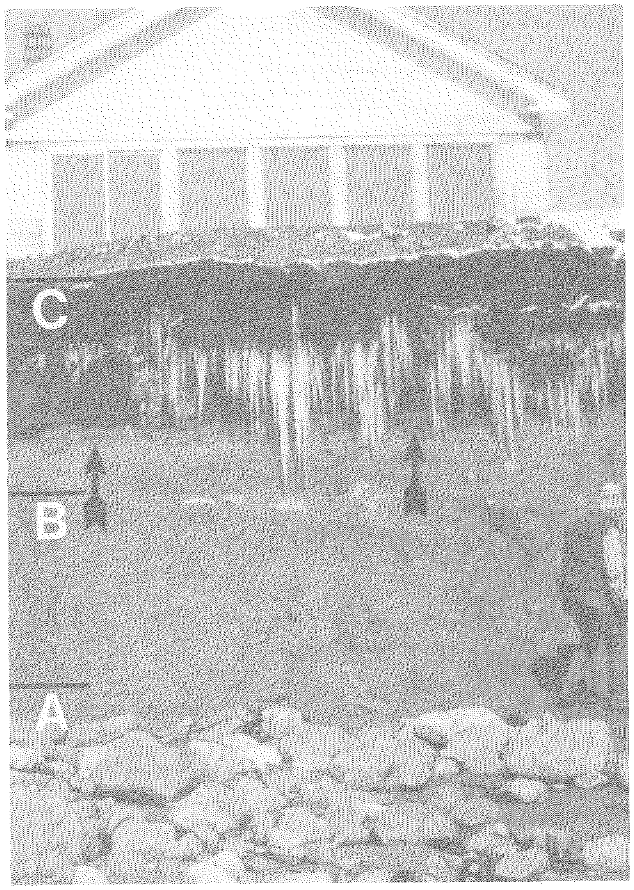


Figure 29. Photograph of eroding outcrop at Jonesport, Maine showing Quaternary stratigraphic column. A - Till, B - Presumpscot Formation, C - Holocene Peat.

and gravel beds. These beds become finer grained in a short seaward direction and numerous faults and water-escape structures are visible in the most ice-distal exposures of the pit. Blocks of till and large boulders commonly appear as ice-rafted material in the stratified sediment. The top of the moraine is partly covered by relatively well sorted sand and gravel apparently reworked from underlying deposits.

This exposure has been interpreted as subaqueous outwash deposits interfingering with till. The fold in the stratified sediment was apparently caused by a minor advance of the ice over its outwash deposits. The uppermost unit probably represents shoreline deposits formed during the late Pleistocene regression of the sea.

Stops 3-3. Pineo Ridge Delta

Leaders: J. T. Kelley, A. R. Kelley

Purpose: To examine the geomorphology of a large, glacio-marine delta

Route: Leave Tracy Corner 1030 hr; proceed 2 km (1.2 mi) north to Route 1 and travel west 4 km (2.5 mi); turn north to Columbia and proceed 4 km (2.5 mi); travel west on side road for 3 km (1.9 mi); arrive at 1045 hr.

Introduction:

In southern Maine emerged ice-marginal deltas are scattered throughout the area of marine submergence. Most are concentrated in close proximity to the former ice margin where they received sediment-laden meltwater. All formed at the local RSL highstand thereby providing the principal data for the construction of isolines of postglacial emergence (Fig. 12; Stone, 1899; Crossen, 1984; Thompson, et al., 1983; Miller, 1986).

Pineo Ridge (Figs. 2, 32) is an emerged ice-contact delta complex (Borns and Hughes, 1977; Miller, 1986) approximately 85 m (279 ft) above present sea level which formed 13,000 - 12,500 yr B.P.

Miller (1986) has recently interpreted the timing of deposition of glacial features near Pineo Ridge in terms of sea level change (Fig. 33). The coastal moraines visited earlier were deposited beneath sea level and are the oldest Quaternary deposits in the area. As the ice margin retreated into the area of Pineo Ridge, its orientation was controlled by bedrock pinning points and as a result the latest moraines are almost orthogonal to earlier moraines. At Pineo Ridge, ice became grounded and deposited a substantial moraine-delta complex (Fig. 33) at the marine limit (approximately 80 m; 262 ft). Sea level may have remained stationary for a time at this elevation because a prominent shoreline was cut into the front of both the delta and the moraine (Fig. 33). The shoreline, seen at East Base, continues for nearly 100 km to the northeast.